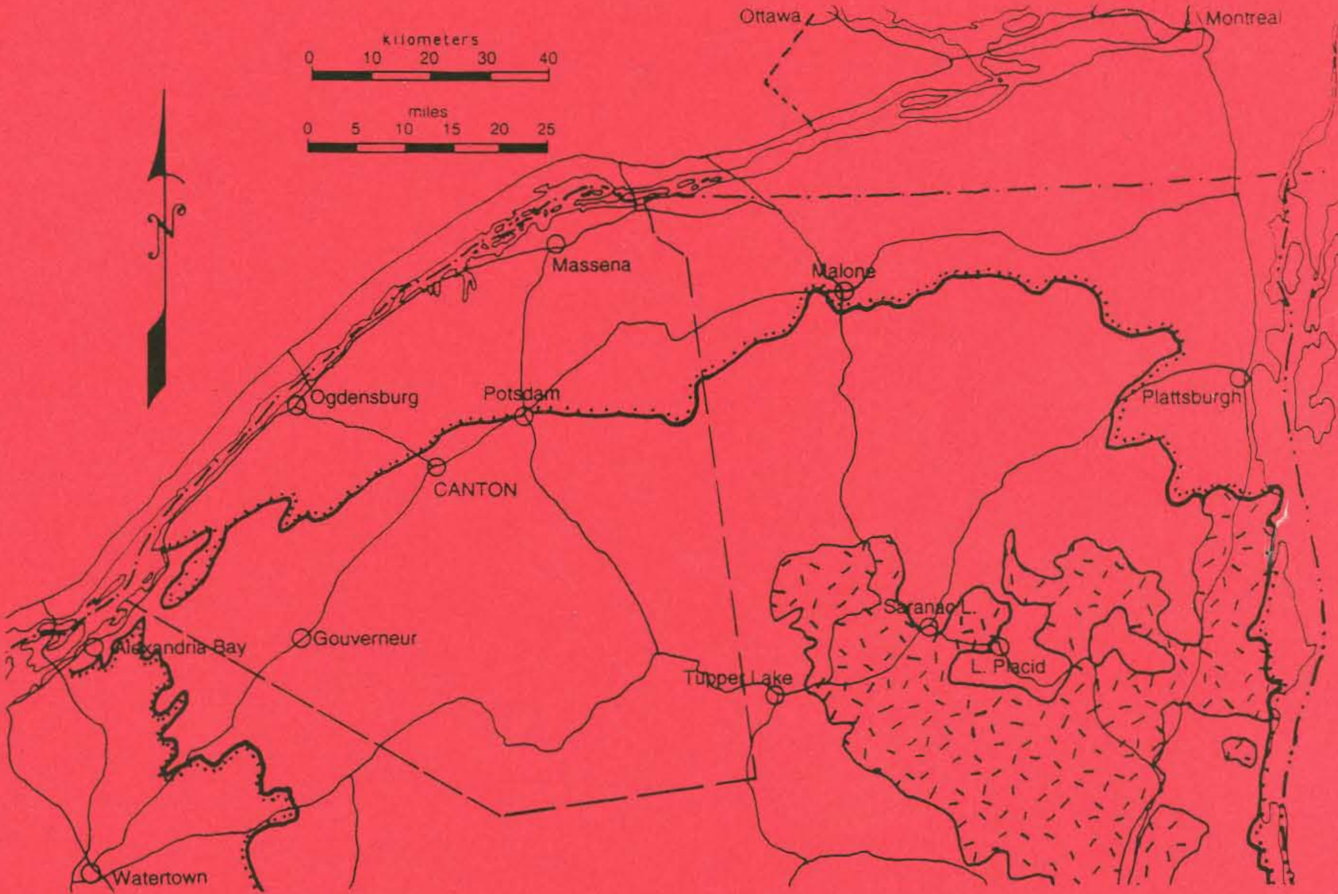
FIELD TRIP GUIDEBOOK

New York State Geological Association



65th Annual Meeting

St.Lawrence University
Canton, New York
September 24-26, 1993

New York State Geological Association

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John T. Bursnall, Editor

Field Trip Guidebook

Department of Geology
St. Lawrence University
Canton
New York, 13617

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James Stewart Street
July 26, 1934 - October 16, 1993

Born and brought up in Chicago, Illinois, Jim attended Wright Junior College and then the University of Illinois in Champaign/Urbana, majoring in Geology. His graduate work was done at Syracuse University, in Syracuse, New York, where he earned both his Master's degree and Ph.D. in Geology. There he studied the glacial geology of the Tug Hill Plateau and portions of the Black River Valley of New York.

Jim came to St. Lawrence in 1966 after working for Texaco in New Orleans, Louisiana, and was a member of the Geology Department faculty for 27 years. During this time he served six and a half years as chairman of the Geology Department, and was elected to two terms as Faculty Delegate to the Board of Trustees, four terms on Faculty Council, three terms on Professional Standards Committee, and held elected positions with the St. Lawrence University Chapter of the American Association of University Professors, and Sigma Xi. He was an elected delegate to the Geology section of the Council for Undergraduate Research. He served as University Marshal from 1988-1993 and was appointed James Henry Chapin Professor of Geology in 1990.

Glacial geology and the interpretation of the meaning of the Earth's surface features was Jim's area of research. It took him into the field with family and friends and served as the focus for his noteworthy geology courses such as The Dynamic Earth, Roadsides and Rivercuts and Geomorphology. Too, it became the basis for continuing professional relationships with many former students. His research was a significant contribution to the recently published Adirondack Sheet of the New York Geological Survey Surficial Geologic Map of the State. His interests in the geology of the Erie Canal System led to his teaching in the History of Geology. He was a member of the Geological Society of America, the National Association of Geology Teachers, the Friends of the Pleistocene, the International Quaternary Association and the History of Geology Society.

Jim worked on numerous faculty and University committees, but most important to him was his teaching, his contact with the students. It was in this area that he made his greatest contributions to St. Lawrence University.

Jim is survived by his wife Sally, two sons, James S. Street, Jr. and David M. Street, SLU '86, and a daughter, Anne F. Street-Gross, SLU '82.

EDITOR'S PREFACE AND ACKNOWLEDGMENTS

The 65th Annual Meeting the NYSGA, held at St. Lawrence, is now behind us and the guidebooks are, at last, about to go out (!). But before they do I wish to thank the numerous people who helped with the planning, organization and running of the conference. In particular I should like to mention the trip leaders, since without their willing support the meeting would not have been possible; also all those who travelled the distance, on an unavoidably inconvenient weekend for some, and braved the characteristic vagaries of North Country weather at this time of year - bright sun (if not downright warm) on the first day and torrential rain (or was it sleet?) on the second. Particular thanks are also due to: Jim Olmsted, Executive Seecretary of the Association, for keeping us to appropriate deadlines early in the organizational process; Kurt Salzburg and the staff at St. Lawrence's Conference Services for their invaluable assistance; St. Lawrence Food Service for their great banquet; Alice Quackenbush for controlling the processing of applications, mailings, phone queries, and everything else; Mike Whitton for hours of 'cutting and pasting'; and last, but far from least, all of the Geology Club members who willingly helped with registration, welcoming party, stuffing packages, and all the other time-consuming and sometimes onerous tasks that are necessary for the smooth running of a conference.

Finally, a separate note of thanks must go to Dave Franzi and Ken Adams who, together with Don Pair, put on a last minute trip when it was realized that Jim Street would not be able lead his intended trip - one that was to have included a summary of some of Jim's research over the past decade or more.

This short editorial piece is followed by the dedication of this volume in Jim's memory.

John Bursnall
Department of Geology
St. Lawrence University
Canton, NY

DEDICATION

We, the staff, students, and alumni of the Geology Department at St. Lawrence University wish to take the occasion of the 65th Annual Meeting of the New York State Geological Association to dedicate this Guidebook, and this conference, to our long time colleague, teacher and friend Dr. James S. Street. Jim's thoroughness, graciousness, and strong scholarly interest in his North Country surroundings are a model for us all, and have been so time and again.

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TRIP A1

PRECAMBRIAN GEOLOGY OF THE NORTHWEST ADIRONDACK LOWLANDS: A STRATIGRAPHIC VIEWPOINT

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INTRODUCTION

The idea of widespread preservation of a coherent stratigraphic sequence in middle Proterozoic rocks in the Adirondack Lowlands has been challenged from several standpoints. Workers in the Adirondack Highlands, and elsewhere in high grade Grenville terrane, have drawn attention to the extent of tectonic realignment that has accompanied folding, ductile deformation and metamorphism. Rock units that were originally discordant have been swept into into parallelism to form "straight gneisses" that resemble a layered sequence. Sequences of metasedimentary rocks have been obscured by the size, volume and frequency of igneous intrusions. Contacts between intrusive and sedimentary rocks have been obliterated by tectonism, the acquisition of gneissic foliation and by metamorphic differentiation. Further complications are introduced where lithologic units have been disturbed or cut out by faulting, and where slivers of much older rocks have been thrusted into the terrane. Thrusting and stacking of crustal segments from southeast to northwest is proposed to have occurred in large parts of Grenville terrane in Canada (Davidson, 1986). For a general review of field and textural criteria used to recognize intense deformation and tectonic realignment, see the summary by Passchier and others (1990).

The Adirondack Lowlands consist chiefly of recognizable metasedimentary rocks and extensive units of fine grained gneisses whose protoliths are a matter of controversy. Metaintrusive rocks here constitute a much smaller proportion of outcrops than in the Adirondack Highlands, and the intrusive bodies are generally smaller in size. Ductile deformation has been extensive in some rock types, particularly the marbles (now annealed?), Hyde School Gneiss and the Huckleberry Mountain granitic gneisses (Tewksbury and others, 1993, and this guidebook). Nonetheless, primary features have been preserved that include stromatolites in carbonates and cross bedding in quartzites. Dikes, apophyses, aplite veins, pegmatites and phenocrysts are preserved in certain metaintrusive bodies. Highly ptygmatic and cross cutting migmatite veins have not been realigned to parallelism in Popple Hill Gneiss, and rocks in some of the fault panels mapped by Brown (1989) preserve a stratigraphic succession comparable with that elsewhere. The case for pervasive tectonic realignment that mimics, but does not represent, a true stratigraphic succession must be demonstrated for local areas in the Lowlands.

Stratigraphic principles are applicable where widespread rock units are shown to be older than the metaigneous rocks that intrude them. Zircons from the widely distributed Popple Hill Gneiss, portions of which are regarded as metavolcanic by Carl (1988), have a U-Pb age of 1214 +/- 21 Ma (Carl and Sinha, 1992) and a whole rock Rb-Sr age of 1297 +/- 41 Ma (Grant and others, 1984). Zircons from another possible metavolcanic unit, the Hyde School Gneiss, range in age from 1219 +/- 52 Ma to 1284 +/- 7 Ma, but ages of 1236 +/- 6 Ma and 1230 +/- 30 Ma for unzoned zircons from two of the bodies have been considered representative (McLelland and others, 1992). Intrusive rocks in the Lowlands generally fall into the range of 1170-1130 Ma as is the case for anorthosite, mangerite, charnockite and granite members of the AMCG intrusive series in the Highlands (McLelland, 1986).

The above remarks are made in the aftermath of a large number (more than 40!) of published U-Pb zircon ages (McLelland and others, 1988; McLelland and Chiarenzelli, 1990; 1991; Chiarenzelli and McLelland, 1991; McLelland and others, 1992) which have had great influence on the understanding of Adirondack geology, particularly in the Highlands. The paper by McLelland and others (1992) on a major Lowland rock type, the Hyde School Gneiss, is of concern because it proposes to overturn a set of stratigraphic conclusions established over decades of field mapping and mining exploration. Hyde School Gneiss, long regarded as a basal stratigraphic unit throughout the Lowlands, is now proposed to be an intrusive igneous rock with no stratigraphic significance. The possibility that metavolcanic rocks could constitute a large proportion of Adirondack rock types is relegated to the backwater of "phase II" in a three-phase summary of the history of Adirondack conceptual schemes (McLelland, 1991). That argument is the subject of a field trip to be held on the weekend following the 1993 NYSGA meeting. Interested readers can obtain copies of the guidebook from the organization known as Friends of the Grenville (FOG, which does not necessarily reflect the state of mind of its members). For the Lowlands and for this guidebook, we plan to "dress up" the corpse of "phase II" which is characterized by paradigms of stratigraphy and stratigraphic correlation. We address some of the issues in the field stops described below.

REGIONAL STRATIGRAPHY

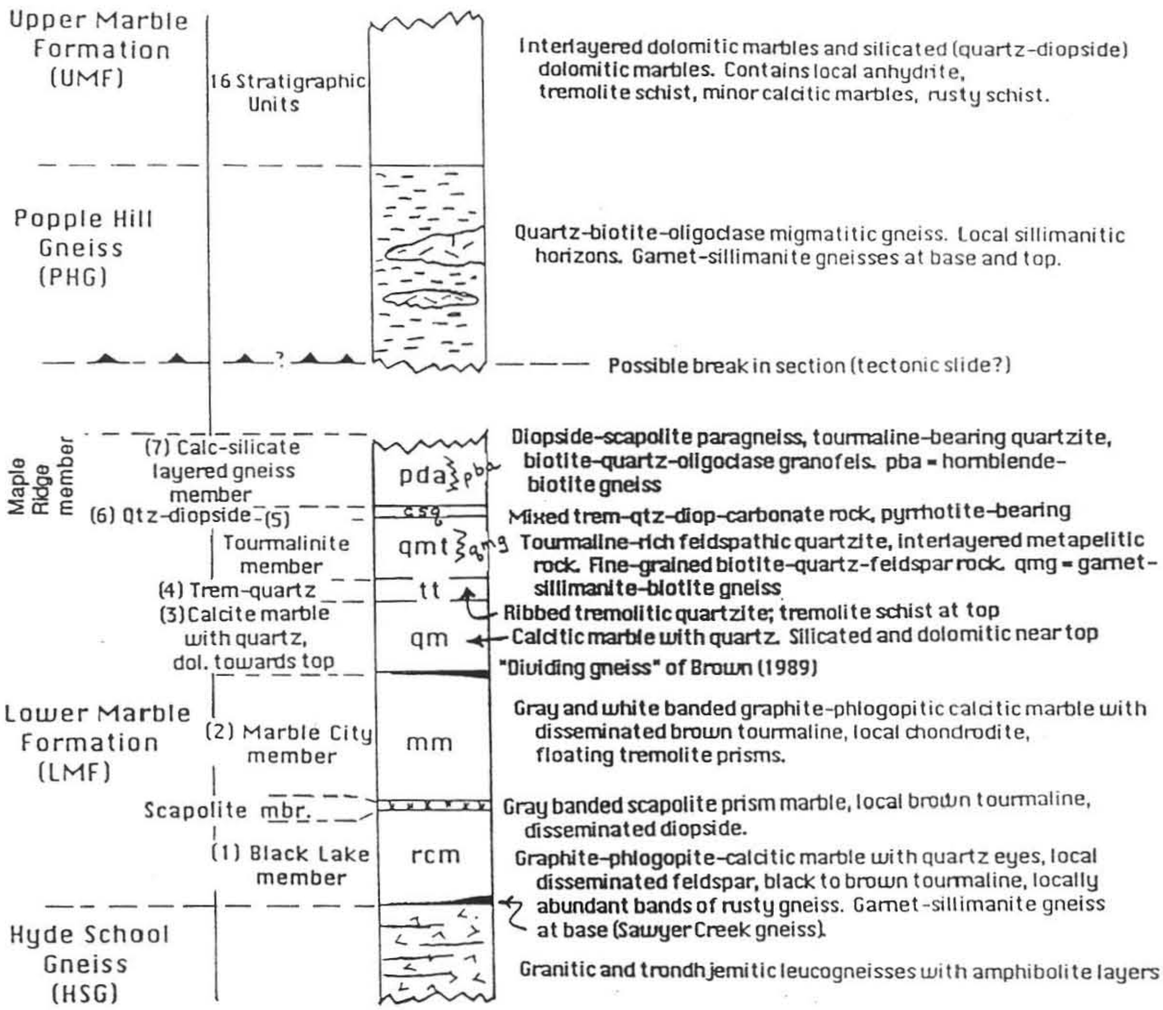
The Adirondack Lowlands represents an extension of the Canadian Grenville Province into New York State, beginning at the eastern end of Lake Ontario. The Lowlands are part of the Frontenac Axis terrane of the Central Metasedimentary belt. They are separated from the Central Granulite terrane of the Adirondack Highlands by the Carthage-Colton Mylonite Zone (also described in the 1993 FOG guidebook) and are underlain by a succession of 1.2-1.3 Ga-old platformal, chiefly carbonate metasedimentary and metavolcanic rocks that reached upper amphibolite facies metamorphism.

Four lithostratigraphic units or formations (Fig. 1) have been proposed to underlie the Lowlands (deLorraine and Carl, 1986; Carl and others, 1990). A geological fold-out map of the Lowlands (Fig. 2) utilizing the four formations will be distributed to those attending the field trip, and copies can be obtained by writing deLorraine. Whether these units constitute a lithotectonic stacking sequence or a true stratigraphic sequence is subject to scrutiny, but the continuity of the units across the Lowlands, their usefulness as a predictive tool, and the compatibility of the sense of tops with structural and geochronological data suggest that a stratigraphic model is indeed viable. Detailed mapping at scales of 1:2400 and 1:4800 has been used in local areas to determine structural relationships and to construct an idealized stratigraphic column for the region. Correlation among stratigraphic columns produced for the Lowlands is shown in Fig. 3, and an idealized NW-SE geologic section across the Lowlands is shown in Fig. 4. Note that rock units high in the section occur only in the southern and central parts of the Lowlands.

The four formations include the Hyde School Gneiss, a complex of basal leucogneisses with thin, concordant amphibolite layers (Fig. 1). The rocks are followed up section by the Lower Marble formation, Popple Hill Gneiss and by Upper Marble formation. Lower Marble formation is subdivided into members that have great lateral continuity and distinctive composition. The formation is dominated by graphitic-phlogopitic-calcitic marbles with disseminated brown to black tourmaline, whereas Upper Marble formation consists of 16 dolomitic and silicated dolomitic members that are lacking in tourmaline. The intervening Popple Hill Gneiss is a thick and variably migmatitic unit of dacitic composition that lies in apparent tectonic contact with the underlying Lower Marble formation.

We have subdivided and described members of the Lower Marble formation (Fig. 1) for several reasons. First, the structural interpretation of a highly deformed terrane with intricate map patterns necessarily begins with a reconstruction of the layering or stratigraphy of the map area. Second, the extent of *apparent* stratigraphic control on the distribution of Hyde School Gneiss needs to be evaluated through map pattern analysis. Third, very detailed geologic maps are necessary to determine the location of regional thrust faults, tectonic slides and other faults. Stratigraphic control provided by geological mapping also places constraints on the location of potential suture zones. The concept of amalgamation of microterranes along cryptic suture zones is an active area of discussion in the Lowlands, particularly in view of the discovery of a 1416 +/- 6 Ma-old leucogneiss dike on Wellesley Island by McLelland and others (1988; 1992). Lithologic successions on either side of postulated suture zones must be understood in some detail before accessments can be made.

We briefly describe the subdivisions of the Lower Marble formation because it is the most widespread formation in the Lowlands and because it directly overlies the Hyde School



Preliminary: errors and/or omissions attributable to deLorraine

Figure 1 Idealized stratigraphic column for the Northwest Adirondack Mountains, New York

Gneiss. Identical lithologies are present to the northeast in Canada and to the southeast in the Adirondack Highlands, and correlation with these areas may be possible.

STRATIGRAPHY OF LOWER MARBLE FORMATION

The basal or Black Lake member of Lower Marble formation occupies most of the terrane northwest of the Beaver Creek lineament (Figs. 2, 3) where the member appears to be much thicker than east of the lineament. The member consists of rusty graphite-phlogopite-calcite marbles with rusty gneiss bands and intercalated quartzite horizons. Coarse grained, homogeneous calcitic marble and massive diopside rock are also present. Accessory minerals include disseminated feldspar and brown to black tourmaline. Disseminated green granular diopside is common in the marble, as are quartz "eyes." Upsection are distinctive marker horizons that include a scapolite prism marble, a magnesian marble referred to as the Marble City member (Fig. 1), and a ribbed quartzite with tremolite partings. The quartzite is overlain by a tourmalinite-quartzite-biotite schist member that may constitute a unique marker horizon (Brown and Ayuso, 1985). At the top of Lower Marble formation is the Maple Ridge member that is composed of calc-silicate gneisses that may be in structural contact with the overlying Popple Hill Gneiss.

Each body of Hyde School Gneiss in the Lowlands is surrounded and overlain by the Black Lake member of Lower Marble formation (Fig. 2). Subdivisions of Lower Marble formation were first elucidated by Brown (1989; 1978; 1969) who worked out the section in the North Gouverneur area. That section was found to have excellent predictive value elsewhere and was expanded by deLorraine to produce an idealized column (Fig. 1).

STRUCTURAL GEOLOGY

Folding

At least four phases of deformation have been proposed for the Adirondack Lowlands (Brown, 1971; Brocoum, 1971; Foose, 1974; Foose and Carl, 1977; deLorraine, 1979; Wiener, 1981; Guzowski, 1979; Weiner and others, 1984). The earliest recognized phase of deformation occurred at a high metamorphic grade to produce isoclinal folds with axial planar foliation that defines the prominent regional foliation. No unequivocal major early folds are known, and minor folds are rare. Some regionally extensive mylonites are deformed by second phase folds and are interpreted as refolded, early-phase tectonic slide surfaces.

Second phase folds are isoclinal in style and moderately to strongly overturned to the southeast. They deform early axial planar foliations and have curvilinear hinges that "porpoise" within their axial surfaces. Axial surface traces trend NE-SW to define the prominent regional grain. These have been described as sheath folds (deLorraine and Carl, 1986; deLorraine in Bohlen and others, 1989; Tewksbury and others, 1993). Good examples of second phase isoclinal sheath folds include the Sylvia Lake syncline (Fig. 2) that hosts the

Antwerp – Rossie granitoids

Amphibolite – metagabbro

Hermon and other granites

Syenite

Upper Marble

Popple Hill Gneiss

5, 6, 7. Udif. px. calc-silicate; biot; tour. gneisses

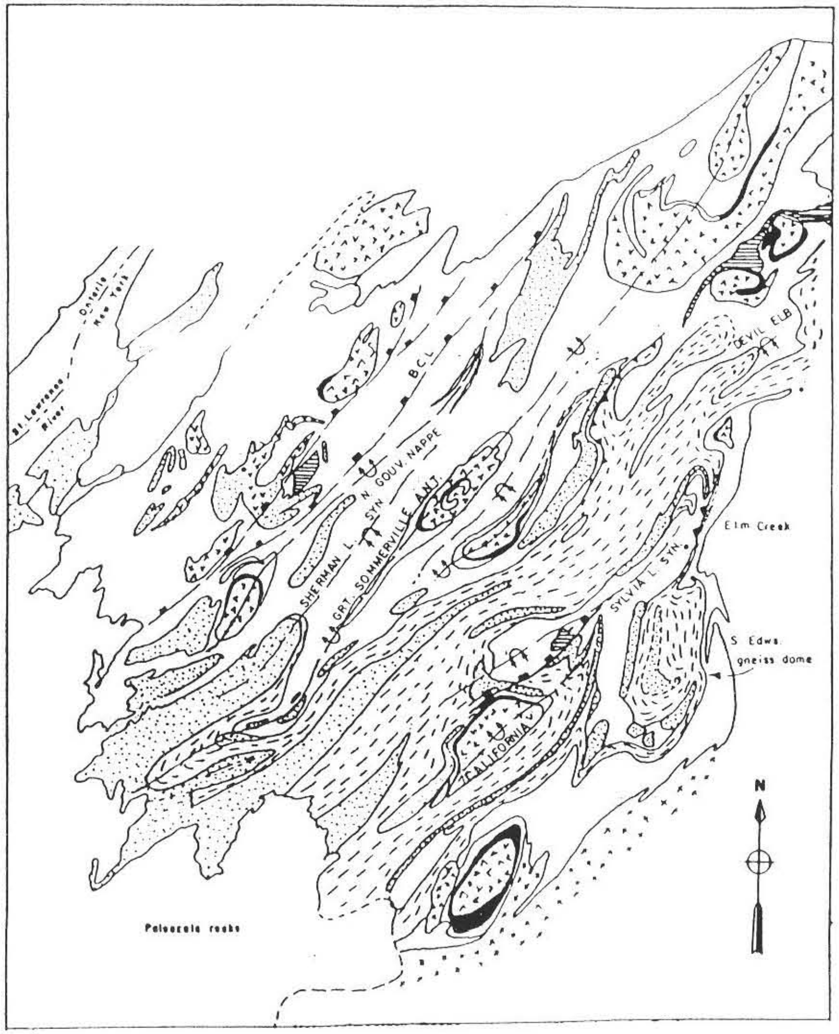
Tremolitic quartzite

nm Bndd magnesian mbl, "Marble City" mbr.

rcm Rusty cal mbl, "Black Lake" mbr

Sill – garnet gneiss "Sawyer Creek gneiss"

Hyde School Gneiss

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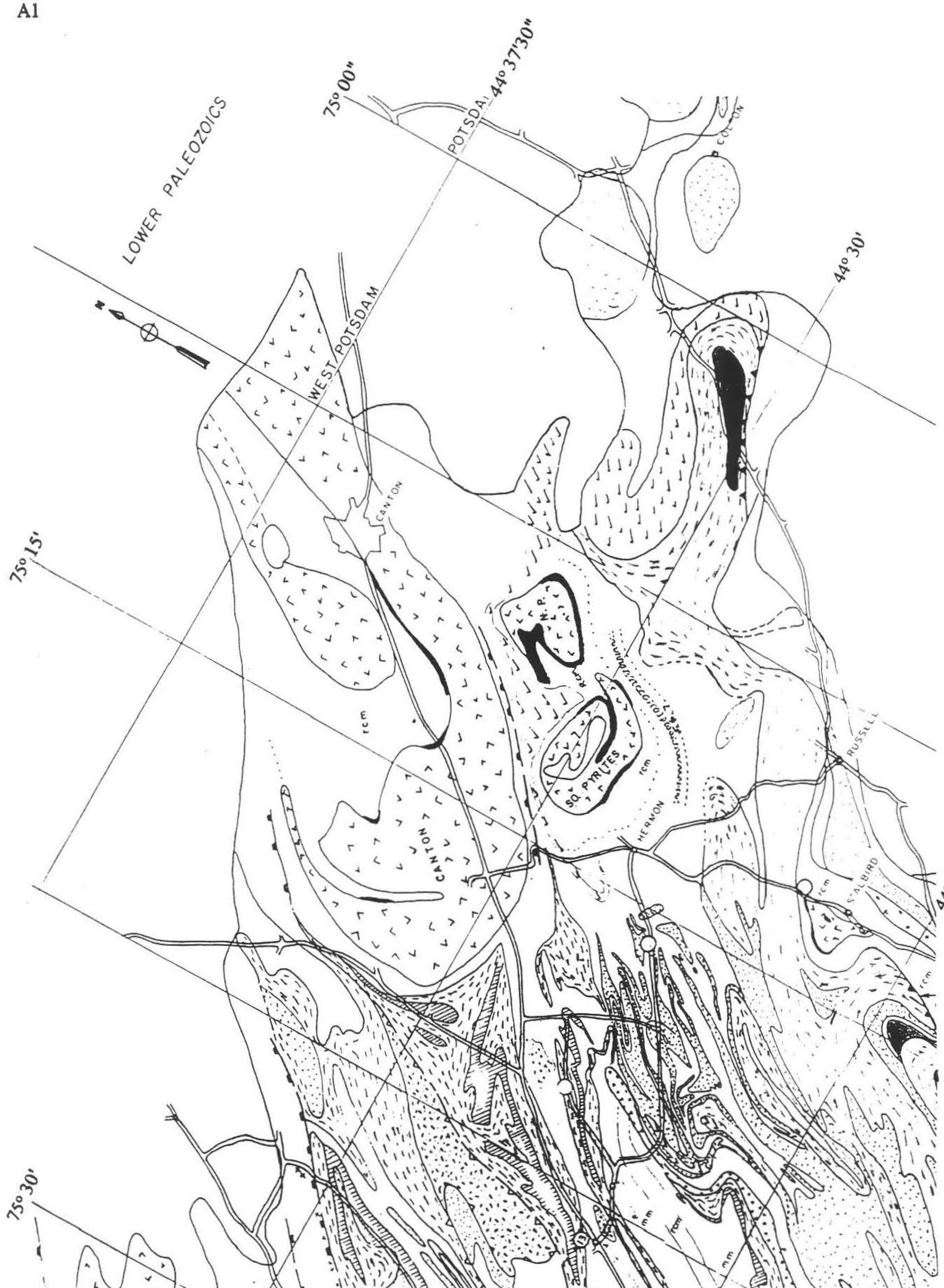
FISH CREEK
HICKORY LAKE
EDWARDSVILLE
POPE MILL
HYDE SCHOOL
BRASIE CORNER
Beaver Creek Lineament
GOUVERNEUR
RESERVOIR HILL
Black Creek Fault
Grass Creek Fault
ROSSIE
CHIPPEWA BAY
ST. LAWRENCE COUNTY
JEFFERSON COUNTY
DODD'S CREEK
PAYNE LAKE
Pleasant Lake Fault
SOMERVILLE
CANADA
UNITED STATES
ROCKPORT
ALEXANDRIA BAY
REDWOOD
rcm
mm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WEST POTSDAM
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HERMON
CANTON
44° 22'30"
75° 30'
mm

GOUVERNEUR
RESERVOIR HILL
KELLOGG CORNERS
CALIFORNIA
SOMERVILLE
ANTWERP
PHILADELPHIA
THERESA
LOWER PALEOZOICS
NORTH WILNA
CARTHAGE
HARRISVILLE
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LEWIS COUNTY
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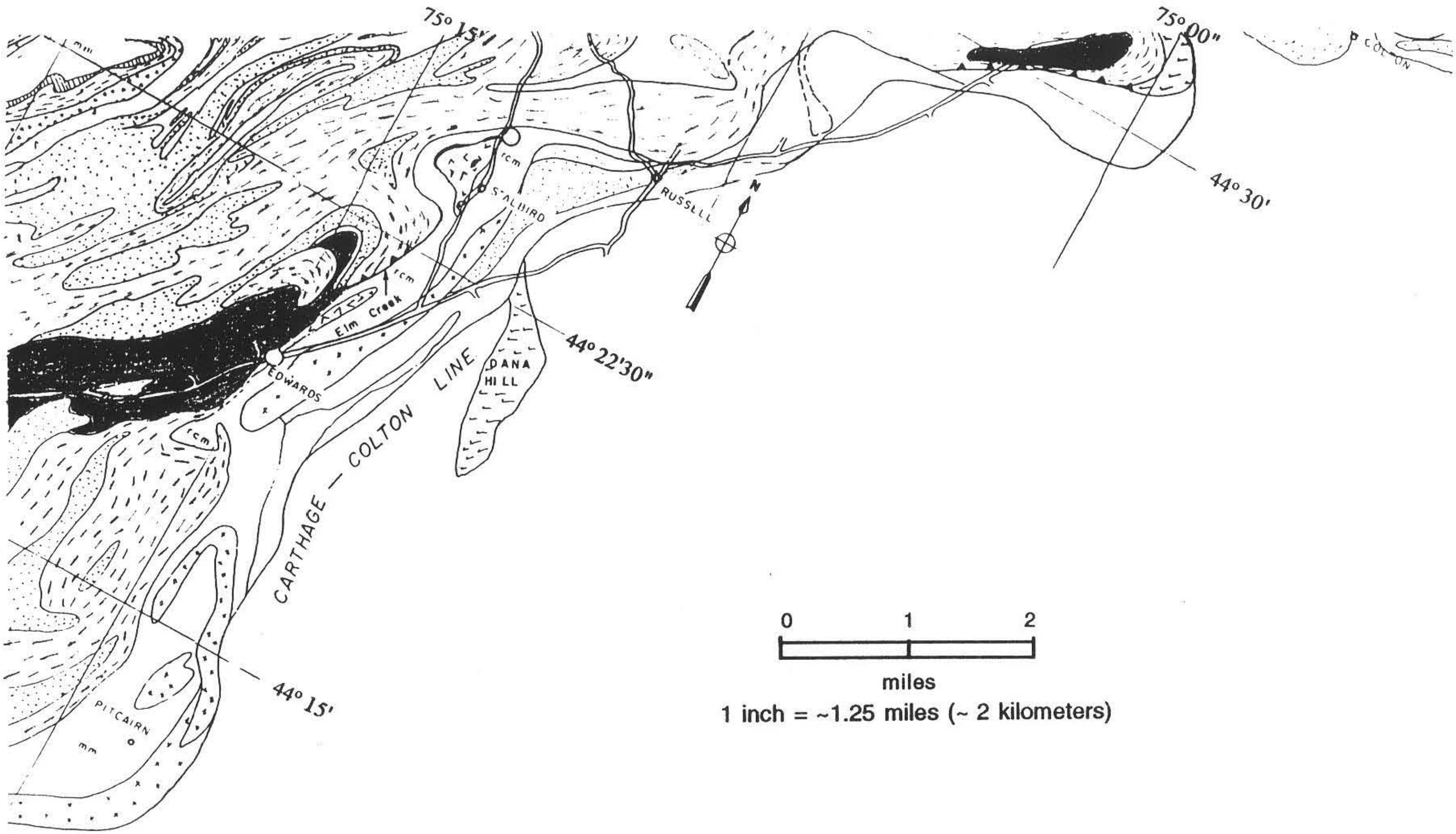
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RUSSELL
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Elm Creek
EDWARDS
44° 22'30"
DANA HILL
CARTHAGE — COLTON LINE
44° 15'
PITCAIRN
mm
0
1
2
miles
1 inch = ~1.25 miles (~ 2 kilometers)

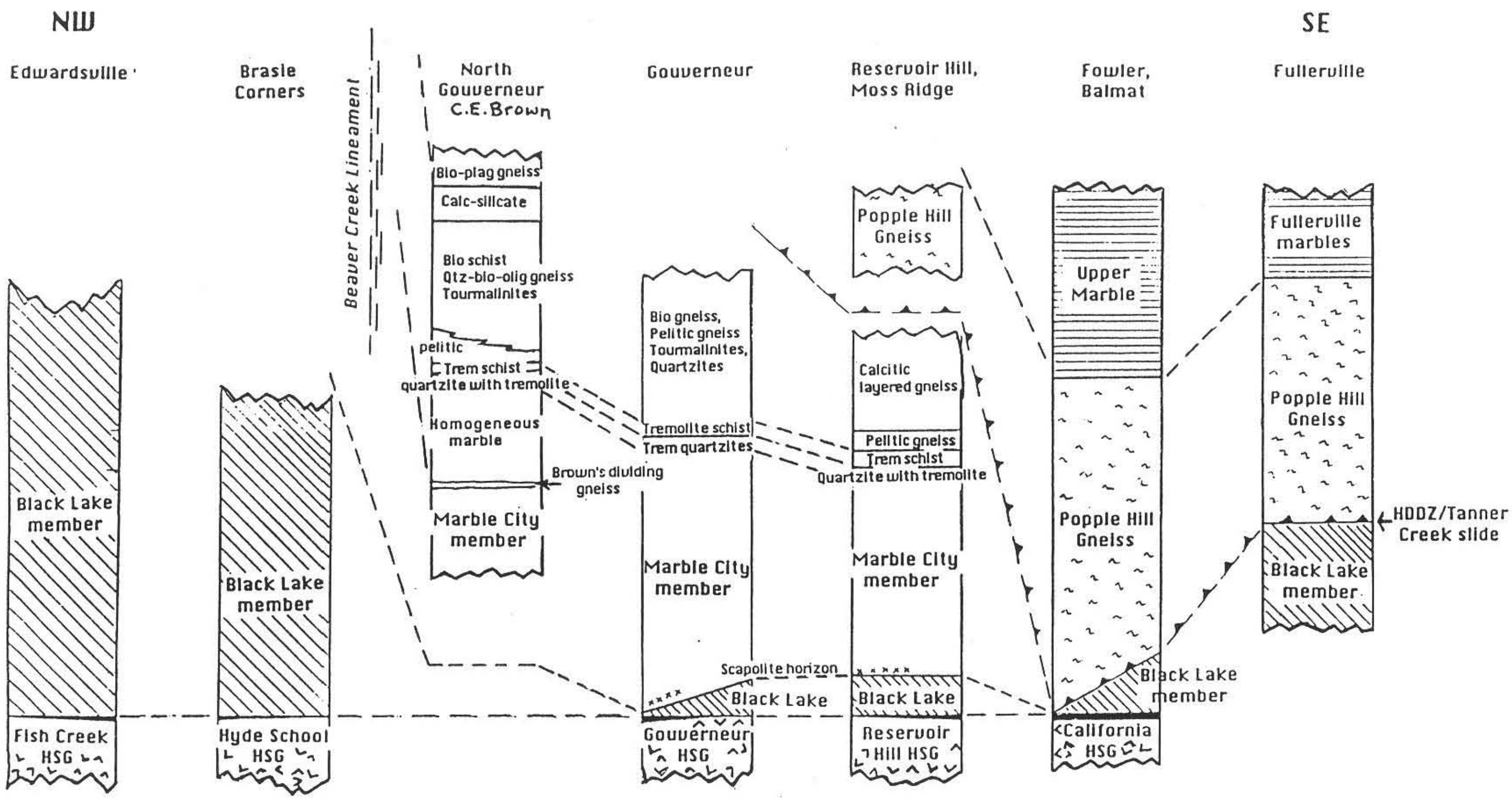

Figure 3 Proposed stratigraphic correlations from NW to SE across the Northwest Adirondack Mountains

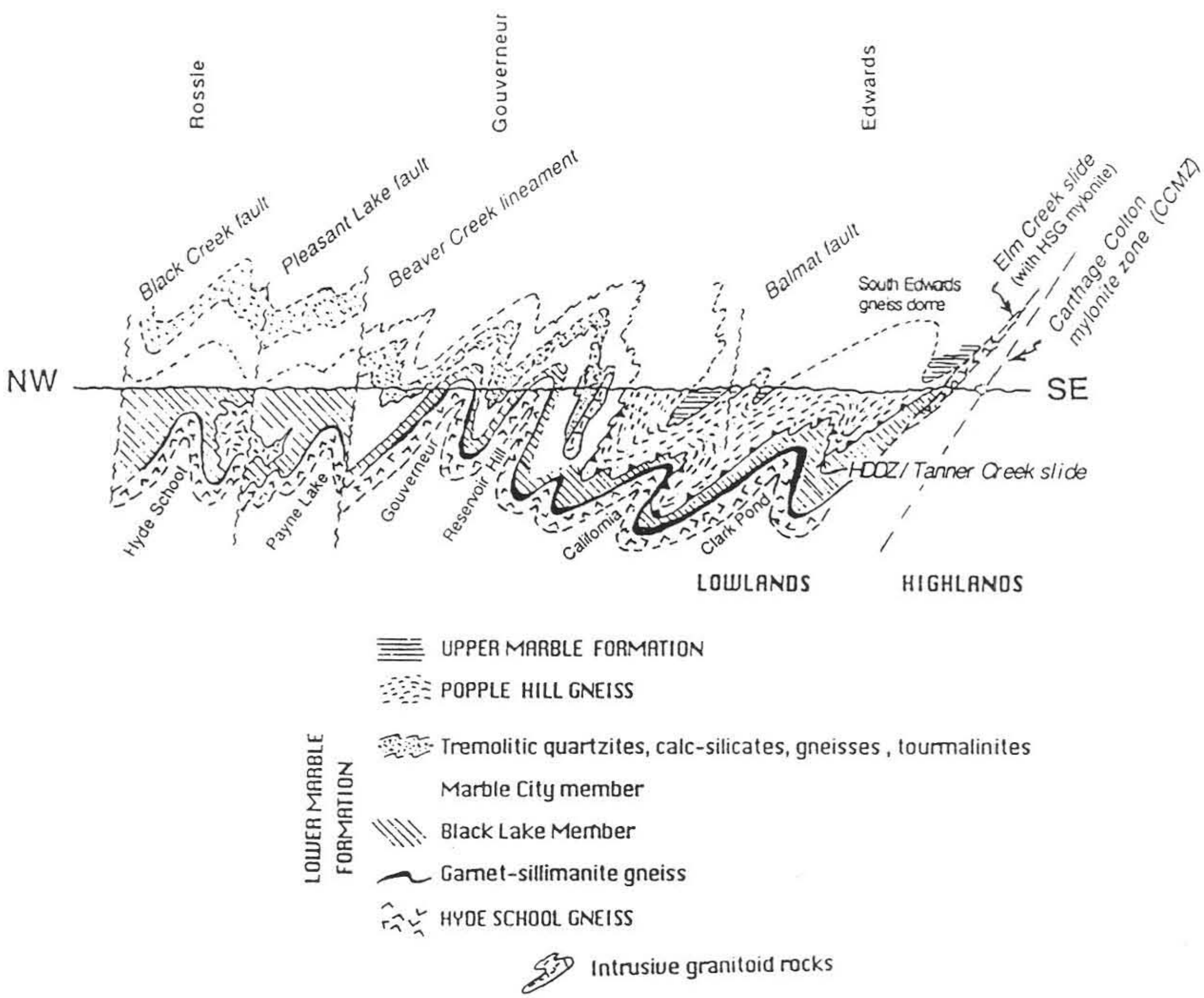

Figure 4 Proposed cross section from NW to SE across the Northwest Adirondack Mountains

zinc orebodies of the Balmat-Edwards district. The geometry of this structure is fairly well understood from diamond drilling and zinc mining. Also included are the California anticline, the Great Somerville anticline, the Sherman Lake syncline, and the North Gouverneur nappe that is cored by the Payne Lake body of Hyde School Gneiss.

Third phase folds are more or less co-planar with the local second phase folds in the Balmat mines, and the trends are also N-NE. Third phase folds tend to be open to tight in style and are recognized where they refold axial surfaces of second phase isoclines. Fourth phase folds are gentle warps to open folds with NW trends that have been recognized locally (Brocoum, 1971; Foose, 1974; deLorraine, 1979; Wiener, 1981; Brown, 1989).

Interference of fourth phase folds with earlier NE-trending folds was previously thought to have produced the dome and basin map pattern throughout the Lowlands (Foose, 1974; Foose and Carl, 1977; deLorraine, 1979; Wiener, 1981; Wiener and others, 1984). Detailed analysis of the Sylvia Lake syncline and the Pierrepont sigmoid, in conjunction with zinc mining at Balmat-Edwards, led to the suggestion that these doubly plunging synclines were sheath folds, and that "domes" cored by Hyde School Gneiss were actually the apical projections of complimentary anticlinal sheath folds (deLorraine and Carl, 1986; deLorraine, in Bohlen and others, 1989). Detailed mapping and fabric analysis by Tewksbury provided evidence for major sub-horizontal shear regimes that later D2 deformation amplified into sheath folds.

Fold interference, thus, is seen by recent workers to have played a relatively minor role in the formation of the regional dome and basin pattern. Although cross folding is capable of explaining the pattern, it cannot easily explain the ovoid outcrop patterns made by overturned bodies of Hyde School Gneiss. Heart and anchor patterns would be expected for these bodies unless the movement line of the NW-trending folds was contained entirely within the axial surfaces of the earlier, NE-trending folds. Hyde School Gneiss may be more susceptible to ductile shear, but the same folding occurs in metasedimentary envelopes around the bodies of Hyde School Gneiss, suggesting that they too were sheath folded.

The top and base of Popple Hill Gneiss are mylonitic, suggesting that ductile shear was heterogeneous and limited to discrete surfaces at contacts with the carbonates. But Popple Hill Gneiss must also be involved in the sheath folding because its contact with the Upper Marble formation defines the Sylvia Lake syncline. The question arises as to whether stratigraphic continuity can be maintained under conditions of high strain associated with sheath folding. The answer, in part, is that folds such as the Sylvia Lake syncline, Great Somerville and California anticlines, are regional folds whose overall geometry *approaches* that of sheath folds in that the hinge lines are recurved and doubly plunging. Extension and elongation has produced local stratigraphic discontinuities, but large-scale continuity across the NW Adirondacks can be demonstrated.

Faulting

Brittle Faults:

Two broad classes of faults, "brittle" and ductile, occur in the Lowlands, some in conjunction with high grade metamorphism and folding in ductile regimes, such as thrust faults, tectonic slides and shears. Others of a brittle nature occur where the offset may have been post-metamorphic.

Northeast striking, steeply dipping "brittle" faults have broken the Lowlands into fault panels (Figs. 2, 4), each of which ostensibly differs in structural style and stratigraphic content. "Brittle" is used advisedly in that early movements may have occurred in a ductile regime producing mylonite, whereas the latest movements were brittle and marked by brecciation. The association of these and other faults with mafic plutons is suggestive of a long, protracted history of faulting and intrusion. Important faults include the Black Creek fault, Pleasant Lake fault, Beaver Creek lineament, Oswegatchie fault and the Balmat fault (Figs. 2, 4). Brown (1989) showed that early fault movement was dominantly strike-slip and accompanied by granite and pegmatite intrusions which were later mylonitized. Evidence of late recurrent movement includes offsets of Cambrian Potsdam sandstone and the presence of breccia (Brown, 1989).

Careful mapping of units of the Lower Marble formation suggests that differences in the stratigraphy on either side of a major fault may be more apparent than real. For example, a stratigraphic breakdown on opposite sides of the Beaver Creek lineament reveals that the Black Lake member of the Lower Marble formation is not unique to that area but is exposed in various places SE of the lineament. The member dominates the terrane to the west, whereas members higher up the section are dominant to the east (Figs. 2, 4). Prior to Brown's work, marbles equivalent to the Black Lake section were not recognized east of the Beaver Creek lineament. Recent mapping by deLorraine shows that the member continues across the lineament to the east and that thin shells surround each dome of Hyde School Gneiss (Fig. 2). The member thickens abruptly west of Beaver Creek lineament where it contains more intercalated clastic lithologies such as quartzites and rusty gneisses.

We emphasize that within and across the fault-bounded panels, each body of Hyde School Gneiss is enveloped by the Black Lake member (Fig. 2). East of Beaver Creek, the upper members of Lower Marble are mapped outwardly from shells of the Black Lake member around the Gouverneur and Reservoir Hill bodies of Hyde School Gneiss. Exceptions include the Clark Pond and Stalbird bodies which lay adjacent to the Carthage-Colton line. Here local faulting has excised large parts of the Lower Marble formation. For example, rocks underlying the Pitcairn-Edwards-Stalbird areas (Fig. 2) are dominated by the Black Lake member, whereas the Clark Pond-Harrisville area is dominated by the Marble City member, and the North Russell-Pierrepont-Colton area by uppermost members of Lower Marble Formation. At Edwards, the Black Lake member is in structural contact with the Upper Marble formation; here the entire section of Popple Hill Gneiss has been excised along the Elm Creek slide (as mapped by deLorraine). Thin, mylonitized slivers of Hyde

School and Sawyer Creek gneisses occur along the contact between Upper and Lower Marble formations where they mark the trace of the slide. The lack of recognition of a tectonic contact between the two marble formations is responsible for the proposal that the Lowlands contained but a *single* marble formation (Foose, 1974; Wiener and others, 1984). The geological relationships near Edwards and Russell are consistent with a "NW side down" movement along the Carthage-Colton mylonite zone, possibly resulting from post-Grenville tectonic unroofing and extension (Heyn, 1990). Interestingly enough, Gilluly (1934) first proposed a major tectonic dislocation near Edwards.

Synmetamorphic Faults:

Synmetamorphic tectonic slides and shears are often difficult to recognize and map due to limited exposures and to the possibility of layer-parallel transport. Tectonic slides and shears are well known from the Balmat zinc mines within the Upper Marble formation. Two prominent tectonic slides within the Sylvia Lake syncline at Balmat (Fig. 5) include the Sylvia Lake slide which separates the Fowler and Sylvia Lake orebodies. These orebodies are interpreted to have been contiguous before separation by transport in a counter clockwise sense (Fig. 5). The apparent offset is on the order of 4000 ft. The Balmat slide on the lower limb of the Sylvia Lake syncline places unit 6 of Upper Marble in direct contact with Popple Hill Gneiss. Units 1 through 5 have been excised.

Only a few slides of regional extent have been reported in the literature. Foose (1974) placed the Moss Ridge slide and the Tanner Creek slide (Figs. 2, 3) at the base of Popple Hill Gneiss. Ambers and Hudson (1985) and Hudson and others (1986) described the Hailesboro ductile deformation zone, also at the base of Popple Hill Gneiss. The Tanner Creek and Hailesboro deformation zones represent the same mylonitic horizon that is isoclinally folded by a second phase isocline at Devils Elbow (near Hermon village). To the southwest, the mylonites are folded by the Great Somerville anticline and Sherman Lake syncline which also are major second phase isoclines. Thus, the Tanner Creek/Hailesboro ductile deformation zone is interpreted by us as a folded tectonic slide that may occur at the base of an early phase thrust sheet (Figs. 2, 4). Popple Hill Gneiss was emplaced over the Lower Marble formation along this slide. Its eastern contact is the edge of Popple Hill Gneiss immediately to the west of the Harrisville-Pitcairn-Stalbird-Russell marble belt. At Edwards, it is unclear whether the Elm Creek fault *is* the Hailesboro ductile deformation zone-Tanner Creek slide or a re-activation of the slide along the Carthage-Colton line.

Thrust emplacement/tectonic interleaving can account for the anomalously abbreviated sections of Lower Marble that surround the Stalbird and California domes of Hyde School Gneiss. The upper section of Lower Marble may have been cut out along the base of this postulated thrust sheet. Diamond drilling and detailed mapping (1:4800 scale) at the NE end of the California body of Hyde School Gneiss by Zinc Corporation of America show that graphitic-phlogophitic-calcitic marbles of the Black Lake member overlie the Hyde School Gneiss, separating it from Popple Hill Gneiss. Rusty gneisses also are exposed on the surface and are present in the drill core. Absent are the Marble City and higher members of the Lower Marble formation.

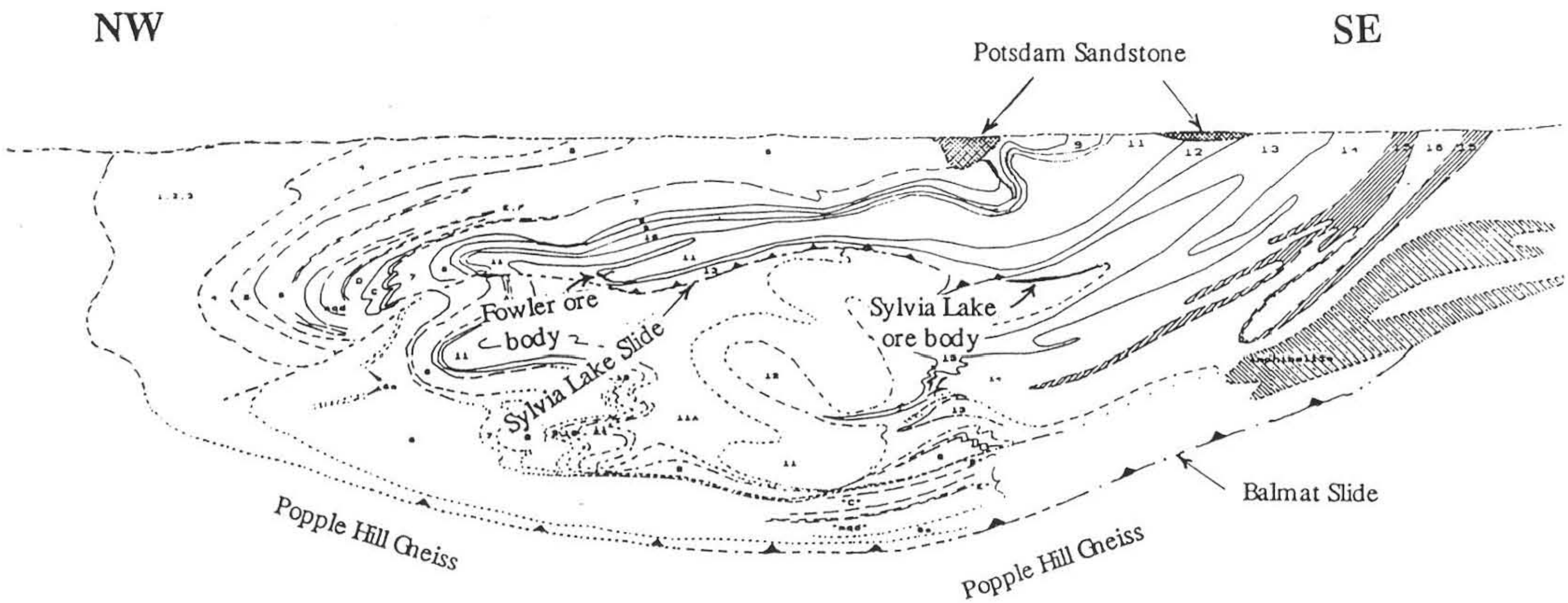

Figure 5 Generalized NW-SE cross section through the Sylvia Lake syncline showing tectonic slides within units of the Upper Marble formation. The units are described in Fig. 7.

ROAD LOG AND STOP DESCRIPTIONS

Cumulative mileage	Miles from last point	Route description
Start		Junction of Main Street (Route 11) and Park Street in the center of Canton at Canton village Park. Follow Route 11 signs out of Canton towards Gouverneur.
23.7 (38.2 km)	23.7	Gouverneur village limits. Kinney's warehouse on left.
24.7 (39.8 km)	1.0	Turn left (S) on Grove St. at the second stop light in Gouverneur, at the Community Bank and the village square.
25.9 (41.7 km)	1.2	Stop 1. Opposite the Cornell farm on Grove St. (ask permission).

STOP 1 Black Lake member of Lower Marble formation, SW end of the Reservoir Hill body of Hyde School Gneiss.

The purpose of this stop is to examine the Black Lake member of Lower Marble formation that overlies the Reservoir Hill body of Hyde School Gneiss. Marbles here are identical in lithology with those exposed west of the Beaver Creek lineament. The marble terrane west of the lineament was labelled the "Black lake Metasedimentary Belt" (Engel and Engel, 1953), and the lowermost member of Lower Marble derives its name from that area (deLorraine and Carl, 1986). Black Lake marbles are quite distinct from the marbles upsection. They are characterized by the presence of ribbed, rusty-weathered, calc-silicate gneiss bands, disseminated black tourmaline, feldspar grains, phlogopite and granular diopside (Fig. 1). Also common are coarse graphite, scattered quartz "eyes" and intercalated quartzites. Massive rusty diopside rocks and a thick, homogeneous calcitic marble unit occur west of the Beaver Creek lineament. Absent at stop 1 is the sillimanite-garnet gneiss that typically surrounds bodies of Hyde School Gneiss. It is informally labelled the Sawyer Creek gneiss after thick exposures near Sawyer Creek along the margin of the California body of Hyde School Gneiss (deLorraine and Carl, 1986).

Note the minor folds in the marbles. Are one or two phases of folds present? One might interpret an early phase defined by rootless, isoclinal intrafolial folds that are refolded by more or less upright to overturned disharmonic folds.

At first glance, one might regard Hyde School Gneiss at this locality as containing few amphibolite layers. When viewed from the rear, i. e. from the high bedrock knobs to the north, the massive outcrops of leucogneiss can be seen as steeply dipping slabs or layers, thanks to the presence of deeply weathered amphibolite layers. Foliation in the leucogneiss is folded about the Reservoir Hill anticline, suggesting that an earlier deformational fabric has been refolded. The anticline is interpreted as a doubly-plunging, southeast overturned, second phase sheath fold that is coeval with the Great Somerville anticline of Buddington and the Sylvia Lake syncline in the Balmat-Edwards mining district.

An important point to make here is the consistent and predictable appearance of Hyde School Gneiss below the Black Lake member of Lower Marble formation. Upsection of this member is a scapolite prism marble that separates the member from an overlying gray and white banded marble called the Marble City member (Fig. 1). The Marble City member will be examined in stop 3. It lacks the rusty weathering gneissic bands and disseminated feldspar observed in the Black Lake member.

We emphasize the contrasting carbonate and calc-silicate rock types because they might otherwise appear to constitute a homogeneous sea of undistinguished marbles. The marbles should not be dismissed simply as "tectonic grease." Stromatolites have been locally preserved, and the persistancy of contacts between relatively thin members for great distances has provided information about the geometries of major folds. Mapping of the thin units is the basis for structural work and zinc exploration in the Sylvia Lake syncline. Unit 6 in the Upper Marble formation, for example, has several 10 to 30 ft thick subunits of distinctive texture and composition that can be traced around the Sylvia Lake syncline by drill core from Balmat to the Hyatt mine, a distance of at least 8 miles (12.9 km). Contacts between rocks of undoubted metasedimentary origin are conformable with each other and, in many cases, with the overlying and underlying gneissic formations. These contacts are interpreted as reflecting original compositional and lithologic differences; that is, they constitute a stratigraphic sequence.

Thus, the appearance of Hyde School Gneiss at a predictable horizon is regarded as the consequence of depositional patterns modified to various degrees by metamorphism, folding, ductile deformation, ductile faulting and partial melting. Stop 12 will add basaltic intrusion to the list, but these late dikes are not to be confused with an earlier amphibolite layering.

Continue south on Grove St.

26.2 (42.2 km)	0.5	Turn right (S) on River Drive and onto the Oswegatchie River bridge.
26.7 (43.0 km)	0.5	Turn right (N) onto Main St. in Hailesboro.
27.2 (43.8 km)	0.5	Turn right (N) onto Routes 58/812.
27.9 (44.9 km)	0.5	Stop 2. Park along Route 58/812, 0.2 km south of Gouverneur village limits.

STOP 2 Scapolite marble member of the Lower Marble (with basalt dike).

Rock here overlies the carbonate and rusty gneiss of stop 1. This characteristically pitted, light gray member or horizon of the Lower Marble formation consists of scapolite-bearing, graphite-phlogopite marbles with quartz segregations. It can be traced southwest of the Gouverneur body of Hyde School Gneiss to aid in defining the "Great Somerville anticline" of Buddington that parallels Route 11 through Somerville village. Scapolite is gray because of finely disseminated graphite; it varies from coarsely bladed to small elongate crystals.

The 1 m thick basaltic dike within the scapolitic marble is N 60 E, vertical and undisturbed in contrast to the highly segmented and metamorphosed dikes to be seen at stop 3. These and other NE-trending Lowland dikes may have been produced by Late Precambrian rifting and the opening of the Iapetus Ocean basin. The dikes are generally vertical, fine-grained, undeformed, amygdaloidal, and show chilled borders against the wall rock. They must have intruded brittle rocks at shallow depths. Many have undergone saussuritization and contain epidote veinlets. They intrude all Precambrian rocks in the Lowlands, but do not intrude patchy remnants of overlying sedimentary rock such as the Cambrian Potsdam sandstone. They show deep weathering (some are altered to hematite) where exposed near the pre-Upper Cambrian unconformity.

Narrow dikes like this one are likely to be strongly saussuritized and to yield unreliable directions of paleomagnetization. Seven samples from thicker, well-crystallized dikes, however, show a grouping of north-seeking poles in the southeast quadrant of the lower hemisphere of an equal area net. Books and Brown (1983) were unable to match the grouping with data from dikes in Ontario.

A hydrothermal event (phase 4 of Brown, 1983) may have reset the ages of many basalt dikes. K-Ar ages from one dike of 405 +/- 11 and 440 +/- 10 m.y. obtained on feldspar and pyroxene respectively (Brown, 1975) are not compatible with the pre-Upper Cambrian age required by field relationships. Ages for Ontario dikes ranging from 600 to 800 m.y. (Park and Irving, 1972) are likely to be representative of Lowland dikes, and the ages of 405 and 440 m.y. may approximate the time of hydrothermal activity (Brown, 1983). Hydrothermal activity is also proposed as the cause for alteration along joints that has produced epidote and reddish feldspar in some Lowland gneisses.

Badger (1993) has studied the largest of the Lowland dikes, a 14 km long, NE-trending dike that cuts Precambrian rocks including the Hyde School body of Hyde School Gneiss (see map of Brown, 1988). The dike shows alkalic affinities and has uniform chemical composition. Tectonic discrimination plots are indicative of within-plate magmatism and OIB-type (Ocean Island Basalt) mantle plume source. These dikes may be related in origin to late Precambrian, rift-related magmas of the Ottawa Graben, the eastern Adirondack Mountains and western Vermont.

Continue north to turnaround.

28.2 (45.4 km)	0.3	Turn around at intersection with old Route 58 (on right). Proceed south on Route 58/812.

29.2 (47.0 km) 1.0 Stop 3. Park along Route 58/812 near the intersection with Main St.

STOP 3 Marble City member of Lower Marble and the broken basaltic dike of the "train wreck outcrop."

The roadcut is located just south of the SW end of a major regional fold hinge, the Reservoir Hill anticline. Numerous minor folds in the marble may reflect the proximity of that anticline. Reconstruction of a refolded isocline suggests an "S" asymmetrical sense when viewed to the north; the "S" is consistent with the location of the fold on the eastern limb of the Reservoir Hill anticline. Axial surfaces of many minor folds have shallow to moderate dips to the west, perhaps giving an indication of the degree of overturning of the anticline.

This massive, coarsely crystalline, blue-gray banded calcitic to slightly magnesian marble is graphic and phlogopitic. We correlate it with Brown's (1989) unit "mm" in the northern part of the Richville quadrangle and North Gouverneur area. Marbles at stops 1 and 2 are lower in the section.

This roadcut was proposed as the type section for a single northwest Adirondack carbonate-bearing formation named the Gouverneur Marble (Wiener and others, 1984). We regard it as a member of the Lower Marble Formation. The rock was quarried throughout the area for building stone in the early 20th century. Note the small town glory that once was Gouverneur (informally called "Marble City") as preserved in the upper levels of the downtown buildings.

The dike here is metamorphosed and disrupted in contrast to the dike at stop 2, but displacement is not great in view of the extensive folding in the marble. The antiform observed in the outcrop probably existed in its present form when the dike was intruded (Schoenberg, 1974; VanDiver, 1976). The lower part of the dike appears to have been a sill, whereas the upper part cuts across the banding in the marble. Possibly the subsequent refolding, as indicated by the shallow broad syncline to the right of the dike, may have caused brecciation and displacement of the dike blocks.

The dike blocks are recrystallized and re-equilibrated near the contact with the marble. They contain little or no plagioclase, abundant diopside and meionitic scapolite with lesser microcline, sphene, tremolite, biotite, quartz, opaque, tourmaline and apatite.

Proceed south on Route 58/812.

29.5 (47.5 km) 0.3 Stop 4. Park along Route 58/812 just north of the bridge over Mattoon Creek.

STOP 4 Steer's Head outcrop where Antwerp-type granitoid gneisses have intruded the Lower Marble.

Gneisses of granite to granodioritic composition comprise a string of large, disconnected, boudin-like masses that extend from Antwerp village northeasterly for 40 km to Moss Ridge-Battle Hill near Gouverneur village. These rocks intrude chiefly the Marble City member of Lower Marble, but near Antwerp village they intrude Popple Hill Gneiss. Radiometric ages include an Rb-Sr whole rock age of 1197 +/- 53 Ma by Douglas Mose (in Carl and others, 1990), and a zircon U-Pb age of 1183 +/- 7 Ma by Jeff Chiarenzelli (in McLelland and others, 1992). These intrusions are slightly older than most anorthosite, mangerite, charnockite and granite (the AMCG intrusive suite of McLelland, 1986) in the Highlands.

Antwerp granite at this outcrop has surrounded and isolated a segment of cream colored marble that is shaped like a "steer's head" (resembling an illustration in a Zane Gray western novel). Major minerals include microcline, plagioclase, quartz, hornblende and biotite. Minor minerals include tourmaline, magnetite, ilmenite, apatite, zircon, titanite and secondary chlorite. Note the veins of quartz and tourmaline in symplectic intergrowth.

Intrusive origin is not easy to demonstrate for many Lowland gneisses, given the multiple deformation and high-grade metamorphism. Dikes and sills within marbles are especially susceptible to rupture, displacement and rotation, and many of the contacts are tectonic in origin. Selvage of scapolite, diopside, mica and quartz occurs between the metagranite and marble at this outcrop, but the presence of selvage may be indicative of regional rather than contact metamorphism. Note at this outcrop a convolute, thin band of selvage that extends from the gneiss into marble (Fig. 6). The source of components in the band may lay within the metagranite. Other features indicative of an igneous origin include aplite veins, cross-cutting relationships and the distribution of carbon and oxygen isotopes at the granite-marble contact.

Marble and granite have been analyzed for oxygen isotopes at the Steer's Head outcrop by Cartwright and Valley (1991). The marbles away from the contact have delta ^{18}O values typical for Adirondack calcite (21-24 ‰). Antwerp gneisses away from the marbles have delta ^{18}O values within the range of Grenville granitic rocks (11-12 ‰). Within a few meters of the contacts, however, delta ^{18}O values increase in the metagranite and decrease in the marble. The result is a sigmoidal shaped delta ^{18}O profile with the inflection points occurring near the contact. This steep gradient is believed to have formed by high temperature, fluid-hosted diffusion of oxygen isotopes from one rock into the other. The question is whether the profile formed during regional or contact metamorphism.

Cartwright and Valley (1991) also observe a change in marble mineralogy toward the contact with the metagranite. Phlogopite content decreases and the diopside content increases, possibly because of a shift to the right in the following reaction:

$$\textit{phlogopite} + 3\ \textit{calcite} + 6\ \textit{quartz} = \textit{K-feldspar} + 3\ \textit{diopside} + H_2O + 3\ CO_2$$

This reaction buffers XH_2O and is favored by increased temperature during contact metamorphism. Another XH_2O buffering reaction involves tremolite as follows:

$$tremolite + 3\ calcite + 2\ quartz = 5\ diopside + H_2O + 3CO_2$$

Tremolite is described as disappearing towards the contact with the Antwerp metagranite, possibly as a result of a shift to the right in this reaction.

A change in the abundance and habit of graphite also is reported in the marble. Coarse (1-2 mm) crystalline flakes of graphite more than 1 m from the contacts give way to graphite free, or finer-grained, disseminated graphite near contact with the metagranite. This change probably resulted from oxidation during intrusion of the granite. Water diffusing from the granite probably caused the breakdown of graphite with the production of methane.

Cartwright and Valley (1991) argue that isotopic exchange occurred during contact metamorphism. Evidence includes the following: (1) Antwerp metagranite would have low porosity, and it is unlikely that enough grain boundary fluid remained in the granite to allow diffusion during regional metamorphism. (2) The systematic changes in marble mineralogy are indicative of contact metamorphism. The observed mineral assemblages also are stable at conditions of regional metamorphism, and little devolatilization would have occurred then. They argue that only during the contact metamorphic event was enough fluid available in the marble.

A larger conclusion drawn from this and other isotope studies at the contacts with metaigneous rocks is that large volumes of pervasive metamorphic fluids have not infiltrated these localities during the regional metamorphism, or at any time thereafter. The sigmodal-shaped profiles would not have been preserved had this been the case (Cartwright and Valley, 1991).

Continue south on Route 58/812.

30.2 (48.6 km) 0.7	Stop 5. Park along Route 58/812 near intersection with Smith road.

STOP 5 Basal, sillimanitic Popple Hill Gneiss.

We are near the base of a thick gneissic unit called the Popple Hill Gneiss and near the contact with Lower Marble. Generally, the gneiss is migmatitic, gray and layered (see stop 6), but here it contains rod-like sillimanite up to 12% by volume (Hudson and others, 1986). Lineations defined by the sillimanite plunge 8° SW. Also present are feldspar porphyroblasts and lenticular rotated garnets with helicitic textures defined by quartz and sillimanite inclusions (Ambers and Hudson, 1985).

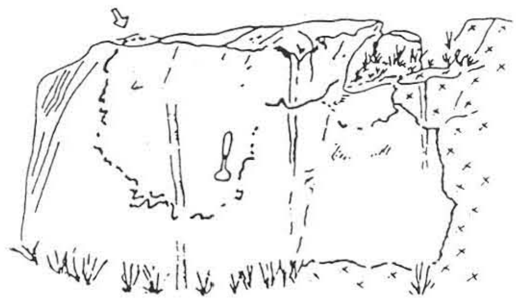

Figure 6 Antwerp metagranite (x's on right) and convolute vein that extends into the adjacent marble. Hammer for scale.

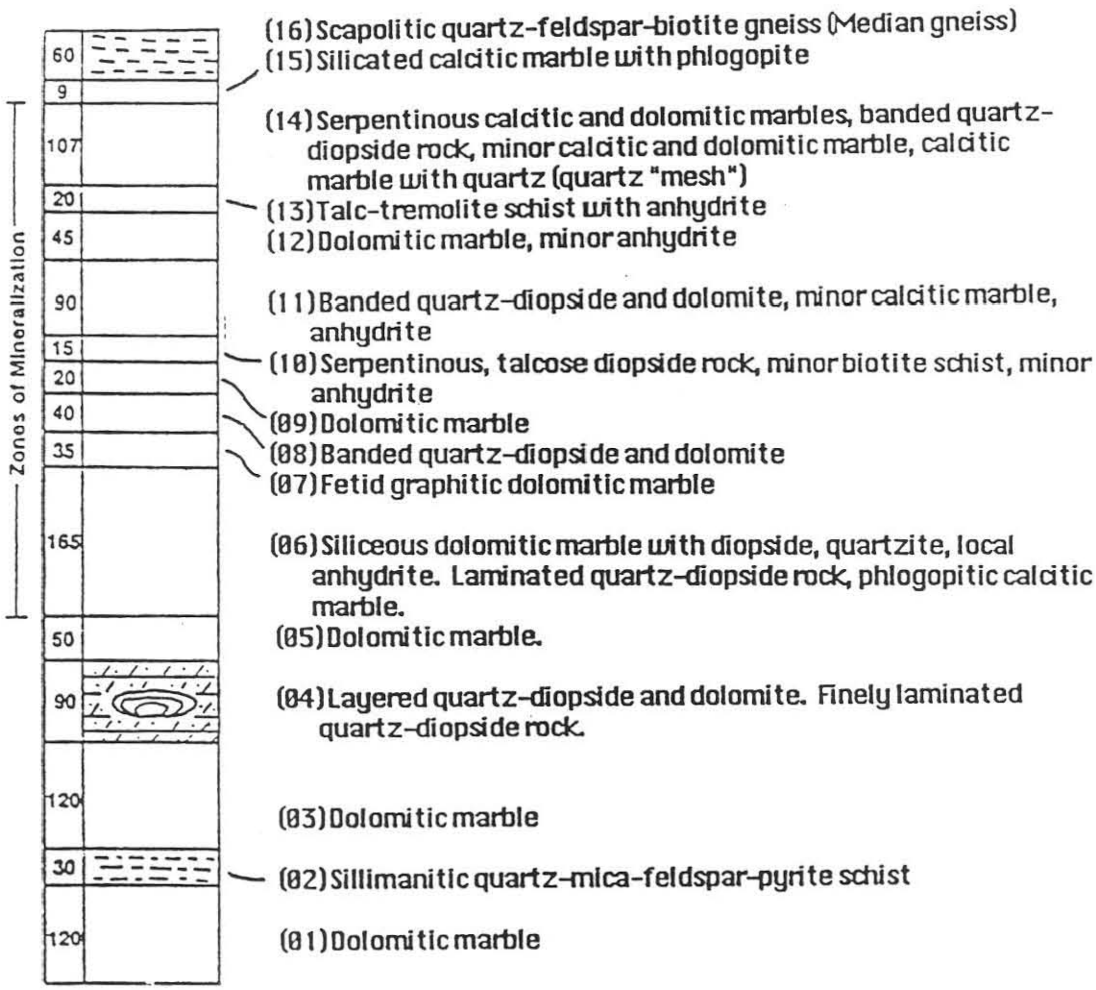

Figure 7 Stratigraphic column for the Upper Marble

This outcrop has been regarded by some as representing a pelitic protolith whose aluminous composition was appropriate for the growth of sillimanite. The presence of carbonate seams in the gneiss at Devil's Elbow and at Philadelphia village is compatible with an interpretation of a lime-bearing protolith. Stop 5, however, is proposed to be the site of ductile faulting that occurred during or near the peak of metamorphism, with the gneiss having moved northeasterly relative to the marbles in an essentially strike-slip motion (Hudson and others, 1986). In this model the presence of stress is believed to have influenced the growth of sillimanite, perhaps by shifting the following reaction to the right:

$$2\,feldspar + 2\,H^{+} = sillimanite + 5\,SiO_2(aq) + 2\,K^{+} + H_2O$$

(Wintsch and Andrews, 1988)

The mylonites are described as Sr-depleted compared to layered Popple Hill Gneiss from which they are presumed to be derived. A whole-rock Rb-Sr isochron of 1177 +/- 40 Ma is younger than ages obtained for the layered gneiss elsewhere and is interpreted as approximating the age of ductile deformation (Hudson and others, 1986). The age also approximates that of AMCG rocks in the Highlands and the Antwerp granite at stop 4; the age falls within the 1170-1130 Ma-interval of high grade metamorphism for the Lowlands as determined by Mezger and others (1991).

This mylonite zone has been extended to connect with the tectonic slide mapped by Foose (1974) near Moss Ridge. Informally called the Hailsboro ductile deformation zone, the mylonite zone may extend across the Lowlands. The Marble City member of Lower Marble lies against Popple Hill Gneiss at this outcrop, and stratigraphically higher members are absent. The gap depicted in our stratigraphic column at the base of Popple Hill Gneiss (Fig. 1) is based on this relationship and on mapping by Foose (1974). Ductile faulting also has disturbed the contact between Popple Hill Gneiss and the units within the Upper Marble in the zinc mines at Balmat. A tectonic slide on the lower limb of the Sylvia Lake syncline places unit 6 in contact with the Popple Hill Gneiss, units 1 through 5 having been excised (Fig. 5). Drill cores penetrating the gneiss generally show blastomylonitic fabrics whereas the overlying carbonate rocks generally are recrystallized "granoblastic mylonitic marbles."

Continue south on Route 58/812.

32.2 (53.0 km) 2.7	Stop 6. Park along Route 58/812 at the intersection with the Popple Hill road

STOP 6. Type locality of the Popple Hill Gneiss.

General:

This outcrop is typical of the migmatitic phase of the gneiss, and sillimanite is much less abundant here than at stop 5. Popple Hill Gneiss lies between two carbonate-bearing units, the Lower Marble of stops 1-4 and the Upper Marble in the Sylvia Lake syncline to the south (stops 7 and 8). The gneiss outcrops in a broad loop along the outer perimeter of the

Sylvia Lake syncline and in the Balmat zinc mining district. It outcrops continuously across the Lowlands in a hard-to-farm landscape of bedrock knobs for 70 km from Philadelphia to Colton village.

Because much of the protolith may consist of volcanic ash, Carl (1988) urged that the name be changed from the Major Paragneiss (Engel and Engel, 1953) to the colloquial expression for poplar trees and the hill through which this section of Route 58 was cut (see the road sign). Here is exposed the typical gray, layered, fine grained plagioclase-K-feldspar-quartz-biotite-garnet-sillimanitic gneiss that is strewn with boudins and convolute (ptygmatic) quartzo-feldspathic veins, megacrysts of K-feldspar, layered amphibolites, sill-like bodies of leucogneiss, and other metaintrusive rocks. Even a 1 cm thick seam of orthoquartzite extends vertically through the outcrop as a fracture-filling of Upper Cambrian Potsdam sandstone.

Geochronology:

A whole rock Rb-Sr isochron of 1296 +/- 21 Ma (Grant and others, 1984) and a zircon U-Pb age of 1214 +/- 21 Ma (Carl and Sinha, 1992) have been obtained for Popple Hill Gneiss. The discrepancy between the two ages has not been resolved. Both determinations were made on suites of samples from three outcrops that included one common locality here at Popple Hill. The 1.2-1.3 Ga-age of the gneiss must also apply to the Upper Marble whose stratigraphic position above Popple Hill Gneiss is well established by drilling.

The age of Popple Hill Gneiss is particularly important in view of the 1415 +/- 6 Ma-age of a leucogneiss dike on Wellesley Island in the St. Lawrence River (McLelland and others, 1988; 1992). Host rocks there must be much older than Grenvillian, perhaps >1500 my, but the extent of their distribution is not known. If these older rocks are widely distributed, then the ca. 1230 Ma-age assigned to Hyde School Gneiss by McLelland and others (1992) would be a revelation. Hyde School Gneiss would be younger than the surrounding rocks and, therefore, intrusive in origin (McLelland and others, 1988; 1992). The previously accepted model would be rendered invalid, i. e. that the Hyde School Gneiss protolith was volcanic tuffs (Carl and VanDiver, 1975), that the bodies represented apical projections of a folded but continuous sheet, and that the gneiss lay near the base of a Lowland stratigraphic sequence (Carl and others, 1990). These recent age determinations of Popple Hill Gneiss, however, do not support an intrusive origin for nearby bodies of Hyde School Gneiss, and the argument regarding an intrusive origin has shifted to other grounds.

A 40 m thick body of metagranodiorite that we regard as intrusive into Popple Hill Gneiss occurs at the northern end of this outcrop. It has given a whole rock Rb-Sr isochron age of 1147 +/- 35 Ma (Grant and others, 1984). This age indicates that the intrusion was contemporaneous with movement along the Hailsboro ductile deformation zone (stop 5) and with intrusion of the AMCG suite in the Highlands. Evidence for intrusive origin includes the young age, the presence of K-feldspar megacrysts (phenocrysts?), amphibolite that occurs as angular blocks (xenoliths?) rather than layers, and a more mafic composition than Popple Hill Gneiss. Contacts with the gneiss have been tectonized so that cross cutting relationships are not observed, or the body may be a sill. The intrusion is cut by numerous quartzo-feldspathic veins.

Migmatite and the Availability of Fluid during Metamorphism:

The belt of Popple Hill Gneiss crosses the isotherms of regional metamorphism that signify a transition from upper amphibolite to the granulite facies, from ca. 640°C in the SW to 740°C in the NE (Seal, 1986). Popple Hill Gneiss, thus, has the potential for study of mineral and chemical change as well as the availability of fluids during prograde metamorphism. The unit has been thoroughly sampled (Engel and Engel, 1953; 1958; Stoddard, 1980; Seal, 1986; Edwards and Essene, 1988; Ehrhard, 1986; Powers and Bohlen, 1985; Hoffman, 1982; see summary in Valley and others, 1990).

Quantitative estimates of the values for water activity are based on the equilibria of reactions that include:

(1) *annite (biotite with Fe in octahedral sites) + sillimanite + quartz = almandine + sanidine + H_2O,* and

(2) *muscovite + quartz = sanidine + sillimanite + H_2O.*

Calculated values of water activity range from 0.1 to 0.5 (Valley and others, 1990). Values are uniformly low, although not as low as in higher grade rocks in the Highlands. Values decrease from Gouverneur to higher grade metamorphic rocks to the NE near Colton.

Valley and others (1990) note that the gneiss is the largest migmatite unit in the Adirondack Mountains. The type locality here is repleat with convolute quartzo-feldspathic leucosomes, some parallel to the foliation and others cross-cutting. The low values of water activity recorded for the gneiss can be used to argue that the leucosomes had formed by partial melting. Water produced during dehydration of mica and other minerals would be partitioned into the incipient melts, and the unmelted residue would be further dehydrated and made harder to melt. In support of this idea, Valley and others (1990) point to the local variability in water activity throughout the unit, suggesting that fluid compositions may have been internally buffered by solids or by melt. Buffering is consistent with the presence of migmatite and with partial melting.

Leucosomes within the gneiss, thus, may be locally derived and not introduced. Carl (1988) proposed an *in situ* origin on the basis of (1) the lack of leucosomes in either Lower or Upper Marble formation, (2) the presence of Ba-, Sr-, and Rb-rich leucosomes in Popple Hill Gneiss outcrops whose mesosome is enriched in Ba, Sr and Rb, (3) the presence of heavy REE-depleted leucosomes where mesosome is similarly depleted, and (4) the dominance of K-feldspar leucosomes in layered gneiss versus plagioclase leucosomes in the amphibolite layers. An origin by metamorphic differentiation was proposed for those leucosomes with exceptionally high K_2O content that lacked minimum melt composition (Carl, 1988).

The studies of Valley and others (1990) are interpreted to rule out the passage of large amounts of fluids through Adirondack rocks during high grade metamorphism. On a local level, however, fluid migration into partial melts remains a possibility, and Popple Hill Gneiss may be unusual in that it retained much of the melt as leucosome (up to 25% at some outcrops). Higher grade gneisses in the Adirondack Highlands may have lost much of their melt fraction to consist chiefly of restite.

Minor folds are abundant in this outcrop. Many display overturned "S" asymmetries as viewed to the NE, suggesting that a major syncline lies to the east. In fact, the Sylvia Lake syncline is located east of this outcrop and may have influenced the orientation of minor folds. A number of folds, however, display "Z" asymmetries. Do these folds suggest a separate or later fold event? Not necessarily. Note that many folded veins exhibit cross-cutting relationships, and recall that fold asymmetries depend upon the orientation of the veins prior to folding. Thus, it is possible to have "S" and "Z" folds in the same outcrop that are coeval.

Continue south on Route 58/812.

33.9 (54.6 km)	1.0	Leave Route 58 by turning right (S) onto Route 812.
34.1 (54.9 km)	0.2	Turn right (N) onto the Sylvia Lake Road toward the mine and mill of Zinc Corporation of America.
35.3 (56.8 km)	1.2	Stop 7. Gate entrance to Zinc Corporation of America. Park on the right side next to the marble outcrops.

STOP 7. Stromatolites in Upper Marble, units 4 and 5.

NO HAMMERS AND NO SAMPLES, NOW AND FOREVERMORE (PLEASE).

Unit 4 of the Upper Marble (Fig. 7) in the Sylvia Lake syncline contains the first known biological remains described from the Adirondack mountains (Isachsen and Landing, 1983). This stromatolite-bearing rock is exposed in the woods above the roadcut. It consists of an alternating sequence of white dolomite, serpentinous-talcose diopside rock, and quartz lenses that may represent a chert-bearing, silty, peritidal dolostone intercalated between pure dolostones. Diopside formed by reactions between quartz and dolomite, and the rims of serpentine and talc owe their origin to retrograde metamorphism. Because the sequence is upside down, unit 5 underlies unit 4 in the roadcut. Unit 5 is a white to grayish, coarse-grained dolomite that contains sparce lenses of quartz and talcose diopside. Occasionally the dolomite is fetid.

The term stromatolite is applied to laminated carbonate sediment that occurs as bulbous heads or stacks. The laminations follow the outline of the structure and may be terminated at the edge of individual heads. Individual laminae may be thicker on the top and drape over the edge of the head to make a steep angle. Stromatolites are well displayed at this outcrop, and perhaps their late recognition can be attributed to an overturned position. The delay could also be attributed to frayed wiring in the twilight zone between the eye and brain of numerous observers, including your field trip guides who did not expect to see what they saw.

Stromatolites in unit 4 are domal SH-V stromatolites that occur as simple domes up to 12 cm high and 40 cm wide (Isachsen and Landing, 1983). They are composed of broad, convex laminae that consist of dolomite and iron-poor diopside (cream-colored rather than green), and they lack structural detail in thin section. The preservation of stromatolites may be due in part to their composition. Some are incompletely preserved with laminated quartz-diopside rock making up the preserved part. Silica replacement of algal structures during diagenesis can account for the preservation, whereas the unreplaced parts would recrystallize during metamorphism to coarse dolomitic marble. Fully replaced quartz-diopside mounds would form a rigid but minor structural buttress during deformation.

This outcrop lies on the upper, overturned limb of the Sylvia Lake isoclinal syncline. The stromatolites indicate that the upper limb is inverted and that this NNE-plunging complex fold is indeed a syncline as geologists of St. Joe Minerals (now ZCA) had proposed many years ago. Stromatolites have been recognized in units 4 and 11 of Upper Marble in recent years, and deLorraine has photographed domical stromatolites and finely laminated, possibly algal mat structures at the 2500 ft. level in the Balmat mine.

Turn around in the ZCA mine entrance.

36.4(58.6 km) 1.1	Turn right onto Route 812 South.
36.5 (58.8 km) 0.1	Bonnie's Diner, famous for home made pie.
37.6 (60.5 km) 1.1	Stop 8. Park on Route 812 just south of the Sylvia Lake fishing-access road in Balmat village.

STOP 8. This newly completed roadcut (late 1992) in units 14 and 15 of the Upper Marble formation (Fig. 7) represents a rebuilding of Route 812 that followed collapse of the old road into an underground mine.

This roadcut occurs in the core of the Sylvia Lake syncline in the Upper Marble formation at Balmat. Subunits of unit 14 comprise the northern part of the cut, whereas the brownish and green rocks in the south part belong to unit 15, a phlogopite/tremolitic calcitic marble (Fig. 7). Unit 14 is composed of a variety of rock types including quartz-calcite marble (the "quartz-mesh limestone" of Buddington), thinly layered quartz-diopside rock, and serpentinous dolomite and calcitic marbles, to name a few. At the southern hinge of the Sylvia Lake syncline near Balmat, minor fold plunges are northerly or slightly west of north. Since the sides of the roadcut here are parallel to the plunges of minor folds, it is difficult to see many folds on the outcrop faces. On the southern end of the cut it is possible to see folds developed in unit 15. Minor folds associated with the Sylvia Lake syncline deform the schistosity in units 14 and 15 and, hence, the minor and major folds are interpreted as second

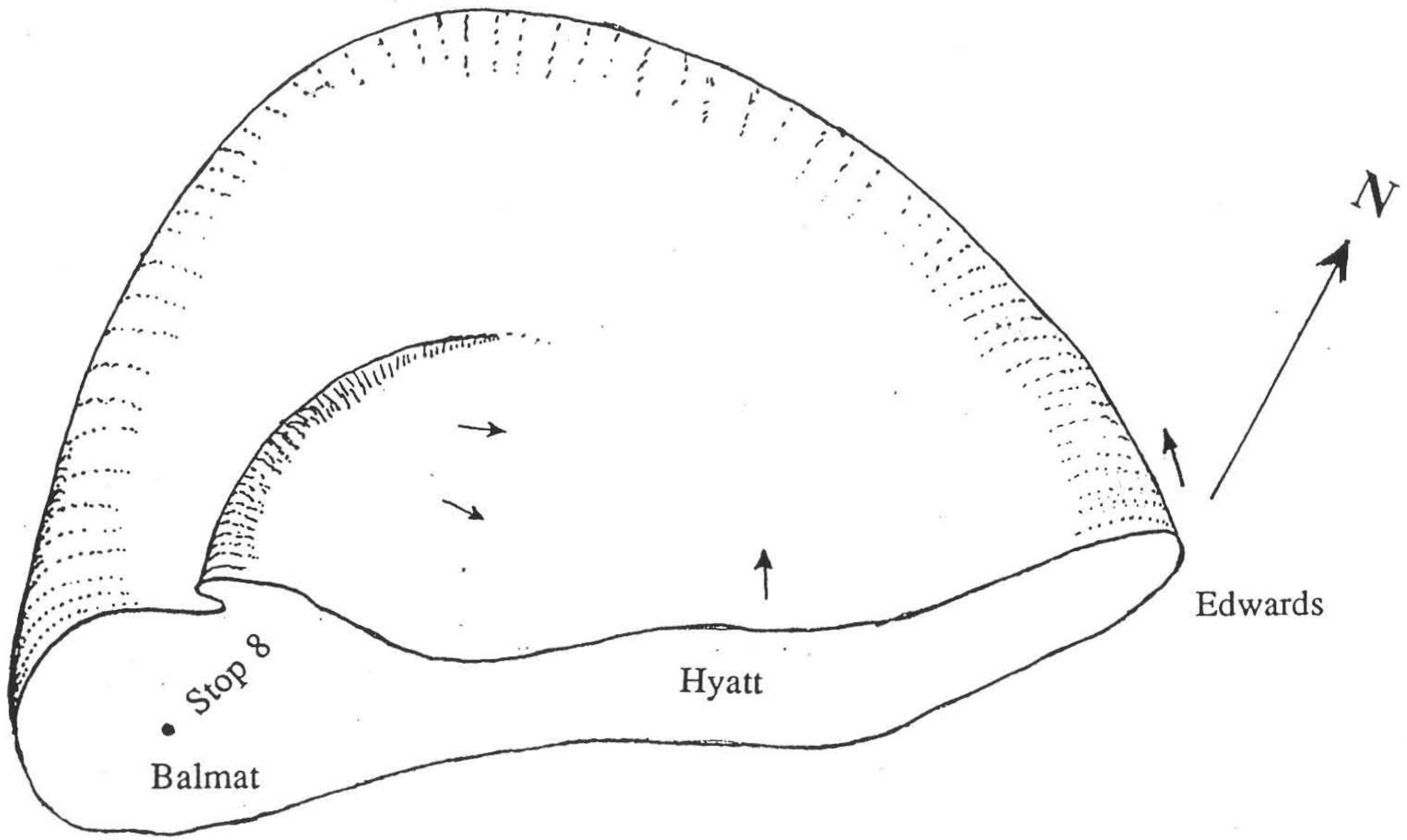

Figure 8 Sketch of the Sylvia Lake syncline as a sheath fold.

phase isoclines. Down plunge (in the mines) the fold axes swing to the NE. At the other end of the syncline at Edwards, fold axes plunge to the NW. The overall geometry of the fold, thus, is lobate, and the Sylvia Lake syncline is interpreted as a sheath fold (Fig. 8).

Turn around at the access site road. Go north on Route 812.

38.6 (62.1 km)	1.0	Bonnie's Diner (try the peanut butter pudding pie).
39.0 (62.8 km)	1.4	Intersection Route 58. Turn left (N) to Gouverneur.
45.3 (72.9 km)	6.3	Intersection Route 11 in Gouverneur. Continue straight ahead on Route 58 (N).
45.6 (73.4 km)	0.3	Sharp left turn to stay on Route 58. Proceed to Natural Dam and Brasie Corners.
47.4 (76.3 km)	0.8	The James River Corp. paper mill (on left) at Natural Dam.
50.7 (81.6 km)	3.3	Oswegatchie River bridge.
55.7 (89.7 km)	5.0	Turn right onto California road in Brasie Corners.
56.7 (91.3 km)	1.0	Flat solution valley that surrounds the Hyde School body of Hyde School Gneiss which is in view straight ahead.
59.5 (95.8 km)	2.8	Stop 9. Park on side of road.

STOP 9. Amphibolite layering in the Hyde School body of Hyde School Gneiss (Fig. 9).

Layering: Large and Small Scale

Hyde School Gneiss with numerous, thin biotitic-amphibolite layers is exposed in the smooth bedrock knobs overlooking a small pasture. We are located at the southern end of an F_2 fold that parallels the Hyde School antiform (see map of Brown, 1988). The orientation of numerous minor folds in the amphibolite layers can be related to the Hyde School antiform, interpreted here as a major, second phase sheath fold. The antiform deforms an earlier fabric, including NW-SE trending lineations and rootless isoclines that may be related to Tewksbury's early shear fabric. Nearby are isoclinally folded amphibolite layers whose axes trend NW-SE, perhaps having been rotated into parallelism with the direction of early phase, NW-directed tectonic transport.

We are near the contact between two pink alaskite units of Brown (1988), one (aa) distinguished by thin parallel layers of biotitic amphibolite, and the other (al) a massive unit with fewer layers (Fig. 9). Another mappable unit (pgd) consists of gneisses ranging in composition from biotite granite, trondhjemite, quartz monzonite, granodiorite to quartz diorite. All are foliated and contain amphibolite layers. Detailed mapping of the large-scale layering enabled Brown to interpret complex refold patterns, and we will discuss his work in the shade of a small tree.

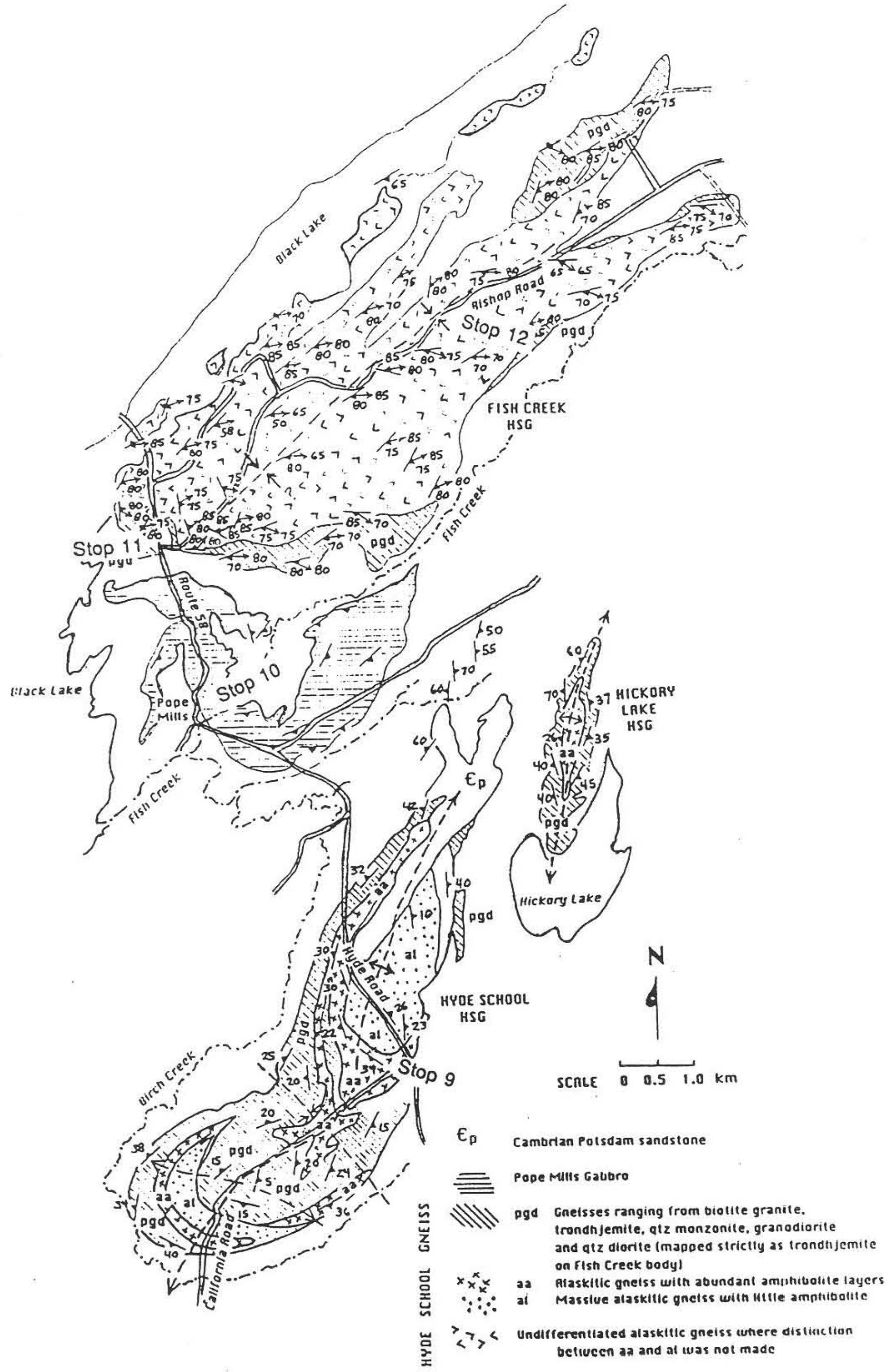

Figure 9 Location map for the Pope Mills Gabbro, Hyde School and Fish Creek bodies of Hyde School Gneiss. Data from Dietrich (1957), Buddington (1934), and Brown (1988).

Hyde School Gneiss, thus, differs from granitoid plutons of the Adirondack Lowlands in being compositionally zoned on a grand scale. Brown (1988) wrote that "the alaskite, granite, and diorite gneisses of the antiforms are a parallel-layered sequence that, despite abundant evidence of internal mobility and local melting, have not intruded each other or broken out of their elliptical structural shells as a melt." In contrast, the younger Lowland granitoid rocks show abundant evidence for intrusive origin. The Rockport granite and Gananoque syenite occur as dikes, sills and apothyses within metasedimentary rocks that outcrop along the St. Lawrence River. Even the oldest rock recorded so far, a 1415 Ma-old leucogneiss on Wellesley Island, is described as a dike within calc-silicate and tourmalinite layers (McLelland and others, 1988).

One objective of this stop is to observe the scale of the amphibolite layering, folding, and boudinage. These features are characteristic of Hyde School Gneiss. Note the absence of other lithologies or inclusions. No other granitoid bodies in the Lowlands exhibit layering like this, and most, if not all, have xenolithic inclusions of country rock attesting to their igneous provenance. Hyde School Gneiss is unique in being restricted to one stratigraphic horizon, containing layer-parallel amphibolites, and in the absence of xenolithic inclusions of the adjacent host gneisses. Feldspathic quartzites, calc-silicate and garnet-sillimanite gneiss occur within bodies of Hyde School Gneiss but, like the amphibolites, they also occur as parallel layers generally < 1 m thick.

Thus, Hyde School Gneiss consists of *conformable* lithologic layers on a large and small scale. Does the layering result from tectonism during intrusion, tectonism after intrusion, or is it a primary feature related to the deformation of ash and lava beds? Do the amphibolites represent basaltic dikes that were realigned prior to complete consolidation of the granitoid magma, so that igneous breccias and textures are preserved? Do the amphibolites represent dikes that were realigned strictly by solid state deformation? Do they represent slab-like sections of amphibolite wall rock that had been loosened during shearing and permissive intrusion of the granitoid magma? All scenarios have been proposed in one form or another.

A less complicated approach for the origin of Hyde School Gneiss begins with a layered complex of alternating ash-flow tuffs and minor metasediments, capped in some cases by outpourings of more fluid lavas and ash of trondhjemic and dioritic composition. Solid state deformation, with localized partial melting in low pressure sites, and late stage pegmatite intrusion, would account for features cited by others as evidence for intrusive origin. If the amphibolite layers were once basaltic dikes that had intruded the complex, then transposition to parallelism with the surrounding rocks had to occur *prior to* the 1170-1130 Ma period of high grade metamorphism in the Lowlands, because the amphibolite bodies were deformed after they had acquired the layered form. If the amphibolite layers represent original lithologies, e.g. protoliths of iron-rich, calcareous shales or basaltic ash, then deformation could have occurred simultaneously with high grade metamorphism.

High Temperatures and Contact Metamorphism?

The most compelling evidence for intrusive origin of Hyde School Gneiss cited by McLelland and others (1992) is the presence of perthite within the gneiss and the presence of a high grade mineral assemblage (garnet, sillimanite and spinel) in gneisses that surround Hyde School Gneiss. Perthite may form from exsolution of hypersolvus feldspar, the existence of which implies temperatures higher than those proposed for Lowland metamorphism.

It is interesting to note that the high grade mineral assemblage might have been used as evidence for higher temperatures of *regional* metamorphism had the assembly been found elsewhere than adjacent to Hyde School Gneiss. A search is underway to determine if the assemblage occurs in pelitic rocks elsewhere in the Lowlands. Given the paucity of Lowland pelitic rocks, and the remarkable selectivity that Hyde School Gneiss magma presumably has shown in intruding only pelitic horizons, we think it appropriate that the upper ranges of temperatures proposed for Lowland metamorphism be reconsidered. Also needed are U-Pb ages of garnet from the sillimanitic gneisses adjacent to Hyde School Gneiss. If the ages are >1200 my (similar to the age of the gneiss) then contact metamorphism is a possible cause. If the ages are <1200 my, then regional metamorphism is likely.

The proposal that stratigraphic principles cannot be applied in the Lowlands goes hand in glove with the model of intrusive origin for Hyde School Gneiss. Pelitic rocks can be conveniently distributed where needed as a host for granite intrusions. One may dismiss the present view of Popple Hill Gneiss as a lithologic/stratigraphic unit confined to the south-central Lowlands. It may be presumed to be distributed elsewhere, intruded by Hyde School Gneiss magma and partially melted so that the residue is left as a garnet-sillimanite gneiss that surrounds Hyde School Gneiss (McLelland and others, 1992).

On the other hand, we suggest that Hyde School Gneiss may represent a layered volcanic complex overlain by the lowest member of Lower Marble formation, whose pelitic composition was favorable for growth of garnet, sillimanite and spinel during the period of high grade regional metamorphism from 1170 to 1130 Ma.

A Stratigraphic View of Hyde School Gneiss:

The picture that emerges from detailed field mapping of members of Lower Marble formation is that the bodies of Hyde School Gneiss occur at the same stratigraphic horizon across the Lowlands. Each body is associated with the Black Lake member of the Lower Marble Formation. It is difficult to imagine how independent intrusions could acquire such similar surroundings unless the carbonate cover was deposited unconformably over each dome as proposed by Wiener and others (1984). Previous references to the gneiss appearing at various horizons across the Lowlands are incorrect. The California body of Hyde School Gneiss was cited as being in contact with Popple Hill Gneiss, with the inference being that the contact was intrusive in origin (Buddington, 1929; McLelland and others, 1992). However, the marble is exposed on the surface, and Zinc Corporation drill core shows that the Black Lake member of Lower Marble occurs between Hyde School Gneiss and Popple Hill Gneiss.

Xenoliths of adjacent metasedimentary rocks occur in Hermon-type and in most other metaintrusive rocks throughout the Lowlands, particularly at the margins of a body. Diligent search has produced few xenolithic candidates in Hyde School Gneiss other than amphibolite. Cushing (1916) observed as much when he wrote of inclusions in Hyde School Gneiss as consisting of amphibolite "no matter what the nature of the bordering Grenville rock is" (p. 18).

The conformable amphibolite layers in Hyde School Gneiss have been interpreted as slabs that were stoped and separated along foliation planes by incoming magma (Levy and others, 1993). This process requires the fortuitous convergence of the following unrelated factors: (1) The necessary presence of vast quantities of amphibolitic material immediately below the Black Lake member of Lower Marble. (2) The necessary requirement that the batholith intrude but one stratigraphic horizon and that it stoped only the amphibolite layers. Apophyses, dikes and sills of Hyde School Gneiss are non-existent in the surrounding metasedimentary rocks. (3) The requirement that the magma be extremely fluid in order to incorporate slabs that vary in thickness from 1 to 60 cm or more. Many so-called slabs have uniform thicknesses for tens of meters. (4) The requirement that magma of stock-like to batholithic dimensions was confined to one or several substrata of appropriate pelitic composition that, in turn, could be subjected to contact metamorphism to produce the ever-present garnet-sillimanite gneiss.

Continue straight ahead to intersection with Route 7 (Hyde road).

59.7 (96.1 km)	0.2	Turn left onto Route 7 toward Pope Mills.
62.7 (100.9 km)	3.0	Intersection with Route 184. Turn left (W) toward Pope Mills.
63.5 (102.2 km)	0.8	Intersection with Route 58 in Pope Mills. Continue on Route 58.
64.3 (103.5 km)	0.8	Stop 10. Park along Route 58. Do not climb the outcrop to the right (E) onto the lawn of the trailer.

STOP 10. The Pope Mills gabbro near Pope Mills village (Fig. 9).

The purpose of this brief stop is to examine a folded and metamorphosed mafic intrusion within the Lower Marble formation as a prelude to examination of the Fish Creek body of Hyde School Gneiss. The Pope Mills gabbro was interpreted by Buddington (1934) as a relatively large mass of "pyroxenic amphibolite" that had been sheared prior to intrusion of syenitic magma. The intrusion of syenite into foliation planes, with the help of mineralizers, was believed responsible for the "migmatitic character" of the body, particularly the K-feldspar megacrysts and augen that are strikingly displayed at outcrops further to the south. The body was isoclinally folded and overturned towards the northwest with the dip of foliation varying from 55 to 70° SE (Buddington, 1934).

Xenoliths of banded, greenish calc-silicate from the Black Lake member of Lower Marble occur within the metagabbro. At least three phases of the gabbro are recognized: (1) subrounded to angular and block-like forms of fine grained gabbro that are surrounded by (2) a slightly coarser grained gabbro matrix. Also present, but not at this outcrop, is (3) a biotite and K-feldspar megacrystic gabbro that Buddington regarded as a mixed rock. Examine the weathered horizontal surface on the west side of the road, and note the calc-silicate xenoliths that form a tectonically re-aligned intrusion breccia. Also note the deformed, fine grained gabbro ellipsoids whose long axes trend N 75 E.

Representative chemical analyses of the three gabbroic phases are given in Table 1. The coarser grained matrix (sample 223) is richer in total Fe, TiO_2 and P_2O_5 than the finer grained phase (sample 222). We tentatively regard the finer grained gabbro as autoliths, and the coarser grained matrix as a later, more differentiated, iron-rich phase of the same magma.

Continue north on Route 58.

65.1 (104.8 km)	0.8	Stop 11. Park on Route 58 south of the Bishop road (right) and walk northward to the outcrop.

STOP 11. SW margin of Fish Creek body of Hyde School Gneiss (Fig. 9).

Hyde School Gneiss is exposed in the hinge/axial region of the Fish Creek anticline. Nowhere could a more challenging and complex outcrop be found, because these cuts expose highly deformed alaskitic and trondhjemitic gneisses with parallel layered amphibolites in the core of a regional "anticline that looks like a syncline" (i. e. a sheath fold). Amphibolite layers may have undergone an early deformation that resulted in "chocolate-tablet" style boudinage before being refolded in the hinge of the Fish Creek anticline. Pegmatites with feldspar, green clinopyroxene and titanite assemblages intrude the suite of leucogneisses and also may appear at or near the contact with the overlying Black Lake member. Did the pegmatitic fluids derive some constituents from the carbonates upsection? Mafic dikes, probably offshoots from the Pope Mills metagabbro, are present and superfically resemble the amphibolites. Other intrusive rocks also may be present in these outcrops.

The following scenario for the Fish Creek body is proposed by deLorraine: Begin with a "normal" sequence of Hyde School Gneiss, the trondhjemite phase in contact with overlying Black Lake metasediments that are exposed at the south end of the outcrop at Bishop road. Rocks are subjected to high grade metamorphism and deformation with temperatures at the higher end of those reported in the literature. The earliest phase of deformation produced chocolate-tablet boudinage in the conformable amphibolite layers, accompanied by partial melting in the neck zones of boudins. Pegmatites are intruded, possibly channeled along the interface between Hyde School Gneiss and the Black Lake member. Offshoots of the Pope Mills metagabbro are intruded, and one wonders about contributions to the local heat budget that may have facilitated partial melting in Hyde School Gneiss.

Table 1 Representative samples of a Fish Creek amphibolite layer (190A), the Pope Mills gabbro (222 = fine grained; 223 = coarser grained; K-feldspar megacrystic = 56B), and a basaltic dike (198B) in the Fish Creek body.

	DF190A	DF222	DF223	DF56B	DF198B
SiO2 (wt. %)	48.06	51.29	48.71	51.65	48.79
Al2O3	14	15.46	15.48	16.96	16.03
Fe as Fe2O3	14.49	10.37	13.94	10.26	14.12
MgO	6.54	6.67	5.45	3.46	4.85
CaO	10.8	9.44	9.3	5.77	7.1
Na2O	3.61	3.43	3.16	3.79	4.46
K2O	0.85	1.11	1.4	4.14	1.93
TiO2	1.8	1.13	2.03	1.74	1.88
P2O5	0.29	0.28	0.56	0.91	0.32
MnO	0.24	0.16	0.17	0.11	0.15
LOI	0.17	0.73	0.59		1.45
Rb (ppm)	7	14	24	65	33
Sr	371	606	691	913	473
Ba	176	550	735	2209	581
Y	39	37	45	44	46
Zr	103	55	148	538	178
Nb	1	6	12	6	10
Cr	157	215	91	55	15
Ni	72	92	33	0	36

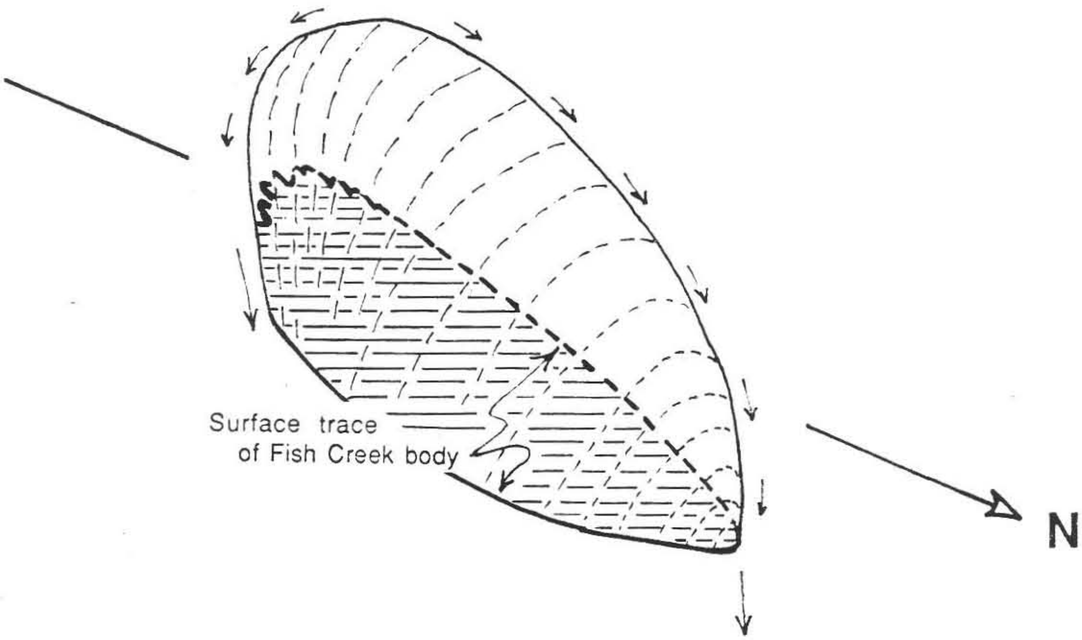

Figure 10 Sketch of a model of the Fish Creek body of Hyde School Gneiss as a sheath fold.

Phase two isoclinal sheath folding then affected the region, producing the NE-SW regional grain. The second phase, sheath fold hinge in the SW end of the Fish Creek anticline resembles an "anticline that behaved like a syncline" because of the northerly plunges of a profusion of minor tight to isoclinal folds (see proposed sketch of the structure of the Fish Creek body, Fig. 10). "M and W" minor folds in the hinge of the anticline were superimposed on amphibolite layers that previously had undergone chocolate-tablet boudinage. Partial melting would be enhanced in the core of the Fish Creek anticline because it would be a low pressure site. Magmatic heat contributions from the sill and dikes offshoots of the Pope Mills metagabbro also may have facilitated partial melting. So-called "intrusion breccias," thus, are best interpreted in a regional framework as broken, boudinaged fragments of conformable amphibolite layers which have been refolded and deformed in the solid state by "M and W" folds. The best place to verify that the amphibolite layers are not randomly or chaotically distributed is to observe them from the top of the outcrop on the west side of the road. Look to the north down the plunges of minor folds.

Turn right (E) onto Bishop road.

66.9 (109.0 km)	1.8	Intersection (left) with Mitchell road. Continue on Bishop road.
69.7 (107.7 km)	2.8	Stop 12. Park on Bishop road.

STOP 12. Mafic layers in the Fish Creek body of Hyde School Gneiss, pasture of the Yoder farm (Fig. 9).

One purpose of this stop is to distinguish late dikes that may be related to the Pope Mills metagabbro (stop 10) from amphibolite layers of the Fish Creek body of Hyde School Gneiss. Metagabbro dikes are shown to post-date the foliated amphibolite layers and an early phase of isoclinal folding. Although the dikes have been deformed and metamorphosed (as is the case with the Pope Mills gabbro), they are clearly younger than the amphibolite layers.

The dikes and amphibolite layers have been regarded as coeval. Cross-cutting relationships as observed here have been used as evidence that all mafic layers had formed as dikes which, in turn, had been transposed shortly after their emplacement within Hyde School Gneiss magma (McLelland and others, 1992). Phenocrystic plagioclase and coarse subophitic orthopyroxene have been used as evidence for an intrusive origin (Levy and others, 1993), but we believe it is the dikes that have been described rather than true amphibolite layers. The dikes have abundant orthopyroxene in various stages of alteration to hornblende. The amphibolite layers are chiefly amphibolite and biotite with little pyroxene.

The Fish Creek body was mapped by Dietrich (1954; 1957) who agreed with Buddington that the Hyde School Gneiss was intrusive in origin (he later changed his mind to an origin as

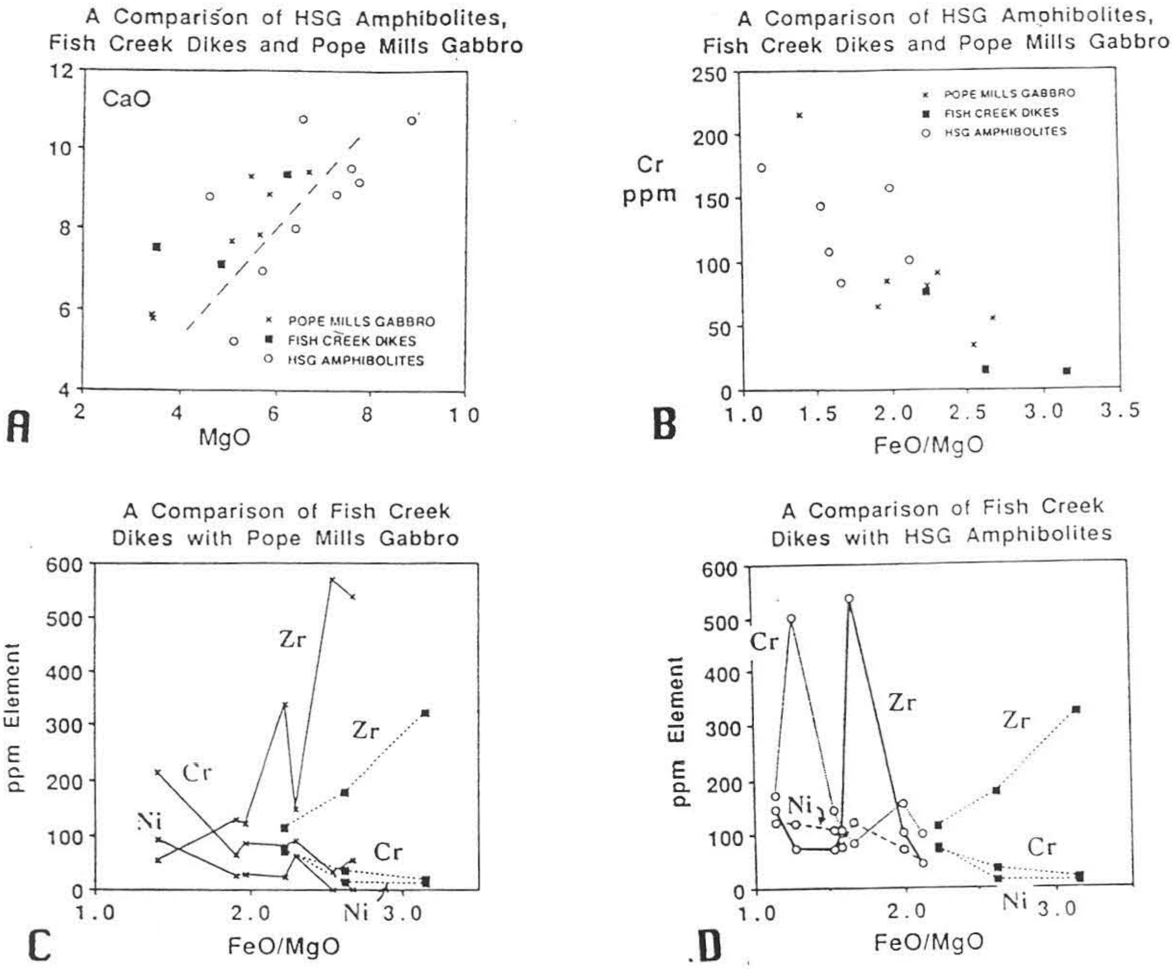

Figure 11. A geochemical comparison of mafic rocks. Pope Mills gabbro (x's), Hyde School Gneiss amphibolite layers (circles) and Fish Creek metabasaltic dikes (solid squares). (A) Pope Mills gabbro and the dikes contain less MgO than most amphibolite layers. (B) Pope Mills gabbro and the dikes are deficient in Cr relative to the amphibolite layers. (C) Trends of elemental abundances versus FeO-MgO ratios. Elemental abundances for the gabbro (x's) vary systematically with increasing Fe-Mg ratios, as is the case for the basaltic dikes (solid squares). (D) Elemental abundances for the amphibolite layers (circles), however, vary erratically with increasing Fe-Mg ratios, with the exception of Ni.

in situ anatectic magma; Dietrich, 1963). The disrupted amphibolite layers were described as xenoliths throughout the 1957 report. In addition to the amphibolite layers, Dietrich recognized 18 tabular "amphibolitized melanocratic dikes," ranging from nearly 2 ft. to a few inches in width (see photos p. 57-58 in Dietrich, 1957).

We will examine several gabbroic dikes exposed in the pastures on both sides of Bishop road. One of the dikes cuts several amphibolite layers, including one that has been isoclinally folded. The dike has been realigned to near parallelism with the amphibolite layers at one end of the outcrop. A 1 m thick dike elsewhere in the pasture cuts several amphibolite layers at an acute angle. One layer can be traced into the dike where it reappears as ruptured fragments that are true xenoliths. Clearly the amphibolite had acquired its foliation and layered form prior to intrusion of the dikes.

We have observed these dikes only in the Fish Creek and Stalbird bodies of Hyde School Gneiss. They are particularly abundant in the Fish Creek body, perhaps because of the presence of the Pope Mills gabbro.

We compare the chemistry of (1) dikes in the Fish Creek body, (2) Hyde School Gneiss amphibolite layers, and (3) Pope Mills gabbro (Table 1). In general, Hyde School Gneiss amphibolites contain more MgO, Cr and Ni than the Pope Mills gabbro. Advocates of a metaigneous origin for the amphibolite layers may regard enrichment in these elements as indicative of a lack of olivine fractionation in their source. We have used plots of MgO vs CaO (Fig. 11A) and FeO-MgO ratios vs Cr (Fig. 11B) to distinguish Hyde School Gneiss amphibolite layers from Pope Mills gabbro. Fish Creek dikes plot with the gabbro.

Fish Creek dikes resemble the Pope Mills gabbro in a tendency toward iron enrichment. Both groups show increasing Zr (and TiO_2 and P_2O_5) content and decreasing Cr and Ni (and CaO) content with increasing Fe-Mg ratio (Fig. 11C). In contrast, the amphibolite layers generally have lower Fe-Mg ratios and show relatively erratic changes in abundances of these elements with increasing iron content (Fig. 11D). We tentatively regard the dikes as fractionated, iron-enriched derivatives of the Pope Mills gabbro. More samples are needed.

Continue straight ahead to the turnaround.

70.1 (112.9 km)	0.4	Turn around in the driveway of the red barn. Retrace the route on Bishop road.
75.0 (120.7 km)	4.9	Intersection with Route 58. Turn left (S) to Pope Mills and Gouverneur.
76.7 (123.5 km)	1.7	Pope Mills. Turn right to follow Route 58.
91.3 (147.0 km)	4.6	James River Corp. paper mill at Natural Dam.
93.1 (149.9 km)	1.8	Stop sign in Gouverneur. Turn right to stay on Route 58.
93.4 (150.4 km)	0.3	Junction with Route 11. Turn left (NE) onto Route 11 toward Richville and Canton.

100.4 (161.6 km)	7.0	Abrupt right (SE) turn uphill towards Richville.
100.6(162.0 km)	0.2	Turn left (N) onto Main St. in Richville.
101.2(162.9 km)	0.6	Stop. Intersection with Route 11. Turn left (SW) and park on the right side of Route 11 adjacent to the outcrops of dark rock.

CAREFUL HERE. TRAFFIC MOVES FAST.

STOP 13. Tourmalinites and quartzites with a thinly layered appearance.

Exposed here are fine grained, gray to deep maroon, tourmaline-bearing metasedimentary rocks that include quartz-feldspar-mica-diopside gneisses, calc-silicates and quartzites. Note the inclined, 1 m thick layer of dark, glassy quartzite that contains tiny, vitreous grains of dark tourmaline. These grains could be overlooked or mistaken for biotite. We correlate the rocks with tourmaline-bearing gneisses and quartzites of unit "qmt" (Fig. 1) in the Beaver Creek area (Brown, 1989). Lithologically similar rocks have been intruded by the Rockport pluton in the Thousands Islands area, suggesting that "qmt" or its equivalents may be of widespread occurrence in southeastern Ontario.

Tourmalinite is abundant in the Lower Marble and may have crystallized from boron-rich basinal brines during diagenesis and metamorphism (Brown and Ayuso, 1985). Feldspathic quartzites and quartz-feldspar-mica granofels contain layers with up to 50% tourmaline whose composition is dravite-uvite with 0.85 to 4.25 wt.% Na_2O, 0.39 to 4.04% CaO, and up to 13.67% MgO (which is near end-member dravite composition). Li-rich compositions have not been found.

Tourmaline also occurs in sheet-like bodies of metagranite that intrude "qmt," but the tourmaline is compositionally distinct from that in the metasedimentary rocks (Brown and Ayuso, 1985). Ratios of FeO-(FeO+MgO) range from 0.55 to 0.75 in granites and pegmatites, compared to ratios of 0.15 to 0.58 for tourmalines in quartzites. Previous workers believed that the tourmaline in the metasedimentary rocks had originated from the granitic intrusions, but the reverse may be true. Intrusive rocks may have picked up (and compositionally modified?) the tourmaline from the metasedimentary rocks, as is suggested by the distribution of the tourmaline-bearing intrusions. Abundant tourmaline occurs in the metagranite that intrudes "qmt" rocks near Huckleberry Mountain (Richville quadrangle), whereas tourmaline is less common in granites that have intruded other rock units.

Some rock types within the unit "qmt" are intensely limonite-stained by the weathering of pyritic zones. Pyrite mining was carried out as a source of sulphur and sulphuric acid for the paper industry at Pyrites, Stellaville and elsewhere in northern New York in the late 19th and early 20th centuries.

Go straight ahead (SW) on Route 11.

101.7 (163.7 km)	0.5	Turn right (NW) onto the Richville Bridge road.
102.3 (164.7 km)	0.6	Cross the Oswegatchie River bridge and turn right.
103.1 (166.0 km)	0.6	Turn right onto River road.
107.0 (172.3 km)	3.9	Stop 14. Park along this narrow road.

STOP 14. Hermon-type granitoid gneiss of the Gray's School body.

At least two textural types of granitoid metaintrusive rocks were mapped in the Lowlands at the turn of the century, a fine-grained, so-called "equigranular" gneiss of the Alexandria-type that included the Hyde School Gneiss, and a coarser grained, K-feldspar megacrystic gneiss called the Hermon-type (Cushing and others, 1910). The mapping of numerous small and isolated granitic bodies throughout the Lowlands eventually led to the adoption of local names and to the abandonment of the term "Alexandria-type." But the name Hermon-type is still applied to relatively small, sheet-like and irregularly shaped, K-feldspar megacrystic plutons that are scattered throughout the Lowlands.

Hermon gneisses are generally reddish, medium to coarse grained feldspar-biotite-hornblende-quartz-titanite gneisses of syenitic to granitic composition (note the paucity of quartz at this outcrop). They contain large K-feldspar megacrysts that occasionally are oriented transverse to foliation. This coarse grained rock may give way along strike to become a fine grained equigranular gneiss. Pervasive grain size reduction may be responsible, and K-feldspar augen, flaser structure and quartz ribbons are often observed in thin section in association with a mosaic textured groundmass.

Igneous textures have survived the metamorphism and ductile deformation. Concentrically zoned megacrysts of K-feldspar are observed in thin section to contain zones of patch or speckled or "fire flake" perthite that alternate with wider zones or bands of non-perthitic feldspar. Plagioclase grains and biotite flakes lie against the perthite zones and protrude outwardly from them. We interpret these grains to be Frasel or epitatic inclusions that originally adhered to the face of a K-feldspar crystal that crystallized from a melt. The rims of fine grained pink feldspar that surround the phenocrysts were produced by ductile deformation.

Hermon gneisses generally occur as intrusions within the Lower and Upper marbles and the Popple Hill Gneiss. Conformity with the host rock, due in some cases to tectonism, has caused some to be mapped as lithologic units. They generally occur as sills and less often as dikes or apothyses. Xenoliths of amphibolite are common.

Each intrusive body of Hermon gneiss is characterized by a limited range in major element content. For example, samples from the Gray's School body (including this outcrop) are uniformly syenitic, low in SiO_2 and rich in MgO (Fig. 12; see sample 101A in Table 2). Gneisses near Hermon village are of intermediate SiO_2 content (sample 178), and samples of high SiO_2 and low MgO content (sample 124) are characteristic of the Mott Creek body. We believe that these rocks represent a common intrusive series.

We compare Hermon gneisses with another Lowland metagranite, the "equigranular" Rockport gneiss which outcrops in the islands and along the shore of the St. Lawrence River. Formerly called the Alexandria Bay Formation, this rock was cited by Wiener and others (1984) as equivalent to Hyde School Gneiss and, thus, was made to represent the basal stratigraphic unit in the NW Adirondacks. Rockport gneisses are quartz-feldspar-biotitic gneisses with less hornblende and titanite than Hermon gneisses. Fluorite is a notable accessory phase, as is tourmaline where tourmaline-bearing host rocks have been intruded (Brown and Ayuso, 1985). Rockport gneisses clearly show an intrusive origin as dikes, sills and stock-like masses. Geochemical and field differences suggest that they represent a more highly differentiated and less viscous magma than Hermon-type magma.

Hermon gneisses show a tendency toward calc-alkaline character (Fig. 13A, B) that contrasts with a pronounced alkaline character for Rockport gneisses. Both gneissic groups plot chiefly in the field of "volcanic arc" granites on an Rb vs Y + Nb tectonic discrimination

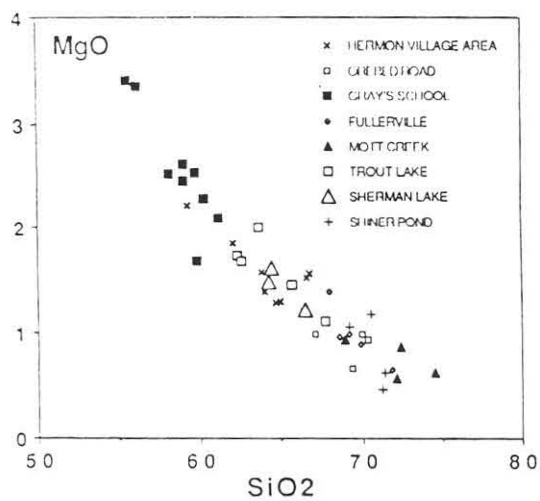

Figure 12 Plot of MgO versus SiO_2 to demonstrate that different intrusive bodies of Hermon-type granitoid gneisses have relatively consistent compositions. The syenitic Gray's School body is exposed at stop 14.

Table 2 Representative samples of Hermon gneisses (top) and Rockport gneisses (bottom). Sample 101A is from stop 14.

HERMON	#101A	#178	#124
SiO2 (wt. %)	58.99	66.78	74.56
Al2O3	16.76	16.54	14.16
Fe as Fe2O3	5.58	4.03	1.78
MgO	2.45	1.56	0.63
CaO	3.88	2.19	1.4
Na2O	4.52	2.57	3.91
K2O	5.51	5.53	4.19
TiO2	0.71	0.61	0.17
P2O5	0.4		0.06
MnO	0.08	0.01	0.06
Rb (ppm)	134	142	129
Sr	915	423	241
Ba	1018	1119	493
Y	44	35	36
Zr	415	267	114
Nb	13	12	0
Ga	18		17
Ce	192		51
ROCKPORT		#3F	#71A
SiO2 (wt. %)		68.37	73.64
Al2O3		14.42	13.62
Fe as Fe2O3		4.29	1.94
MgO		0.42	0.1
CaO		1.39	1.07
Na2O		4.24	3.49
K2O		5.45	5.17
TiO2		0.45	0.15
P2O5		0.05	0.01
MnO		0.07	0.03
Rb (ppm)		105	205
Sr		175	92
Ba		466	417
Y		49	62
Zr		571	136
Nb		10	7
Ga		21	17
Ce		269	80

A1

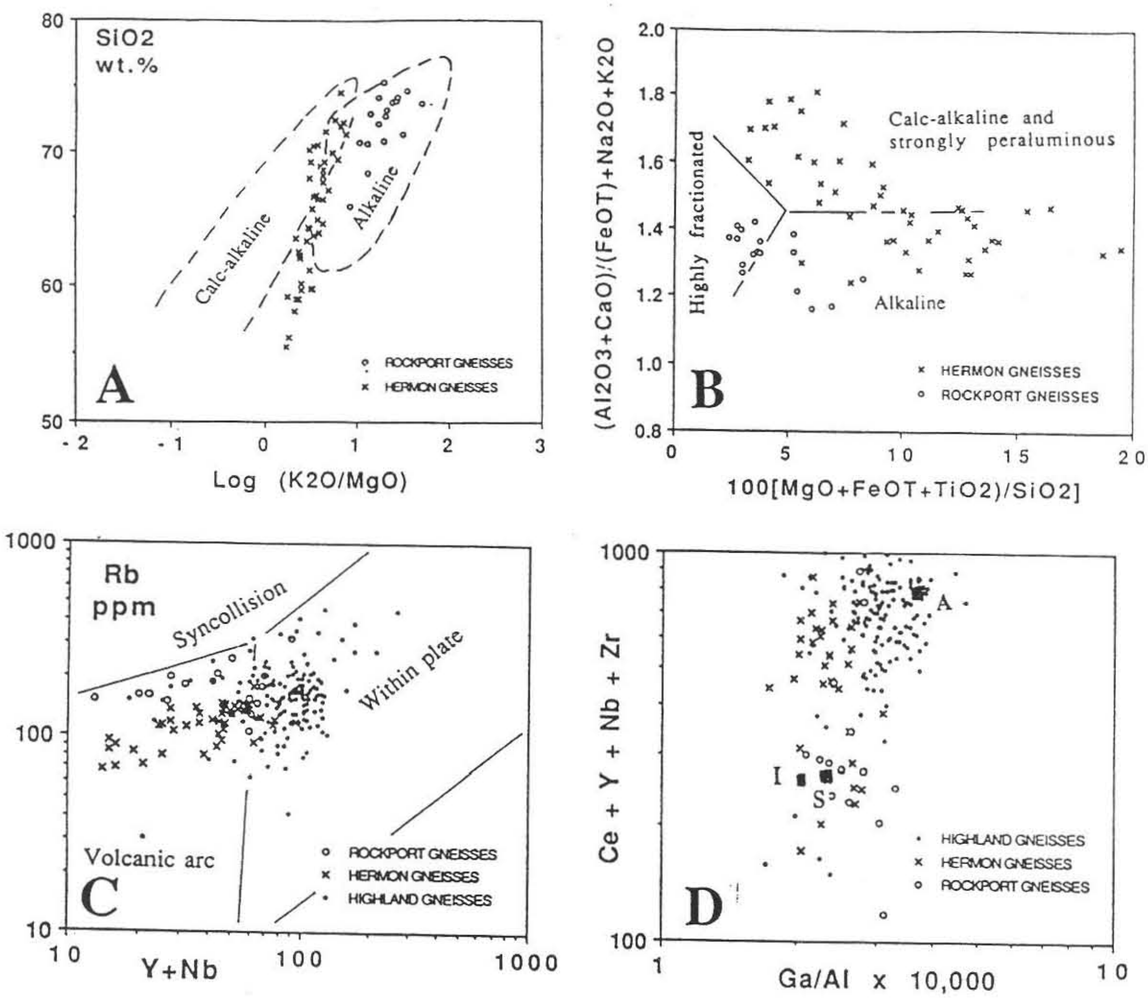

Figure 13 (A) Plots of (K_2O/MgO) versus SiO_2. Hermon gneisses (x's) are transitional between calc-alkaline and alkaline. Rockport gneisses (circles) are alkaline. Fields are based on data from island arc and continental-margin suites, including the Sierra Nevada batholith, after Rogers and Greenberg (1981). (B) Major element classification of granites after Sylvester (1989) who views the "highly fractionated" category as a variety of alkaline granite. Hermon gneisses are transitional between calc-alkaline and alkaline, whereas Rockport gneisses are alkaline. (C) Tectonic discrimination diagram after Pearce and others (1984). Hermon and Rockport gneisses plot chiefly as volcanic arc granites. Highland data (dots) from Whitney (1992, and personal communication, 1991) and McLelland and Whitney (1990). Included are 12 samples taken by us from lower Lake Durant Formation in the Highlands. (D) Plots used to distinguish among I-, S- and A-type granites after Whalen and others (1987). Hermon (x's) and Rockport gneisses (circles) are widely scattered, whereas Highland granitoid rocks (dots) cluster near A-type granite.

diagram (Fig. 13C). Plots of both gneisses are widely scattered among I, S and A-type granites (Fig. 13D). In contrast, granitoid rocks of the AMCG series in the Adirondack Highlands (Figs. 13C, D) plot chiefly as alkaline, A-type "within plate" intrusive rocks (McLelland, 1986; McLelland and Whitney, 1990; Daly and McLelland, 1991).

Proceed straight ahead on River road.

108.3 (174.4 km)	1.3	First stop sign. Turn right onto Maple Ridge road.
108.5 (174.7 km)	0.2	Second stop sign. Turn right.
109.1 (175.7 km)	0.6	Oswegatchie River bridge and intersection with Route 812. Turn right on Route 812 to Dekalb.
109.3 (176.0 km)	0.2	Turn left onto Route 17 in Dekalb toward Dekalb Junction.
110.3 (177.6 km)	1.0	"Y" in the road. Keep left toward Dekalb Junction.
113.3 (182.4 km)	3.0	Intersection with Route 11 in Dekalb Junction. Turn left (NE) toward Canton.
121.9 (196.2 km) the	8.6	Intersection with Park St. in Canton. Turn right (S) to return to SLU physical plant parking lot.

WE HOPE YOU HAVE HAD A GOOD TRIP.

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TRIP A2

THE POTSDAM-GRENVILLE CONTACT REVISITED (I)

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INTRODUCTION

The purpose of this field trip is to examine the relationships at the boundary between Proterozoic Grenville gneiesses within the Adirondack Lowlands and overlying arenaceous rocks, presumed to belong to the Potsdam Sandstone of Upper Cambrian (?) age. It is the first of two that will explore the diverse mineralogy, lithology and structure of a number of arenaceous outcrops in the Canton-DeKalb-Gouverneur area and will provide an overview of the compositional and structural characteristics of some of these problematic rocks.

(Please note that trips A2 and B2 can be taken together, as a single long day sequence, or independently. This fieldguide includes a repeat in Trip B2 of Trip A2 stops 7 and 8 to conveniently provide a shorter day's review.)

The Potsdam-Grenville contact in the St. Lawrence valley represents a hiatus of some 500 Ma. Mineral assemblages within the Grenville gneisses indicate that perhaps 25 km of material was eroded prior to the deposition of the transgressive Cambro-Ordovician sequence, over an undulating Precambrian erosion surface. Many of the isolated, predominantly sandstone, bodies within the Grenville gneiss terrane of the Adirondack Lowlands can be confidently interpreted as outliers of Potsdam Sandstone. The focus of this field trip, however, is to question the affinity of those whose relationship to the Cambro-Ordovician sequence is equivocal. Such enclaves vary in composition from matrix supported breccias to equigranular orthoquartzites and their depositional environment has been variously interpreted as: pre-Potsdam solution pocket infills (related to a karst topography on a Grenville marble surface); fault related debris slides; fault breccias; and fault scarp talus accumulations. Some are seemingly crudely interlayered with the gneisses and are locally foliated. Rare sandstone/breccia dykes within the gneisses may be of significance in understanding the origin of these enigmatic, apparently sedimentary, clastic rocks.

This fieldtrip commences at exposures close to the St. Lawrence Seaway between Ogdensburg and Alexandria Bay - including the text book exposures of the Potsdam - Grenville unconformity east of Alexandria Bay (Fig. 1). This will take from 25 to 40 mins., depending on traffic conditions. We then head south towards the village of Theresa and staert our investigation of the arenaceous enclaves within the gneiss terrane. In the process we will visit some of the classic outcrops of the St. Lawrence Valley and Adirondack Lowlands where arenaceous rocks in contact with demonstrable Precambrian are of uncertain age and origin. Some of these rocks are seemingly deformed along with late Grenville deformation (that is, exhibiting a qualitative strain level not seen in the Potsdam Sandstone; Trip B2) whereas others exhibit intrusive relationships to the gneisses (Trips A2 and B2), or lie structurally beneath them (Trips A2 and B2). The origin of these enigmatic relationships have variously been proposed as:

1) outliers of Potsdam Sandstone;

2) pre-Potsdam, post-Grenville in age;

3) Grenville in age, in that they have enjoyed at least the later stages of Grenvillian deformation.

It is likely that all of these relationships are represented on this fieldtrip - but the question we ask, since most relationships are at least superficially ambiguous, is - "how can we discriminate between them". We hope for fruitful discussion on outcrop that will lead towards a better understanding of the status of these "enclaves". Some of these outcrops have long been recognized as contentious and constitute a longstanding problem. It would be nice to move closer to solving it!

ROAD LOG

Cumulative mileage	Miles from last point	Route description
Start		Junction of Route 11 and Park Street in the center of Canton. Head southwest, cross the Grasse River, and continue straight (onto Route 68 towards Ogdensburg)
16.9	16.9	Junction of Route 68 and Route 37. Turn left (southwest) onto Route 37 and proceed towards Morristown, passing St.
23.6	6.7	Lawrence State Park on the right.

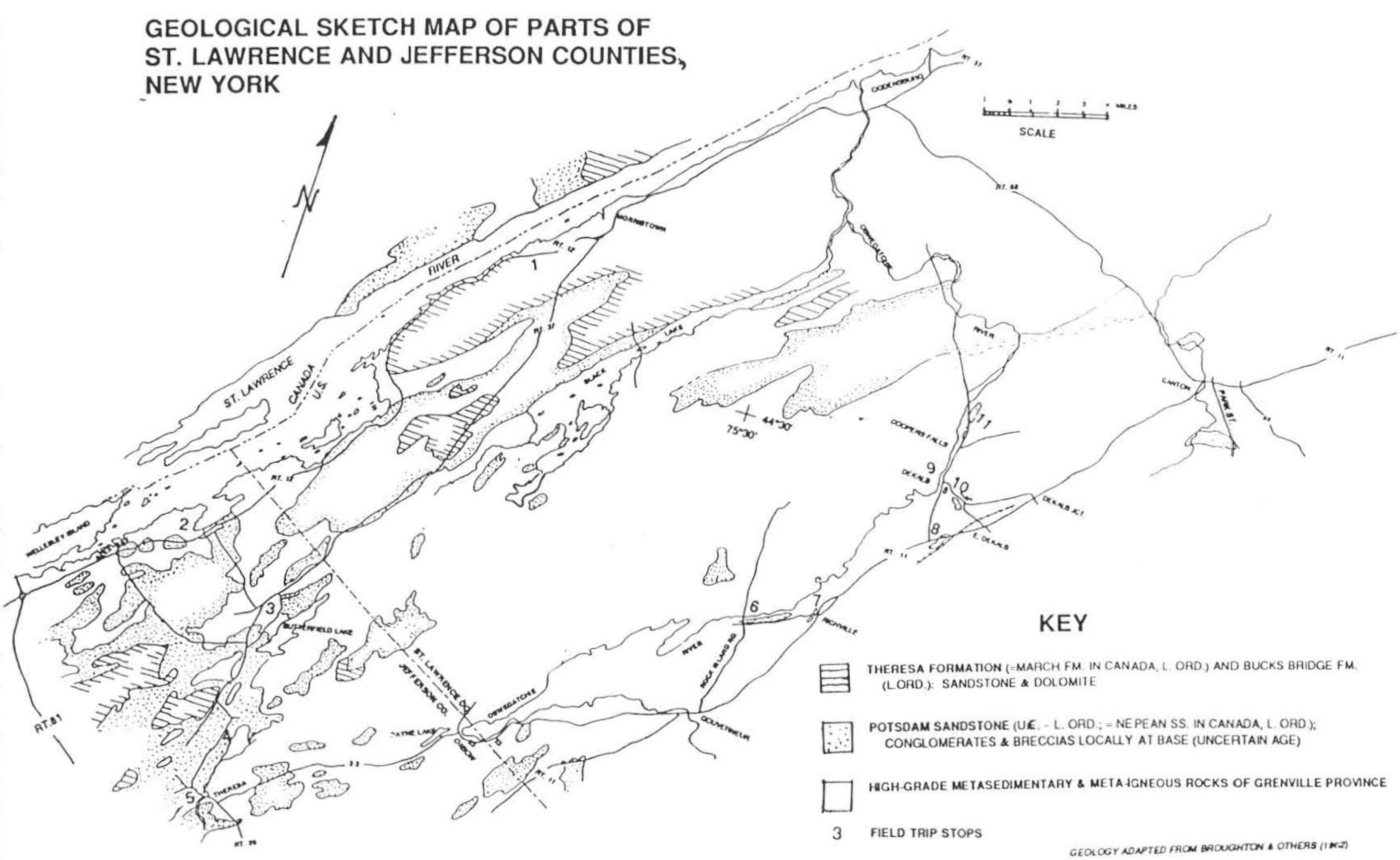

Figure 1: Distribution of arenaceous rocks on Precambrian basement in the Adirondack Lowlands. Location map for stops mentioned in the text.

28.8	5.2	Proceed westwards, bearing right onto Route 12
30.6	1.8	Pass entrance road to Jacques Cartier State Park on right
31.9	1.3	Stop 1. Outcrops of Theresa Formation (see Trip B2, Stop 2, this volume)

STOP 1: Deformation in middle Theresa Formation

Gentle northeasterly-trending folds of Potsdam Sandstone and Theresa Formation are relatively common to the south and west of Theresa (Barber 1977). The examples in this road section are similar in geometry to those observed in that area. They are significant to this field trip in that they clearly demonstrate significant post Potsdam regional deformation. Folds in the Potsdam Sandstone north of Theresa will be seen (Stop 4, Figure 2, this trip) are considered likely to be the same age. A minor fault will be seen in the central part of this section.

Continue on Route 12 passing through some magnificent roadcuts of Theresa Formation and Potsdam sandstone that are featured in Trips A3 (Erickson) and B2 (Selleck). Continue past the village of Chippewa Bay, across marshy ground on the southern shore of the St. Lawrence.

50.2	18.3	Very large outcrops on both sides of road with an obvious contact at their eastern end.
50.4	0.2	Continue to a turnaround (side road on right) and proceed eastwards back to through the roadcut to its eastern end and
50.7	0.3	park. This is Stop 2.

PLEASE WATCH FOR FAST MOVING TRAFFIC

STOP 2: The Postdam Grenville unconformity at Alexandria Bay.

This classic example of an unconformity shows typical, regularly layered, lowermost Potsdam Sandstone lying on deformed Grenville gneisses. Some slip has occurred along the contact resulting in small-scale buckling of a subvertical foliation within the gneisses on the north side of the road.

The nature of the unconformable relationship and the character of the Potsdam Sandstone immediately above the contact at this stop will provide a basis for discussion at subsequent outcrops and is typical of the basal Potsdam in this area, although no trace fossils have been observed (Selleck, this volume). Note the rare conglomeratic horizons and the well-defined bedding.

Continue northwest on Route 12:

51.4	0.7	Pass Log Hill Road on the right.
51.7	0.3	Turn right onto Jefferson County Route 1, up a sharp rise onto the plateau underlain by the flat-lying Paleozoic sequence which is characteristic of this section of the St. Lawrence Valley.
52.3	0.6	Turn right (southeast) onto Limestone Road and continue through the junction at Skinners Corners
55.65	3.35	Turn left at T-junction
55.85	0.2	Junction with State Route 37. Turn left (northeast) and proceed towards Hammond. Note exposures of gneisses in fields to the right. The wooded higher ground ahead on both sides of Route 37 are underlain by sandstones. An excellent example of the unconformity at their base occurs in a roadside outcrop approximately two miles further north.
56.9	1.05	Turn right onto Stein Road. Drive across a bridge, noting gneisses to the north, and head up the escarpment ahead to the junction with Spies Road on the left. Park, and walk back down the hill for approximately 50 yards.
57.2	0.3	Stop 3. Cylindical structure in lowermost "Potsdam"

STOP 3:. Soft-sediment deformation close to the Precambrian erosion surface.

Unusual columnar structures within the sandstones of this region have been described from a number of localities (Van Diver, 1976; Dietrich, 1953). They measure from a few inches in diameter to 10 feet or more. Characteristically they possess a cylindical or cone-like structure with a narrow structureless border zone that cross-cuts the bedding in the surrounding rocks. Concentric sub-vertical layering paralleling the margin may occur and, in the larger examples, the interior may be brecciated. This internal breccia may be comprised of laminated sandstone fragments set in a homogeneous sandstone matrix. This particular example, which must occur close to the Precambrian surface beneath well demonstrates these relationships. Abrupt truncation of bedding within the surrounding sandstone will be seen at the eastern margin of the structure and a brecciated central zone may be observed; individual clasts are typically poorly defined The origin of this and similar structures is not fully understood, but is thought most likely related to localized collapse of semi-consolidated sediment - that is the downward movement of a column of sand. A possible cause is settling or collapse of cavities in the underlying rocks - an hypothesis resulting from the observation that the location of these features is within sandstones seemingly overlying marbles within the underlying Proterozoic sequence. Subsequent stops will support this notion. The homogeneous sandstone matrix surrounding breccia fragments suggests a lack of grain coherence (and resulting aggregate fluidity) that may have been initiated by high pore fluid pressures. The details and mechanics of such an environment will be discussed at this and subsequent outcrops. Similar structures are common elsewhere within the Potsdam Sandstone of this region and in equivalent rocks of the Ottawa basin to the north (Grabs personal communication 1979).

Turn around and head back to Route 37:

57.5	0.3	Junction with Stein and Route 37. Turn left (south) and continue through Redwood on Route 37
61.25	3.75	Pass sign for Crystal Lake on left
62.25	1.0	Turn left onto Joyner Road (at sign for Thousand Islands Zoo)
63.2	0.95	Turn right onto County Route 21 and follow this, down a moderate hill onto a flat open area
64.15	0.95	Stop 4. Park as far on the limited shoulder as possible. Outcrop of deformed "Potsdam Sandstone" on the east side of the road across a disused railroad line.

STOP 4: Folds within the "Potsdam Sandstone" (Figure 2)

PLEASE NOTE THE POISON IVY AT THIS STOP

Further evidence of gentle large-scale folding within arenaceous rocks of presumed Potsdam age is exhibited here (Figure 2). These exposures of sandstone are typical for the area. Bedding is pronounced, with the rock taking on a 'flaggy' appearance locally. Hematite staining may accent layering, including cross-bedding, but also is apparently not controlled be internal structure inplaces cross-cutting the sedimentary structures. Such 'liesegang banding' is quite common throughout the Potsdam Sandstone in this region. Evidence for soft sediment deformation is present and a small monocline behind a pine tree in the center of the face could be a product of differential settling.

Barber and Bursnall (1978) described the folding in this outcrop as follows: "Bedding is folded in a manner typical of the lower part of the Potsdam Formation. Hinge -lines are tenuous and variable in orientation, and individual beds thicken and thin greatly through short distances. The most obvious fold at this locality is one of a series of irregular gentle undulations of bedrock, between which are small areas underlain by Grenville rocks or horizontal Potsdam strata. The hinge line of the fold trends approximately N40^0E. Other folds in the area exhibit a similar trend ....". They suggested that the spatial relationship of many of these folds and internal soft sediment disruption of bedding to topographic highs of the Precambrian basement (in this case caused by a possible west-side down northeast-trending fault 2 miles to the southeast near Theresa village) might have produced "rapid downslope movement during and closely following sedimentation" (Barber and Bursnall, op cit.).

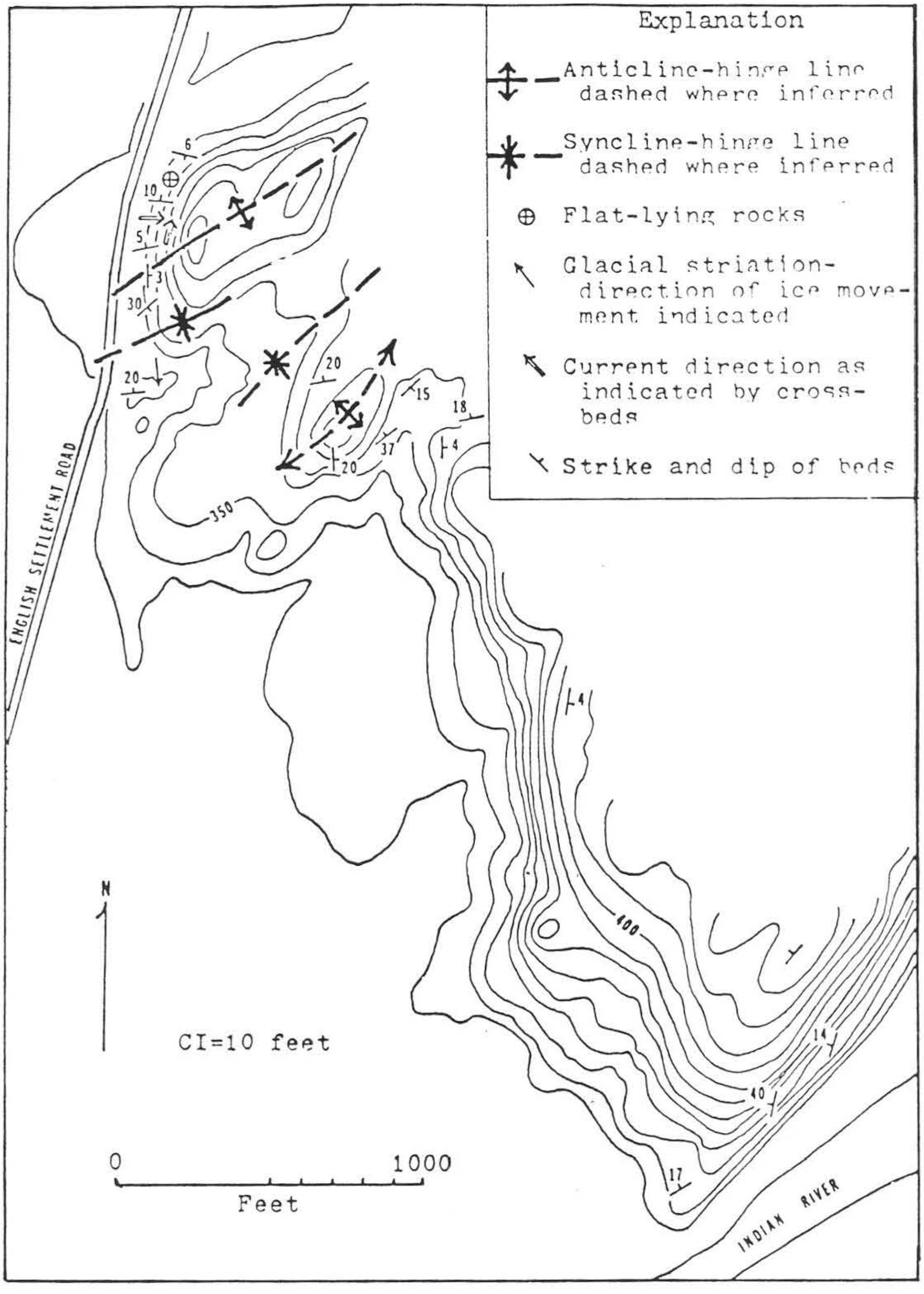

Figure 2: Fold trends within Potsdam Sandstone northwest of Theresa (from Barber and Bursnall 1977).

Continue south on Route 21:

66.25	2.1	T-junction. Turn left.
66.45	0.2	Intersection with State Route 26. Turn left and drive up the hill (fault scarp referred to above) into the center of Theresa.
67.0	0.55	Turn left at top of hill onto State Route 26 and proceed eastwards on Route 26 through the village.
67.2	0.2	Pass the reservoir for the Theresa village hydro plant and a large, overhanging outcrop on the right (unfortunately due for demolition). This is Stop 5, but parking is restricted so continue across a bridge, find a convenient parking space and walk back to the outcrop.

STOP 5: Contact between Grenville gneisses and non-typical "Potsdam" sandstones.

This area exhibits considerable relief of the Proterozoic surface (consider the elevation change between this locality and Stop 4) and the local configuration of the unconformity reflects this. The contact between sandstone and weathered basement gneiss dips moderately to the east from a high point at the western end of the outcrop. Significant discordance in bedding attitude exists between the low dip of well-bedded and color banded "typical Potsdam" at the highest parts of the outcrop to moderate easterly dips at road level. A bluff overhanging the road exhibits well-bedded sandstone overlying a heterogeneous breccia of quartz-arenite at the base. The larger clasts are of quartz and sandstone and towards the base, within a few feet of the unconformity, thin pebbly horizons are present. The weathered surface suggests the possibility of large sandstone blocks or rafts encased in an otherwise structureless mass and healed microfaults are present towards the top of this lower section.

These features suggest a period of instability and possible slumping prior to the deposition of the well-bedded sandstones in the upper part of the exposure. Barber (1977) suggested that penecontemporaneous faulting might be the cause for the lower disrupted material and, although the gneiss-sandstone relationship here might be explained by initial dip of the Precambrian erosion surface, the existence of non-typical basal Potsdam Sandstone at this stop raises the possibility of a pre-Potsdam but post-Grenville depositional period accompanying fault-related deformation. This interpretation requires the existence of a cryptic depositional break within the outcrop and that the presumed slumping is not intraformational with respect to the present exposed section.

Return to vehicle and continue east on Route 26:

67.65	0.45	(measured from Stop 5) Turn left (northeast) onto Jefferson County 194.

67.9	0.25	Continue straight, onto Route 22 towards Oxbow.
76.8	8.9	Follow this to the junction with County Route 25, passing Payne Lake on the left at approximately 7.5 miles. The complexly deformed Payne Lake alaskite body underlies the topographic high to the north (see Tewksbury, Trip A5, this volume).
		Turn right and drive through the village of Oxbow.Continue on Route 25, which continues as Route 52 in St. Lawrence County, passing through Wegatchie and closely following the path of the Oswegatchie River. Note that the flow is to the southwest here - at a position upstream of a major change in direction at Oxbow, from where the river heads northeastwards to the St. Lawrence at Ogdensburg.
		Continue on Route 52 to the junction with Route 11 (traffic lights) in Gouverneur. Turn left, cross the Oswegatchie River and continue on Route 11, pass the "mint-with-the-hole" statue and the village park on your right.
77.4	0.6	Turn left onto Rock Island Road
81.1	3.7	Turn right onto Welch Road and park. Stop 6.

STOP 6: "The Rock Island Roadcut". Complex relationships of sandstone to gneiss.

This outcrop is perhaps one of the most intriguing in the North Country. It is featured in Van Diver's Rocks and Routes of the North Country and is a common stopping point for field parties visiting the area. Our main interest here is to characterize the relationship of the presumed Potsdam Sandstone to the Grenvillian gneisses which at the upper part of the roadcut are calcitic marbles. This deep roadcut provides magnificent exposure and is one of thebetter outcrops in illustrating the variety of marble - sandstone relationships present in the region.

At the southeast end of the cut, a well-bedded quartzarenite illustrates the attractiveness of the Potsdam sandstone erosional remnant interpretation. However, the very complex relationships of the sandstones with the "underlying" marbles invite alternative explanations to a simple erosional unconformity relationship. Moderate dips (~30°), angular marble clasts in quartzarenite, thin sandstone dikes extending outwards and upwards from contacts, and changes in composition of the sandstones close to the marble have all contributed to a karst surface depression infill (Van Diver, 1976). Some of the contact relationships in this section have ben used as evidence for structural control of deposition and basement remobilization (Bannerman *in* Carl and Van Diver, 1971).

Contact zones need to be carefully examined. Iron oxides and iron pyrites armor many of the surfaces, and the potential significance of angular marble fragments, subtle brecciation features particularly in the northern sections (perhaps, autobrecciation), matrix supported

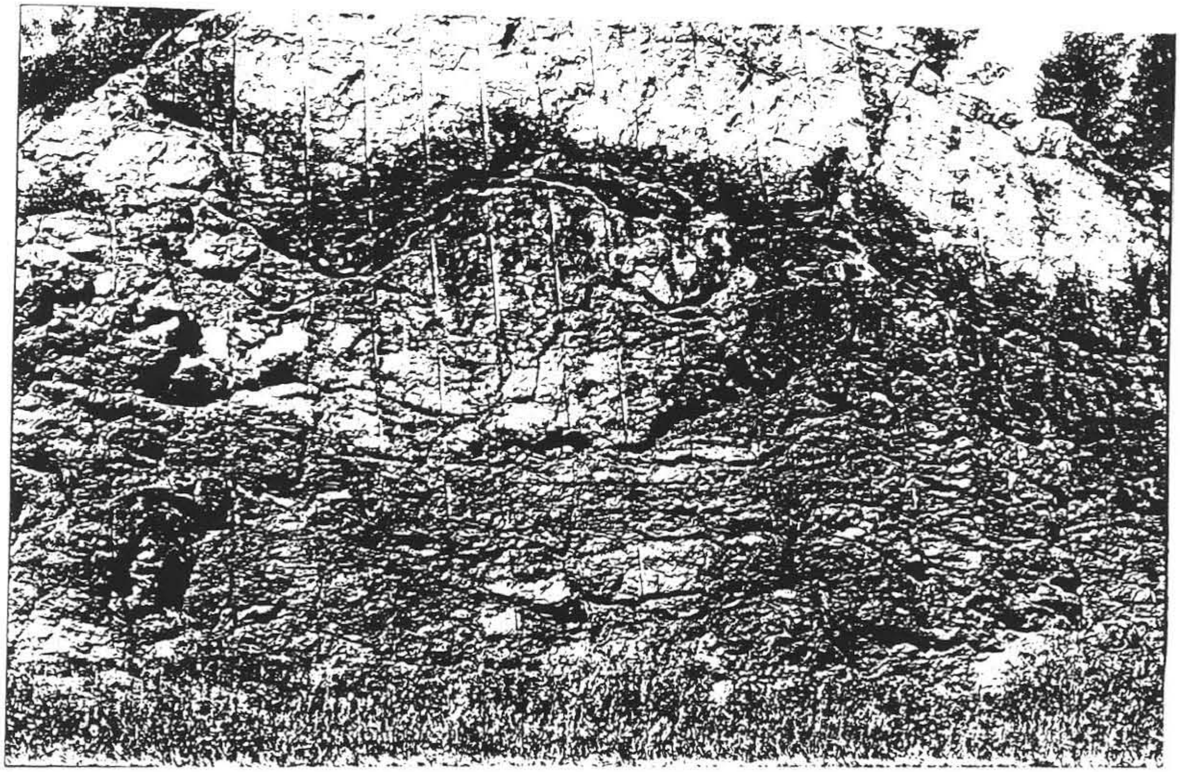

Figure 3: View of the western roadcut at Stop 7. Note quartzarenite and metaquartzite at top of outcrop and the overall structural grain. Breccia lenses are just discernible at bottom of outcrop.

conglomerates (upper west side of northern section), etc., are easily overlooked. A spectacular tourmaline-rich breccia in the latter area further suggest a protracted and complex history for this section. Evidence of significant post-arenite deformation is present at many localities.

Return to vehicle and continue east along Welch Road towards Richville.

84.0	2.9	Junction with Route 11. Turn left and park on the shoulder. Stop 6 includes the outcrops on both sides of the road

BEWARE OF FAST MOVING TRAFFIC !

STOP 7: The Richville breccias.

Please note that description of this stop is repeated in Trip B2

Cross the road and visit the moderately high exposures on the northwest side first.

Karboski et al. (1983) described this outcrop this as containing a "flow breccia" at its base, followed by a densely consolidated breccia with an overlying highly deformed metaquartzite, overlain by an "orthoquartzite", which contains pebble-sized quartz clasts - possibly derived from the underlying metaquartzite. The breccias contain quartz clasts set in a hematite stained, medium grained arenite. Large (0.5 m) phacoidal blocks of the breccia are enveloped by thin shaly borders, the whole giving the impression of a shear zone.

This outcrop certainly inhibits any notion that the unconformity between Precambrian basement and overlying cover is a simple one!

Points to concentrate on are:

1) Possible shear fabrics in the lower part of the outcrop, in part defining the borders of coherent blocks of breccia
2) Compositional variation of clasts in the breccias
3) The relationship between and the textural character of the each of the recognized rock types

If the upper part of this outcrop is indeed comprised of rocks which are part of the Grenville basement as supposed by Karboski (1976) and Karboski et al. (1983) then the current disposition of lithologies seems not to be satisfied by a model that involves karst infill alone (Van Diver, 1976) as seems possible at Rock Island Road. It is possible, however, that the wall collapse of a large solution depression may have allowed a slab of basement to slide into the argillite filled basin. The presence of shear fabrics within the breccia may be accommodated by this model provided that these rocks were only partially lithified at the time.

Cross over to the vehicle and investigate the southern end of the outcrop on the southeast side of the road. Here, a narrow zone of sandstone breccia dips steeply through marble. One of the contacts is sheared indicating high angle faulting. Is there evidence for displacement sense?

[Note: if time permits an outcrop of marble on the nortwest side of Route 11, a few hundred yards to the south should be visited. It contains narrow subvertical veinlets of arenite at the northeast end, which provide good evidence for solution cavity infill]

Return to vehicles and continue northeastwards on Route 11.

88.9	4.9	Junction with Route 812.
89.5	0.6	Stop 8. Outcrops on both sides of road.

STOP 8: Southwest of East Dekalb.

Please note that description of this stop is repeated in Trip B2

Roadcuts due south of Dekalb further illustrate the complexities within these post-Grnville rocks. A long series of low exposures on the northwest side of the road contain well-layered marble (reclined folds at the north end) to the north which in the central section give way to a complex steep contact with rusty weathered arenaceous rocks. A number of small sandstone breccia wedges penetrate downwards into the marble in the vicinity, again supporting the solution-pocket infill model for the larger scale relationships seen elsewhere (*e.g.* Rock Island Road). Evidence for sulfide mineralization is present at some contacts, in common with the Rock Island Road locality (Elberty and Romey, 1990).

Bedding in the sandstone is irregular and breccia/conglomerates are common, particularly lower in the section a hundred yards to the south and at the base of the southernmost outcrop on the northwest side of the road. A poorly defined cylindrical structure exists in the latter section and may be seen on the top of a low outcrop of compect, rusty weathered, quartz-arenite (compare with Stop 3).

Similar conglomerate and breccia occur in a large outcrop on the southeast side of Route 11. Quartz, quartz-arenite, and metaquartzite are common clast compositions and are similar to those at Stop B3 (Stop 9 in Fig. 1). In both the western and southeastern outcrops shaly zones could have been generated by shear and at the northern end of the latter bedding dips are steep (> 60°; Karboski, 1977), yet again supporting the notion of significant post-depositional deformation .

Continue northeast on Route 11

93.15	3.65	Dekalb center
101.0	7.85	Junction with Route 68. Turn right.
101.3	0.3	Canton Town Park

END OF TRIP

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TRIP A3

CAMBRO-ORDOVICIAN STRATIGRAPHY, SEDIMENTATION, AND ICHNOBIOLOGY OF THE ST. LAWRENCE LOWLANDS - FRONTENAC ARCH TO THE CHAMPLAIN VALLEY OF NEW YORK

Note: The following four articles constitute Trip A3. The first of these is the roadlog whilst the subsequent three provide essential supporting information.

Trip A3 (1) **Cambro-Ordovician Stratigraphy, Sedimentation, and Ichnobiology of the St. Lawrence Lowlands- Frontenac Arch to the Champlain Valley of New York**
J. Mark Erickson page 68

Trip A3 (2) **Traces Fossils and Stratigraphy in the Potsdam and Theresa Formations of the St. Lawrence Lowland, New York**
J. Mark Erickson and Thomas W. Bjerstedt page 97

Trip A3 (3) **A Preliminary Evaluation of Dubiofossils from the Potsdam Sandstone**
J. Mark Erickson page121

Trip A3 (4) **Distribution of Trace Fossils preserved in high energy deposits of the Potsdam Sandstone, Champlain, New York**
J. Mark Erickson, Peter Connett, and Andrew R. Fetterman page 133

TRIP A3 (1)

CAMBRO-ORDOVICIAN STRATIGRAPHY, SEDIMENTATION, AND ICHNOBIOLOGY OF THE ST. LAWRENCE LOWLANDS - FRONTENAC ARCH TO THE CHAMPLAIN VALLEY OF NEW YORK

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INTRODUCTION

This trip is being offered for the purpose of introducing students of geology to some of the interesting and important relationships to be found in the first Cambrian(?) rocks to be "properly" described on this continent. Indeed, the name Potsdam is the oldest recognized stratigraphic name in North America accepted under the terms of the Code of Stratigraphic Nomenclature. The unit is, however, not strictly homogeneous, nor are its age and stratigraphic relationships known throughout its geographic distribution. The aim of this field trip is to point out some of the stratigraphic, sedimentologic, paleontologic, and paleoenvironmental variety that has been included under the label of "Potsdam Sandstone" in the region of the type area, northern New York.

This field trip will span the North Country from Chippewa Bay to Champlain, New York, (Figure 1) and will emphasize the basal units in the Paleozoic sedimentary cover of the basement. During its progress the guide will attempt to point out relationships between bedrock and the activities and deposits of the Wisconsinan glaciation which covered the entire field area. The trip will be entirely within the St. Lawrence Lowlands physiographic province as used by Fisher, 1977 (Figure 2).

Please note that the articles in support of this trip (Erickson & Bjerstedt; Erickson; and Erickson et al.) follow this road log.

ROADLOG

In this road log one will find, associated with each stop description, a number of questions that are best addressed by relationships at that particular stop. To stimulate thought and discussion participants are encouraged to make their own observations in an attempt to answer these and any others that arise in the process. The author assures you that he does not have answers to most of these!

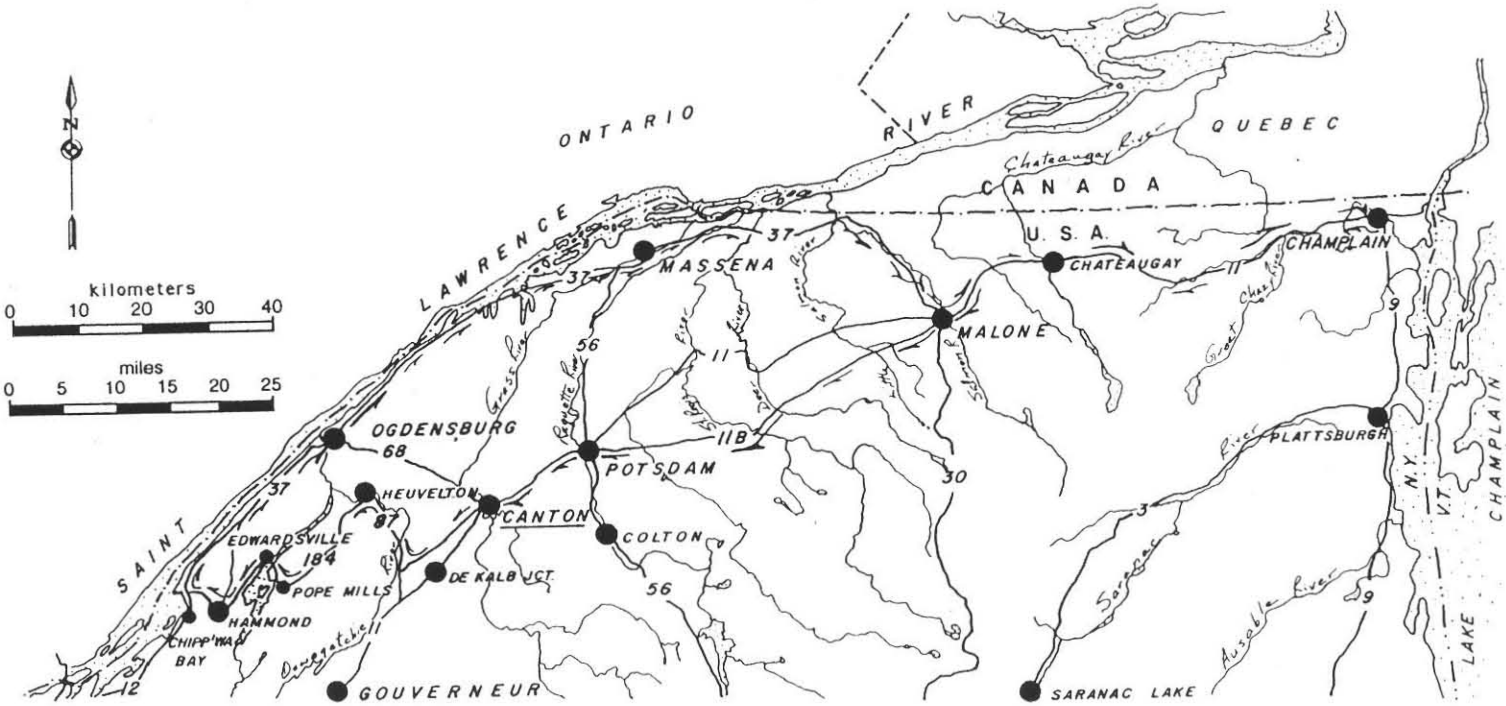

Figure 1: Map indicating route of the field trip.

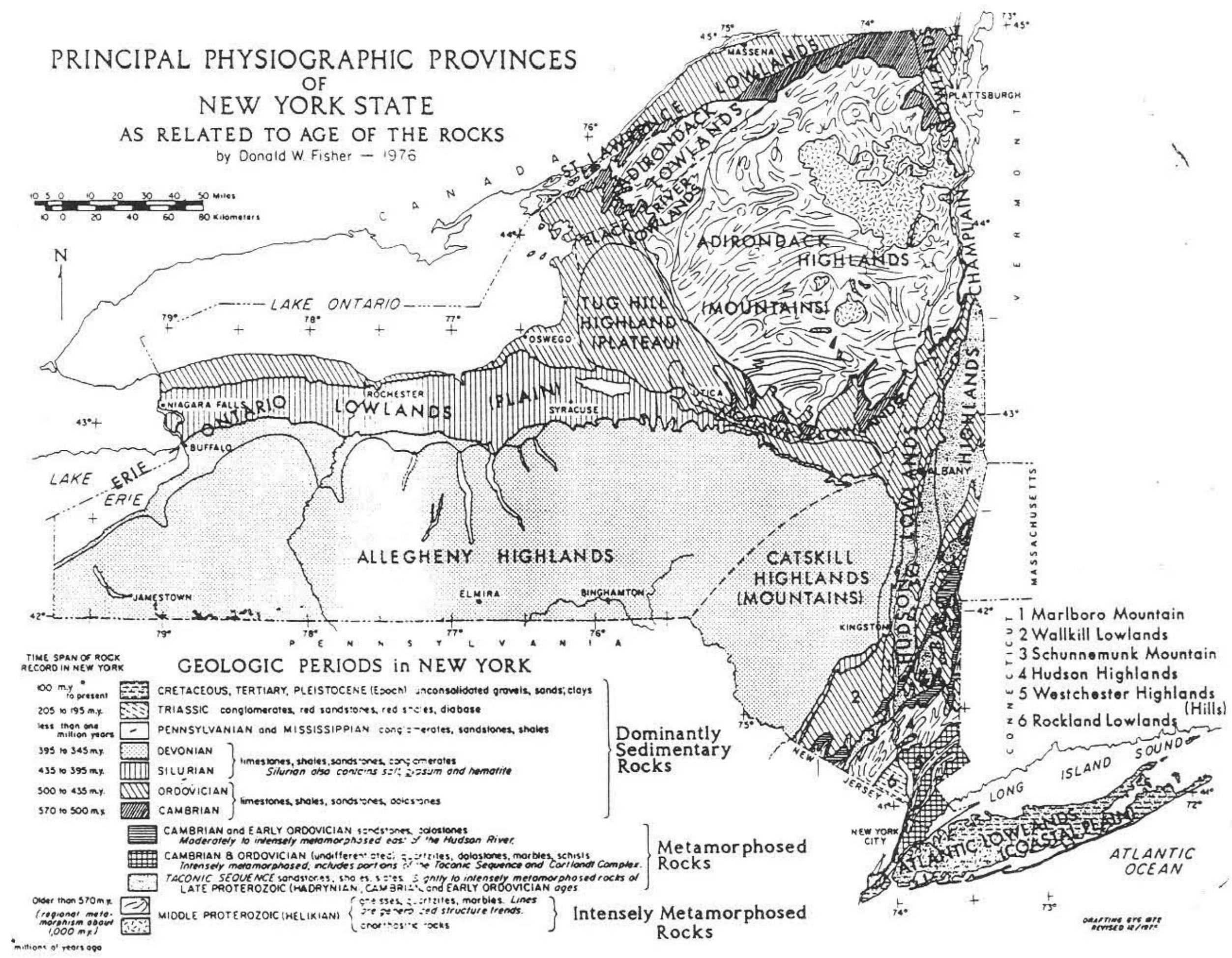

Figure 2: Physiographc map of New York from Fisher (1977) indicating extent of the St. Lawrence Lowlands physiogarphic province as used on this trip.

Questions to be raised by this field trip:

1. What is meant by "Potsdam" - how well defined is the unit? Is the name being applied properly? Is misapplication preventing better understanding of regional tectonic and depositional histories
2. What is the age of the Potsdam Sandstone?
3. What depositional environments are involved in the Potsdam Fm.?
4. What are the relationships between the Potsdam-Theresa-Ogdensburg Formations?
5. What was the biota of the Potsdam-Theresa interval?
6. What was the paleoecology of the ichnotaxa (and of the trace-making organisms) found in these rocks?
7. Is the first appearance of trace fossils in regional rocks indicative of evolutionary change or of depositional environment?
8. What are the stratigraphic relationships of the formation?

Cumulative mileage	Miles from last point	Route description
0.0	0.0	Mileage log begins at the intersection of Main and Park/Court Streets in the center of the Village of Canton, N.Y.
0.1	0.1	Proceed N. on Court Street 1-1/2 blocks to parking lot of the newly expanded St. Lawrence County Court House. Disembark and walk to original wing of the complex centered on Court Street.

STOP 1: St. LAWRENCE COUNTY COURT HOUSE. (Figure 3)

Photo stop - no hammers please!

The field trip begins appropriately at this classic structure because it contains most of the impressive elements of the Potsdam Sandstone as it occurs in the stratigraphic/lithographic type locality in the vicinity of Potsdam, N.Y., 10 miles to the East. The edifice also demonstrates many of the uses of the Potsdam and the skills of the quarry craftsman.

Emmons (1838, p. 214; 1842) named this unit in the Report of the Survey of the 2nd Geological District of New York. Quarries of the Potsdam are no longer accessible; however, the lithologic character of the "type" Potsdam is easily seen in this structure.

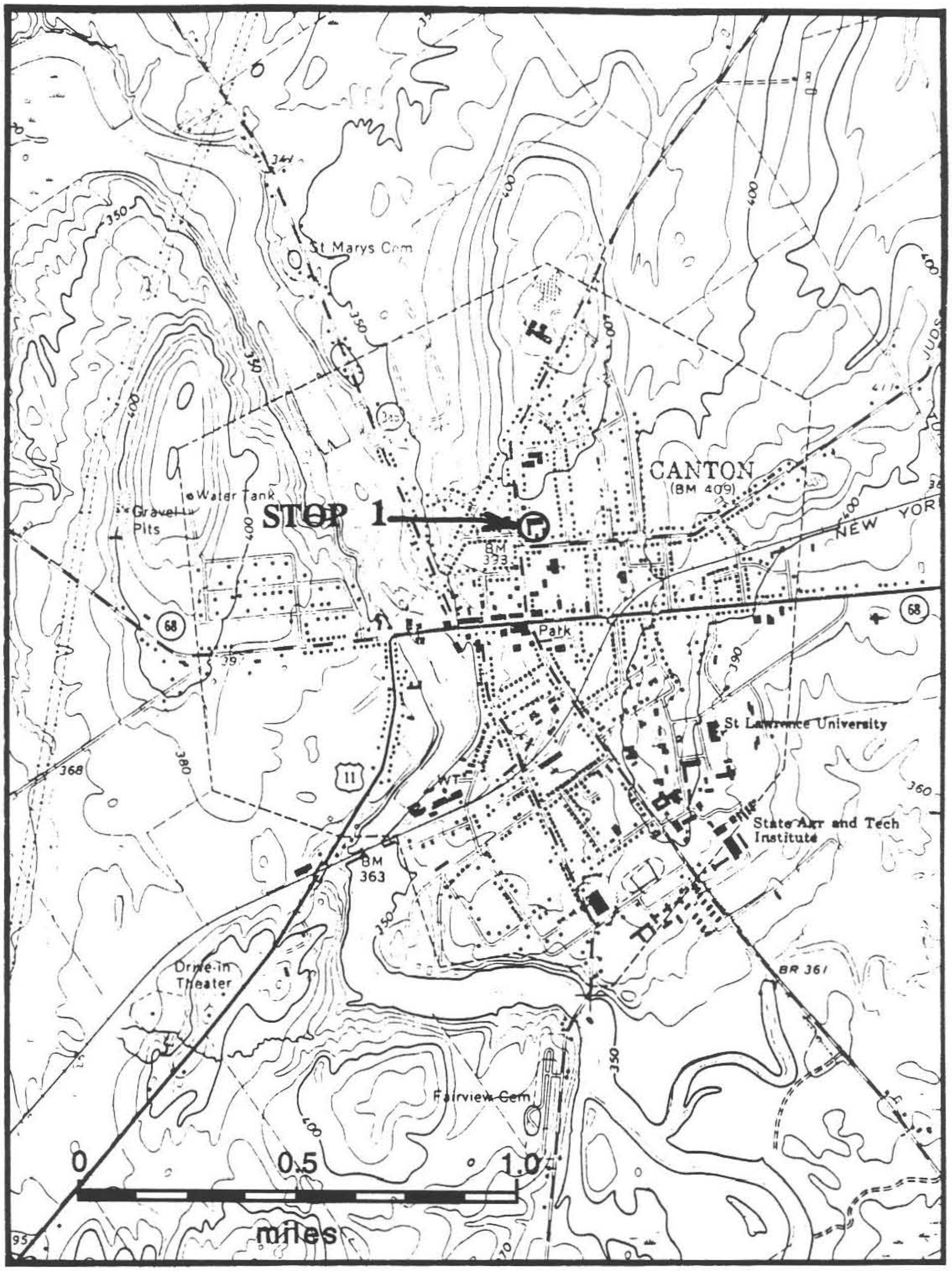

Figure 3: Topographic map illustrating location of Stop 1 on the Canton, N. Y. Quadrangle.

The rock is a fine-to-medium grained, well-rounded, well sorted, laminated, plane bedded or cross-stratified, hematitic, silica cemented orthoquartzite. It has very unique decorative properties because quartz grains were individually coated by hematite before the silica cement was introduced. For this reason the red coloration is protected from weathering, thus facades do not become streaked or stained as is the case with some hematitic or limonitic building stones having calcitic cement that contains the oxides or with pyritiferous limestones and sandstones.

The gray ashlar blocks of the courthouse are a Precambrian unit known as Gouverneur Marble. Red trim is all Potsdam Sandstone. Both were produced in St. Lawrence County. Please note that the newly completed addition to the courthouse attempts to use the same color themes but employs man-made materials rather than native building stone.

Questions to be raised at Stop 1, in particular are:

1. What are the lithologic properties of the Potsdam Sandstone?
2. What depositional environments are represented in the "type" Potsdam Sandstone?
3. Do the rocks contain fossils - either ichnofossils or body fossils?
4. What primary sedimentary structures are present?

0.2	0.1	Return to Main Street (Hwy. 11). Turn RIGHT (West) on Main Street and proceed west. Note sandstone facade on Canton Municipal Building on left. Cross Grass River, pass straight through traffic light and follow Rte. 68 toward the North for two blocks to Old Dekalb Road. The Grass River is one of several northward-flowing rivers whose headwaters are in the Adirondack Highlands. These are tributary to the St. Lawrence River approximately 20 miles to the North. Several have major power dams or paper mills on them thus being developed to the degree that the original character and fishery of the river has been lost. Canton's water supply is drawn from the Grass River.
1.0	0.8	Jct. Hwy. 68 and Old Dekalb Road just before True Value Hardware. Turn LEFT and follow Old Dekalb Road to its end at Hwy. 87. The route passes along the Canton phaccolith, a body composed of alaskite and surrounded by marble and paragneiss, all presumably Grenvillian age (Bloomer, 1965; 1967).
3.3	2.3	Outcrops of Canton Alaskite are visible at crest of hill.
6.8	3.5	The entire region has been glaciated multiple times, most recently during the Wisconsinan Stage of the Pleistocene. Topography of both bedrock and surficial material reflect that process. Outcrops

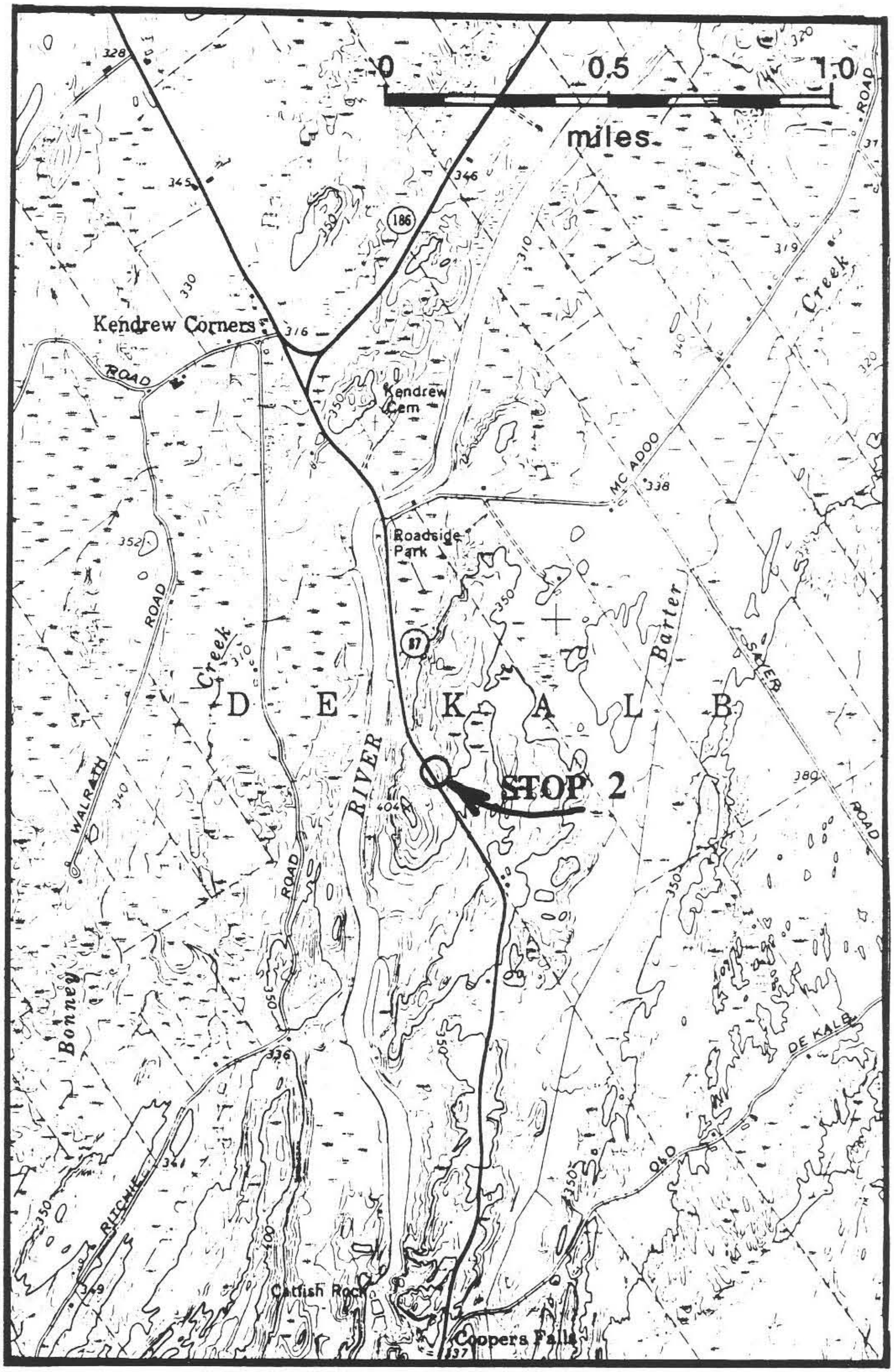

Figure 4: Topographic map illustrating location of Stop 2 on the Rensselaer Falls, N. Y., Quadrangle.

		of alaskite and gneiss having smooth glacial polish and protruding from surrounding till are referred to as "whalebacks" because of their appearance and morphology.
10.0	3.2	Differential scouring of softer marble and subsequent valley filling by alluvium and bog deposits has produced prominent valleys through the alaskite-gneiss terrain. The bus will cross such a valley at the given mileage. The next stop lies within such terrain.
10.01	0.1	Jct. with Hwy. 87. Turn RIGHT (North) and proceed northwestward along the east side of the Oswegatchie River.
11.4	1.3	Stop 2. Bus will turn onto gravelled parking strip on W. side of the highway. Disembark and carefully cross this busy highway to examine low outcrops in roadcut on east side.

STOP 2: POTSDAM OUTLIER or GRENVILLIAN QUARTZITE? (Figure 4)

In the North Country no trip in the Potsdam would be complete without visiting one of the region's outcrops of enigmatic sandstone. Cushing (1916) included this hill as an OUTLIER of Cambrian Potsdam Formation overlying Grenvillian marble.

Bloomer (1965, 1967), after structural mapping and petrographic studies of included minerals, concluded that the quartzites mapped as outliers were in structural and mineralogical equilibrium with enclosing Grenvillian rocks, implying for the sandstones a Precambrian age as well. Others (eg. Brown, 1967) have considered the quartzites to be collapse fillings of , or deposition in, a karst developed on Grenvillian marbles, or fillings of cavities created by hot, high-pressure, acidic brines conducted through the porous Potsdam (see trips by Selleck, and by Bursnall and Elberty in this volume for further discussion). As one examines this outcrop bear in mind that the type Potsdam is an outlier of quartzite on Grenvillian rocks!

Careful examination of the outcrop will reveal at least six lithologic units (including those to be seen in the continuation of the roadcut approx. 75 yards to the north). The lowest unit is white, buff-weathering, fine- to-coarse grained, laminated, trough cross-stratified quartz sandstone, 3+ feet thick dipping NE at 28 degrees.

Unit is overlain disconformably(?) by laminated, plane bedded, more friable, pink and white quartz sandstone. Unit persists through thickness of approximately 20" grading into unit 3, a more massive expression of the same lithology. A 3" set of trough cross-strata that occurs 4" below top of unit 2 may be filling a channel. Unit 2 contains laminae packaged in coursening-upward bundles one to three inches thick.

Tracing unit 2 toward to South end of the outcrop reveals at least two instances where it is brecciated and fills vertical interstices on planes of displacement(?). Flow lamination in the sand matrix parallel (?) to the plane of movement suggests deposition of the matrix by fluid but indices for the direction (stratigraphically up or down) of fluid motion are lacking. Such autobreccias constitute a 4th unit. Lest there be question regarding structural motion on these surfaces, there are well-developed, if a bit weathered, slickensides in the upper portion of the south end of the outcrop.

Walking northward will bring one to a low outcrop of siliceous autobreccia and finally to a second laminated sandstone, the exact relationship of which is unclear.

Coloration in the Potsdam Sandstone is often a point of interest, even contention. Lower surfaces of pink, hematitic(?) color bands on this outcrop sometimes transgress bedding planes and show sharp boundaries having micro-irregularities similar to those of stylolytes. Coloration seems diagenetic not primary depositional in origin in this case. Such may not always be the case for Potsdam coloration, therefore each outcrop should be examined without prejudgment in the matter of color.

Stratigraphic relationships of many sandstone outliers, including those of the type Potsdam (Reed, 1934), to the main lithosome are unknown. Most workers today discuss informally the red, pink (or peach in Canada), pink and white, and white Potsdam. The terms lower and upper Potsdam are used as well. Although relationships among these units are unclear, paleoenvironmental interpretations are likely to eventually afford means for distinction.

At this outcrop consider the following questions:

1. What depositional environments are represented by the various units?
2. What are the structural relationship of the outcrop?
3. What might the sources of coloration in the sandstones be?
4. Are trace or body fossils present?
5. What is the age of this sandstone?

Reboard bus after 20 minutes. Proceed northwestward on Hwy. 87 toward village of Heuvelton.

13.0	1.6	Cross Oswegatchie River.
13.3	0.3	Note roadcut exposure of Precambrian granite gneiss.
19.3	6.0	Village of Heuvelton. This region of St. Lawrence County has been settled by Amish groups so it is probable that we will see horse-drawn buggies or other farm equipment while driving along the south side of Black Lake. These gentle folk are superior craftspersons who contribute much to the diverse culture of this county.

19.6	0.3	Turn LEFT (WSW) on Hwy. 184 and proceed to Pope Mills & Jct. with Hwy. 58.
26.1	6.5	Note stone house in Federal architectural style on right.
29.7	3.6	Note outcrops of pink and white orthoquartzite on both sides of the road. There are no fossils in this rock and it resembles some of the laminated units seen at the last stop. Note the terrain change ahead as bus crosses into metamorphic country rock.
30.3	0.6	Grenvillian gneiss.
31.2	0.9	White, poorly bedded sandstone.
31.9	0.7	Gray gneiss with granitic dikes cutting outcrop.
33.0	1.1	Pope Mills. Note outcrop of marble in the village as bus turs toward North. Bend to the RIGHT (NW) on Hwy. 58 to Edwardsville.
35.7	2.7	Cross Black Lake at Edwardsville.
35.8	0.1	Jct. with Hwy. 6. Turn LEFT (SW) and proceed to village of Hammond along the north shore of Black Lake. The north shore of this long, narrow lake marks the edge of rather flat-lying, Paleozoic sediments.
41.8	6.0	Exposure of Paleozoic sandstone.
43.0	1.2	Blinker light at jct. with Hwy. 37 in Hammond, N.Y. Proceed straight through the intersection.
43.3	0.3	Turn RIGHT in Hammond on Co. Rd. 3 toward Oak Point. Note use of stone in the new house on the right at corner.
44.4	1.1	On left is a bedding plane exposure of white orthoquartzite having a striated, polished surface. It reveals no bed features, nor any fossils. Please notes its elevation with respect to the outcrop of Precambrian gneiss 0.1 mi. ahead and to the left side of the road.
45.8	1.4	After crossing Chippewa Creek, arrive at Jct. with Pleasant Valley Road. Turn LEFT and proceed to Jct. with Hwy. 12. While driving note the escarpment cut in Paleozoic sediments on the right in the trees and the wide valley to the left in which Precambrian granitic gneiss whalebacks stand as evidence of proximity to the Precambrian-Paleozoic contact.
49.1	3.3	Turn RIGHT and travel NE along Hwy. 12 climbing through the outcrop to be visited. It is a road cut made where Hwy. 12 crosses the escarpment just viewed. Take this opportunity to preview the outcrop which we will spend some time examining.
49.85	0.75	Bus will turn LEFT at top of hill on Blind Bay Rd. and LEFT again at "T" intersection, returning to jct. of Pleasant Valley Rd. and Hwy. 12.

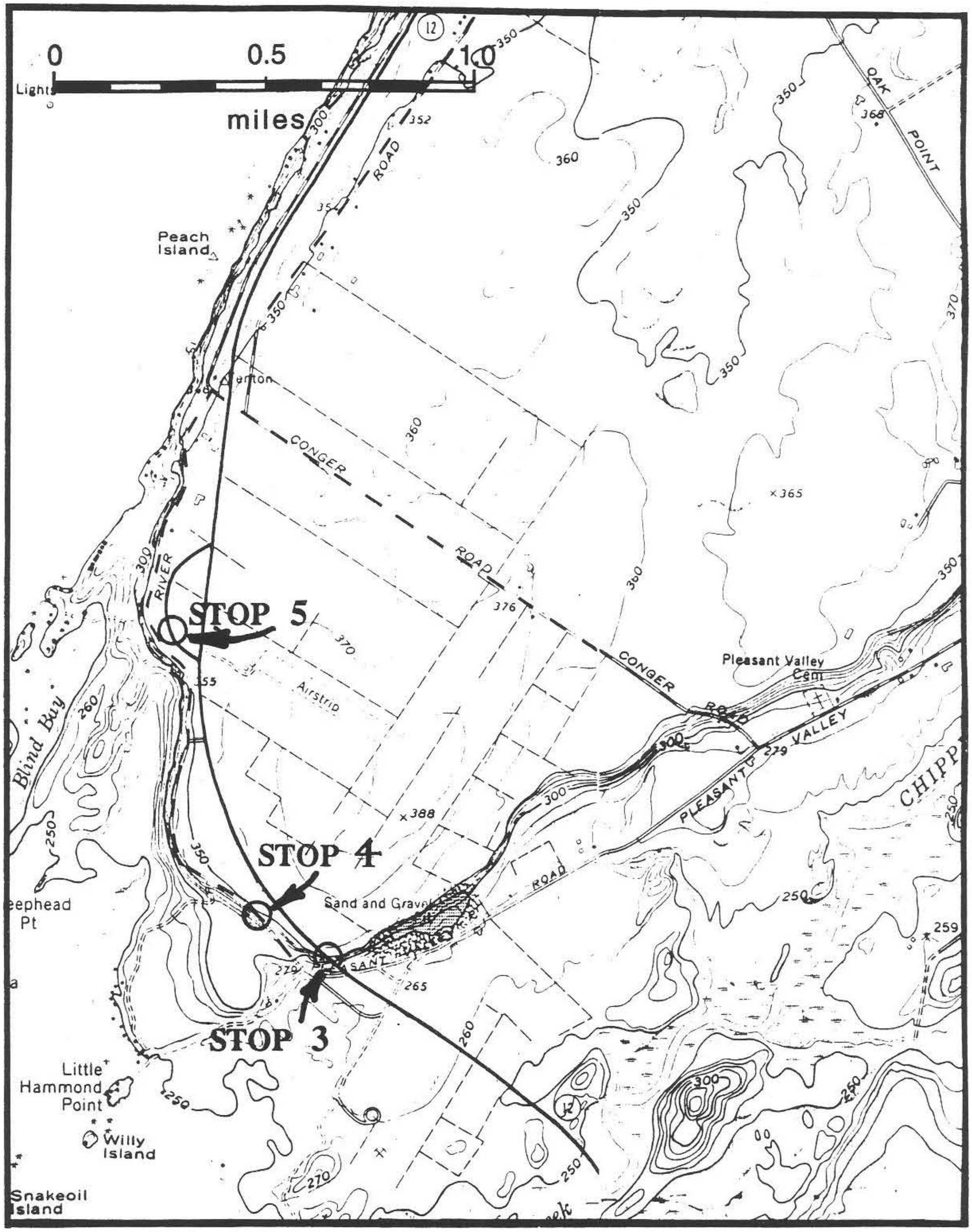

Figure 5: Topographic map illustrating location of Stops 3, 4, and 5 from portions of the Hammond and Chippewa Bay, N. Y., Quadrangles.

50.8	0.95	Turn LEFT on Hwy. 12.
51.05	0.25	Stop 3. Bus will proceed up hill to wide shoulder beyond guardrail. Discharge quickly. WATCH OUT FOR TRAFFIC.

STOP 3: "White" Potsdam and Theresa Formations are exposed in this important outcrop (Figure 5). PLEASE USE PROFESSIONAL CARE - WATCH FOR TRAFFIC AND FOR LOOSE ROCK ON THE OUTCROP.

This outcrop provides one of the most complete exposures of the Precambrian-Potsdam-Theresa transitions in the northwestern St. Lawrence Adirondack Lowlands on the American side of the St. Lawrence River. That is not to say that the section is chronostratigraphically complete; almost certainly it is not, because the more angular, poorer sorted, ferrugenous, often arkosic and siliceous-cemented sandstone facies generally associated with the older (lower) Potsdam are not present in the immediate vicinity. The red rocks exposed approximately 0.3 miles to the SE are Precambrian alaskitic gneiss. Pleasant Valley intervenes between the exposure now being examined and the Precambrian outcrop leaving the possibility that the lower Potsdam section has been removed by erosion.

Other local sections, however, preserve the transition into white Potsdam Sandstone through a few to several feet of quartz cobble conglomerate which marks a marine transgression that is probably younger than Dresbachian in this region. Its precise age is not yet well constrained (Fisher, 1977) in this location, yet lithologic relationships in northeastern New York (Fisher, 1977; 1982) and recent ichnobiostratigraphic studies by Yochelson and Fedonkin (1993) infer a Franconian or even a Trempealeauan age for rocks not containing species of the ichnogenus *Climacticnites*. No *Climacticnites* sp. have yet been recovered from the "white" Potsdam sandstones locally.

The trip first examines a relatively unweathered exposure of Potsdam-Theresa deposits created as a roadcut approximately 25 years ago. This section includes units 3 through 8 on the composite stratigraphic column given herein (Erickson and Bjerstedt). The Potsdam is white, cream or buff, brown-weathering, moderately sorted, fine to coarse, medium-to thick bedded, cross-stratified or laminated, calcite- or silica-cemented orthoquartzite. Current ripples and "herringbone" cross-strata suggest deposition under tidal conditions (see Erickson and Bjerstedt herein). Contact between Potsdam and Theresa is taken to be at the inception of dominant calcareous cement in grey, buff-weathering, more poorly sorted ferrugenous, thin- to medium-bedded, limey sandstones and sandy dolostones. Bioturbation becomes a controlling element of preservation of primary sedimentary structure at this point. Iron is often present as pyrite framboids. Originally these sediments were very rich in organics. Much of that original food resource was removed by trace-making organisms leaving rocks of the unique "Theresa aspect" in which bedding planes are formed between bioturbated sedimentation units seen in this outcrop.

Cross-strata and presence of *Skolithos* sp. suggest that the white Potsdam lithofacies originated as sandwaves built across expansive tidal flats subsequently reworked by shifting tidal currents. Varying eustatic conditions resulted in recurrence of such conditons a number of times during Theresa deposition. Thus a "white Potsdam" lithofacies of rather clean quartz arenite can be found in the Theresa repeatedly. The *Skolithos* ichnofauna is not, however, always contained in it. It seems the Potsdam lithofacies does not always carry the Potsdam Ichnofacies.

At this outcrop the two styles of ichnofaunal development - infaunal suspension feeders and infaunal deposit feeders - are recognized. Bjerstedt and Erickson (1989) and Erickson and Bjerstedt (herein) discuss these trace fossil relationships, particularly as they relate to energy of depositional environment and depth of burrowing. Here it is emphasized that body fossils are essentially absent from this unit so that knowledge of the biota is developed from the ichnofauna. The Potsdam was unable to support deposit-feeding organisms as it was too well-sorted, reflecting high energy conditions of deposition on the tidal platform. Theresa sediments, on the other hand, must have held a rich admixture of organic matter within their more muddy matrix. These are burrowed extensively to depths of more than 5 cm by a wide variety of trace-making organisms. These are illustrated in the accompanying article by Erickson and Bjerstedt (herein).

Questions to be considered at this stop:

1. What were the depositional environments of the Potsdam and Theresa Formations.
2. Why are there no body fossils in these rocks?
3. What is the significance of the depth of burrowing n the Theresa? (Does it reflect ability of the Burrowing organisms, food content of sediments, sedimentation rates, oxygen content, or other environmental factors.)
4. What types of organisms were the likely trace-makers?
5. What was the source for the organics being utilized by deposit feeding ichnofauna?

Walk 0.1 mi up section along route 12. The bus will meet us at the top of the hill after 40 minutes. Please work your way up section steadily.

51.3	0.25	Turn left at sign to Blind Bay.
51.35 .	05	Turn left at "T" intersection.
51.85	0.5	Proceed SW to low roadcut outcrop on scarp wall.

STOP 4: WEATHERED THERESA FORMATION. (Figure 5.)

In the weathered bedding plane exposures found along this roadcut and the scarp to the SE one finds a diverse ichnofauna preserved in hyporelief, or occasionally in epirelief, on bed soles and surfaces. Examples of *Monocraterion* sp., *Phycodes flabelliforme*,

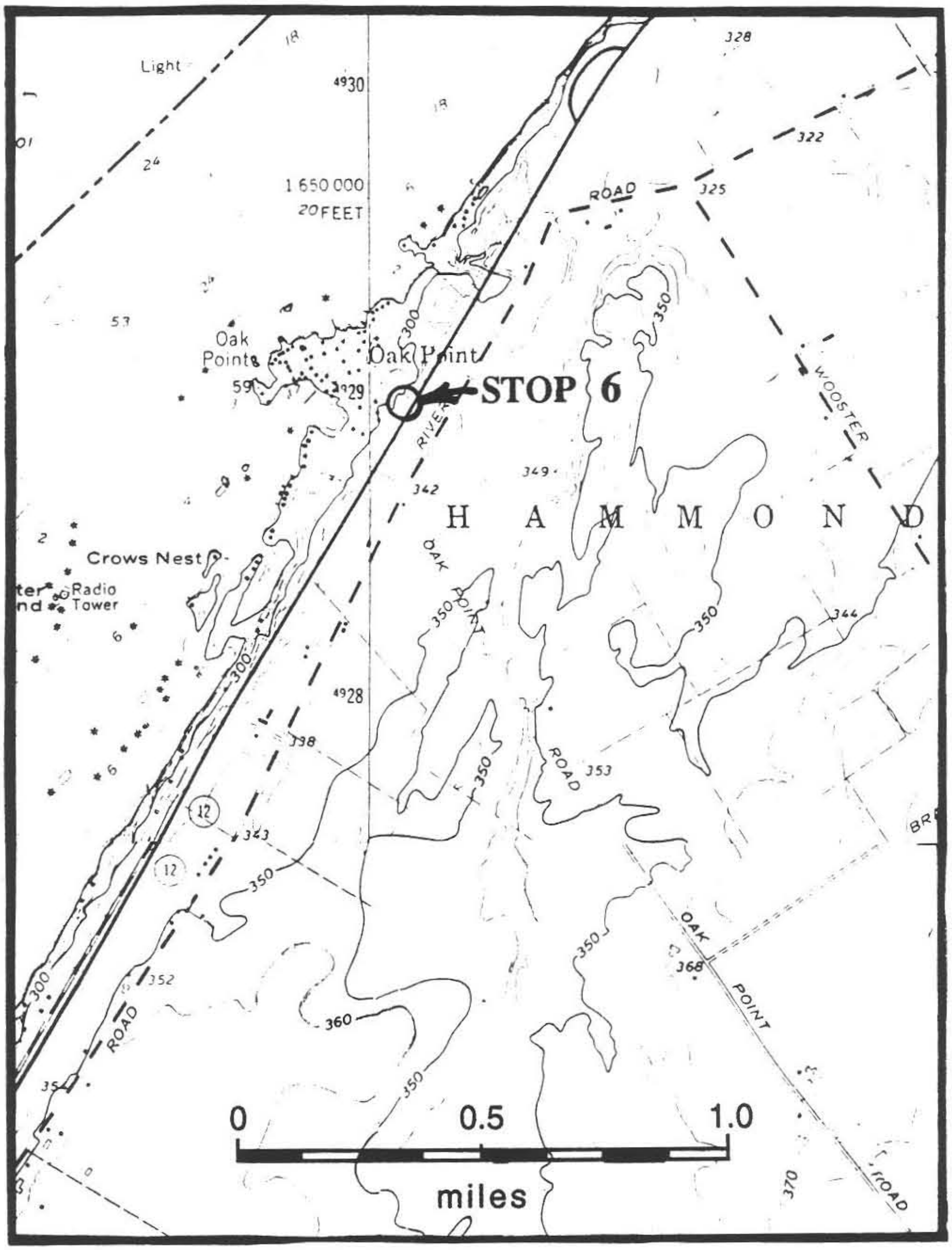

Figure 6: Topographic map illustrating location of Stop 6 from portions of the Oak Point and Hammond, N. Y., Quadrangles.

Planolites sp., *Roselia socialis* and perhaps of *Cruziana* sp., will be found here. Weathering along the scarp face represents approximately 10,000 years of development, whereas that along the roadcut may be much less. Apply the same questions examined at Stop 3.

56.35	4.5	Proceed back to Hwy 12. Turn left (NE).
57.15	0.8	Stop 5. Turn into Rest Stop overlooking Chippewa Bay/Blind Bay and the beginning of the Thousand Islands Region.

STOP 5: ST. LAWRENCE SEAWAY SENIC VIEW. (Figure 5)

Photo pause - 10 minutes. Islands in the St. Lawrence River are formed by Precambrian (Grenvillian) granitic gneiss which is elevated along the Frontenac Arch, a structural high that links the Adirondack Dome to the Canadian Shield proper. This view of the region is probably not unlike that which one would have seen during the Cambrian as the Sauk marine transgression crossed the region. Knobs of grenvillian metamorphic basement became islands about which conglomerate, and subsequently sandstone, were deposited as tidal currents swept between the islands.

The St. Lawrence River, now a "seaway" for ocean-going vessels, flows NE draining the Great Lakes of Canada and the U.S. The international boundary passes up the middle of the river here, but east of Massena, NY the river lies entirely in Canadian territory.

Note boulders of local and regional rock types.

Return to Hwy 12 turning LEFT (NE) once again. You will be traveling on a surface held up by Paleozoic calcareous sandstones of the Theresa Formation. Note the appearance of this St. Lawrence Lowlands topography; it is typical of landscapes with very thin, or no, glacial drift remaining on it..

59.65	2.5	Proceed to Riverledge Rd. Turn LEFT (west) and follow road to Oak Pt. turn.
60.10	0.45	Turn LEFT at sign to Oak Point. Drive around the loop at Oak Point which is nearly at river level. Note Precambrian gneiss. Where loop closes on itself Precambrian gneiss lies NE of road and Potsdam Ss has SW forming ledge behind garage. Contact not exposed but quartz pebble conglomerate occurs in wooded area to ENE of the side roads.
61.0	0.9	Jct. of Oak Pt. road and Riverledge roads. TURN LEFT and proceed .2 mi.
61.2	0.2	Stop beside low outcrop.

STOP 6: SECTION AT OAK POINT. (Figure 6)

The drive around the Oak Point loop has provided a typical example of many of the exposures along both Canadian and American sides of the St. Lawrence River in this immediate region. Exposures along the frontage road contain the Potsdam-Theresa contact.

This outcrop offers another opportunity to investigate the burrowing depths, bioturbation thicknesses, and trace fossil taxa involved in the transition to carbonate rich mudflat sedimentation in lower energy conditions (?) upsection. This section is noteworthy because it has produced one example of an arthropod trackway of large size probably assignable to the ichnogenus *Protichnites* as is discussed herein by Erickson and Bjerstedt. The trackway will be available for examination and discussion at this stop.

Depart Oak Point turning LEFT (NE) onto Hwy. 12 toward Ogdensburg.

61.8	0.6	Note white sandstone bed in blue-gray Theresa Formation on right of road.
64.3	2.5	Slight deformation of Paleozoic strata can occasionally be seen in roadcuts such as that near Brier Hill.
67.2	2.9	Jct. with Hwy. 37. Several exposures of Theresa Formation form roadside outcrops here.
68.4	1.2	Stone fences constructed of slabby, upper Theresa Fm. begin to be a local hallmark in this region. They will be emphasized in the "Stone Fence Motel" that will be seen in about 15 miles on the LEFT (NW) side of the road.
70.5	2.1	Scotch Bush Road. A small quarry opened in the "white Potsdam" (or Heuvelton). Sandstone lies 1.3 miles south along this road. It is now being used to store rubble, but it once presented a fine view of this unit.
75.8	5.3	Stratigraphically the trip has traversed the Theresa-Ogdensburg contact. Bedrock now is Ordovician limestone and dolostone. The highway here passes through a small (no admittance) private, abandoned quarry in the Ogdensburg Formation in which can be found domal stromatolites. The depositional setting of these rocks was described by Kerans (1977) and conodont faunas from two larger quarries in the formation were investigated by Judson (1981).
76.0	0.2	Stone Fence Motel on left.
77.3	1.3	Pass a large stone quarry (behind berm on left) in the Ogdensburg Formation (See Kirchgasser and Theokritoff, 1971). It is beyond scope of this trip.

78.1	0.8	Cross the Oswegatchie River the confluence of which with the St. Lawrence lies in the city of Ogdensburg. Note the "Golden Dome", a highschool hockey arena.
78.9	0.8	Jct. Hwy. 37 with Hwy. 68 to Canton.
80.9	2.0	Bridge to Prescott, Ontario, CANADA. Frequent views of the St. Lawrence Seaway and terraine formed on glacial Lake Iroquois sediments will be seen between Ogdensburg and Massena, NY.
90.8	9.9	Full view of Iroquois Dam, a waterlevel control structure, can be seen to the north.
95.9	5.1	Jct. Hwy. 37 and 345 (to Madrid) in the center of Village of Waddington.
97.4	1.5	Construction of the St. Lawrence Seaway raised water levels in the valley greatly. Many St. Lawrence River tributaries now have drowned valleys above their confluence. That of Sucker Brook, crossed here, and Brandy Brook are typical.
98.0	0.6	Valley of Brandy Brook.
100.7	2.7	Cole's Creek, the drowned valley now being crossed, marks a former channel of the Grass River drainage which was pirated eastward after the marine waters of the Champlain Sea were excluded from the St. Lawrence Lowland by isostatic rebound. Development of the present drainage followed.
105.0	4.3	To the LEFT (N) view a coffer dam constructed across another abandoned channel of the Grass River. This dam is designed to keep the valley from being flooded by waters of manmade lake St. Lawrence.
110.3	5.3	Jct. Hwy. 37 and 56. Continue on route 37 toward Malone.
111.3	1.0	Village of Massena. There are several items of geological interest associated with Massena. It was the site of a major rapids in the St. Lawrence River which became the site for the Moses-Saunders power dam of the NY Power Authority and Ontario Hydro. Massena was also the epicenter of the 1944 magnitude 4.0 (Richter) earthquake of regional significance.The Massena Clay, a blue, marine clay deposited in the Champlain Sea, underlies much of the region. This deposit contains the marine bivalve mollusks *Macoma baltica* and *Hiatella arctica*, a modest foraminiferid (Katz, 1981) and ostracod (Erickson et al., 1984) fauna, and to the south at Norwood, N.Y., the skeleton of a Beluga whale was recently recovered by Bill Kirchgasser (Kirchgasser and Steadman, 993).
115.0	3.7	Jct. with road to Robert Moses Park, (left) and St. Lawrence Mall (right).

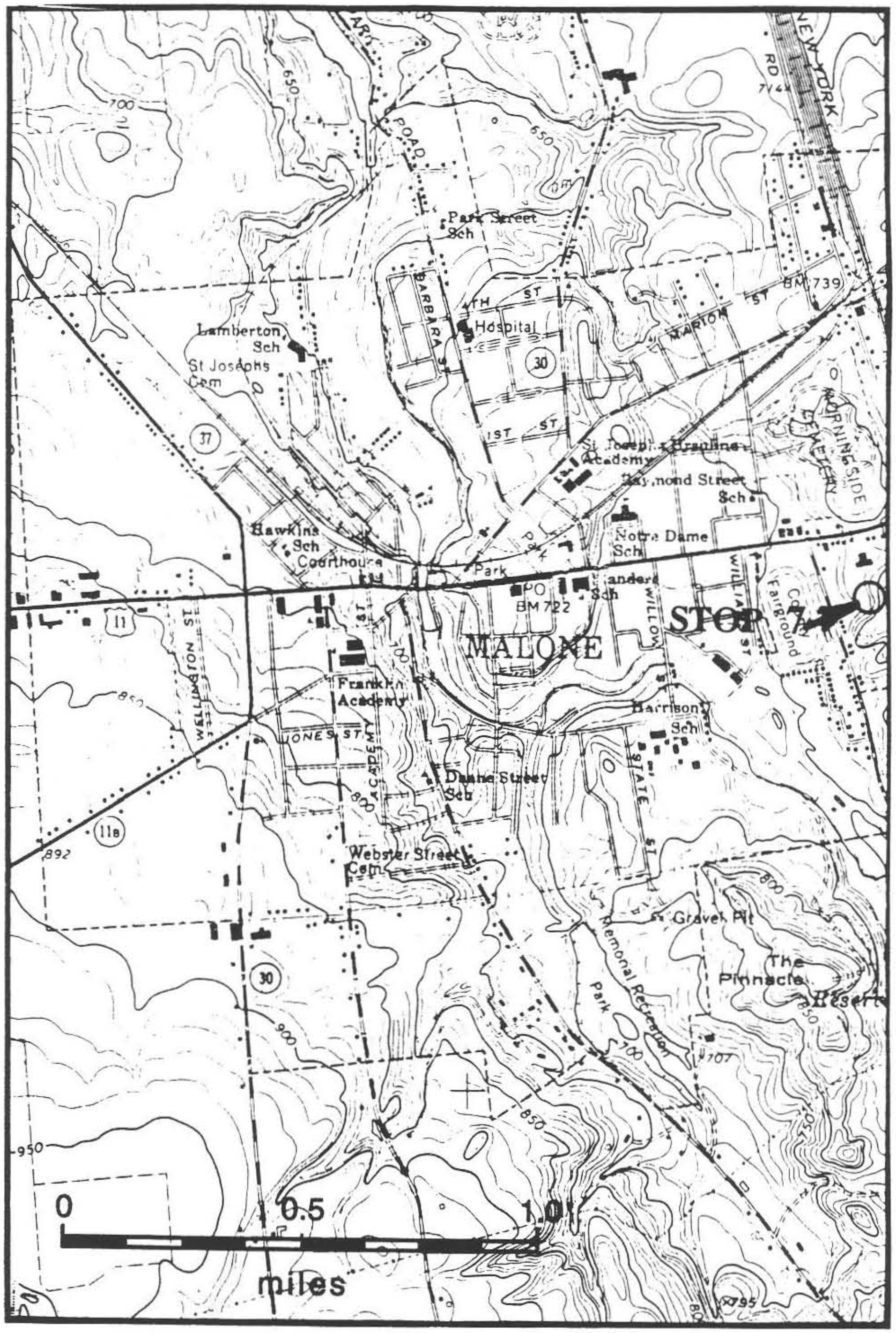

Figure 7: Topographic map illustrating location of Stop 7 on the Malone, N. Y., Quadrangle.

		To the left is the visitor center of Eisenhower Lock, one of the sites where zebra mussels (*Dreissena polymorpha*) were first introduced to U.S. waters. Robinson Bay, northeast of the lock, will be the site of the $14 million St. Lawrence Aquarium and Ecological Center.
		Inexpensive electric power from the NYPA project has made this the site of ALCOA, Reynolds, and GM plants producing and using aluminum. Some serious environmental problems have also been generated: the large earth moving project on the left is part of a PCB mitigation effort.
119.9	4.9	Biomass Poplar Plantation, are experimental, quick-growing strain of the poplar to be used for wood chips which fire a heating plant. This project was spawned by the energy crisis of the 1970's when some institutions like Clarkson University installed wood-fired heating plants.
120.3	0.4	View of the Raquette River to the right.
121.0	1.0	Cross Raquette River.
121.4	0.4	Enter Akwasasne the Mohawk Nation.7
123.6	2.2	Cross St. Regis River.
131.8	8.2	Cross Little Salmon River.
132.3	0.5	Village of Ft. Covington at confluence of Little Salmon and Salmon Rivers.
136.8	4.5	Westville. The Salmon River here flows on bedrock.
138.7	1.9	Cross Salmon River.
142.6	3.9	Terrane begins to "roll" at the approaches to the town of Malone and begins to climb onto the platform of a major delta built into the Champlain Sea (Clark and Karrow, 1984).
148.1	5.5	Jct. N.Y. route 37 with routes 11, 11B & 30 at stoplight, western edge of Malone, N.Y. Turn LEFT and proceed E. on route 11 through downtown Malone, crossing the Salmon River, and begin to ascend hill on E. side of town.
149.4	1.3	Turn RIGHT on Hillside Dirve (before "Parts Plus" auto parts store), proceed to top of hill and make RIGHT turn on Strand Drive.
149.6	0.2	Disembark and carefully cross to driveway of gray house on the corner. This is our stop. Cameras; **no hammers**.

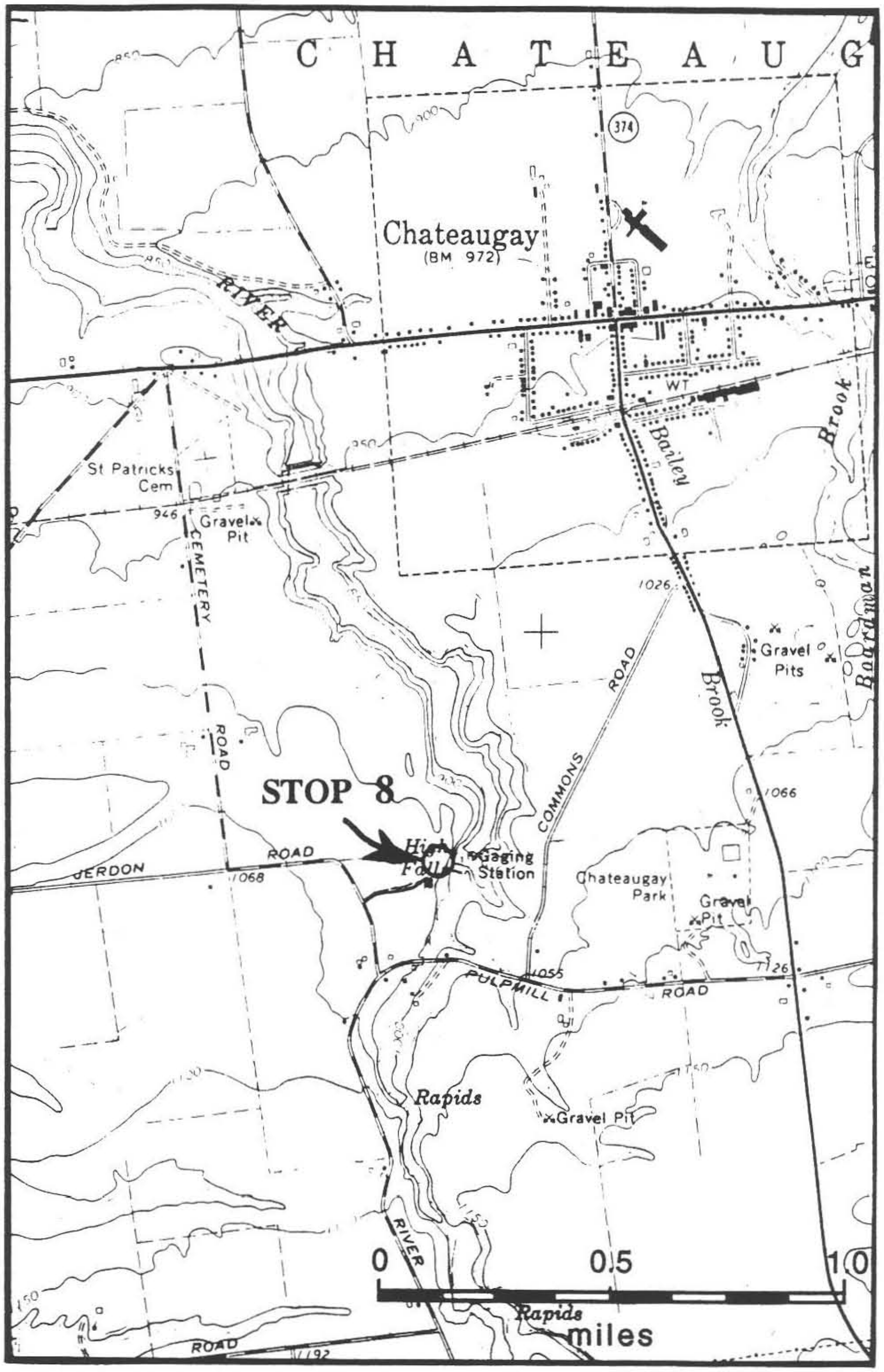

Figure 8: Topographic map illustrating location of Stop 8 on the Chateaugay, N. Y., Quadrangle.

STOP 7: MEAGAN RESIDENCE. (Figure 7)

Field trip members are the guests of the Michael Meagan Family. Walk up driveway and enter backyard through gate in fence. **Please stay on path**. Mr. Michael Meagan is the owner of the unusual Malone dubiofossils which we are here to view. The specimens, which are displayed near the pool, were discovered during excavation in 1992, and were recognized by Mr. Meagan to be unusual. He permitted removal of the specimens to St. Lawrence University where they were studied by Erickson.

Preliminary findings presented herein interpret these structures as slighly-transported fragments of algal mat with included trapped sediments. Please examine the specimens which are part and counterpart of a single block of laminated Potsdam Sandstone. During examination one may wish to keep the following questions (and others!) in mind.

1. What conditions were required to produce such an algal mat?
2. How much transport might they withstand?
3. Were there no animals present in the environment of deposition?
4. Could these be something other than algal/sediment features?
5. What is the age of this specimen?
6. Similar structures seem to be restricted to the Cambrian and very late Precambrian - why?

149.8	0.2	Retrace route to Hwy. 11 East. Turn RIGHT (eastward). You will climb again onto the plain of the Malone delta built into glacial Lake Iroquois (Clark and Karrow, 1984). From this vantage there is a commanding view of the St. Lawrence Valley to the North - the former basin of Lake Iroquois and later the Champlain Sea.
156.8	7.0	Entering Burke Center. The careful viewer will see to the north-northeast a prominent elevation across the St. Lawrence Valley. This may be Covey Hill, a Potsdam Sandstone ridge against which glacial ice lay to dam glacial Lake Iroquois. The lake drained between approximately 12,000 and 11,500 B.P. (Pair and Rodrigues, 1993).
157.6	0.8	Again, cross an abandoned drainage through a deltaic platform related to the ancestral Chateaugay(?) River.
163.6	6.0	Make a Right turn (Southward) on County Road 23, then bear left and follow signs to High Falls Park Campground, a distance of 1.6 miles.
165.2	1.6	Park at designated space and await trip leader to acquire entry to private property through a group rate fee. Pass through campstore and briskly walk to falls along designated trail.

NO HAMMERS PLEASE.

STOP 8: HIGH FALLS PARK. (Figure 8)

This scenic stop affords the opportunity to view a more easterly exposure of the Potsdam Sandstone. Comparison with the sections seen previously at Pleasant Valley Road and Oak Point in St. Lawrence County will support the idea that the sandstone of the Potsdam thicken northeastward across the Laurentian Platform. Here more than 100' of section is exposed. As thickening occurs it is probable that age relationships within the formation become increasingly complicated. Trace fossils have assisted stratigraphers to make regional correlations and paleoenvironmental intrepretations in the absence of body fossils.

At this stop one may wish to examine character of the Potsdam lithologies, bedding characteristics and contacts, weathering profile, and trace fossil content of the rocks. What paleoenvironments are suggested? Time allowed 1/2 hour. Please return to bus promptly when requested.

166.8	1.6	Retrace route to Hwy. 11. Turn Right (eastward).
166.9	0.1	Cross Chateaugay River.
167.2	0.3	Intersection of Hwy. 11 and Depot Street at stop light, Village of Chateaugay, N.Y.
167.9	0.7	Rising out of ancestral Chateaugay River Valley, note retaining wall constructed of local sandstone. On the upland there is another view northward toward the St. Lawrence Valley and Canada. South of the highway lies a newly constructed substance abuse center, one of several penal facilities that have proliferated northward in the past decade.
169.9	2.0	Enter Clinton County, N.Y.
178.8	8.9	Ellenburg, N.Y.
181.9	3.1	Ellenburg Depot, N.Y.
182.8	0.9	The trip has crossed into the basin of the Great Chazy River, the North Branch of which is crossed here.
185.3	2.5	Again the trip crosses a deltaic sand plain, probably an element of the ancestral Chazy River system.
191.3	6.0	Cross Great Chazy River in Mooers Forks.
194.3	3.0	Jct. with Hwy. 22 at blinker light, Mooers, N.Y. Turn Left. Route winds eastward once gain at edge of the village where the modern floodplain of the Great Chazy River can be seen well developed south of the highway.

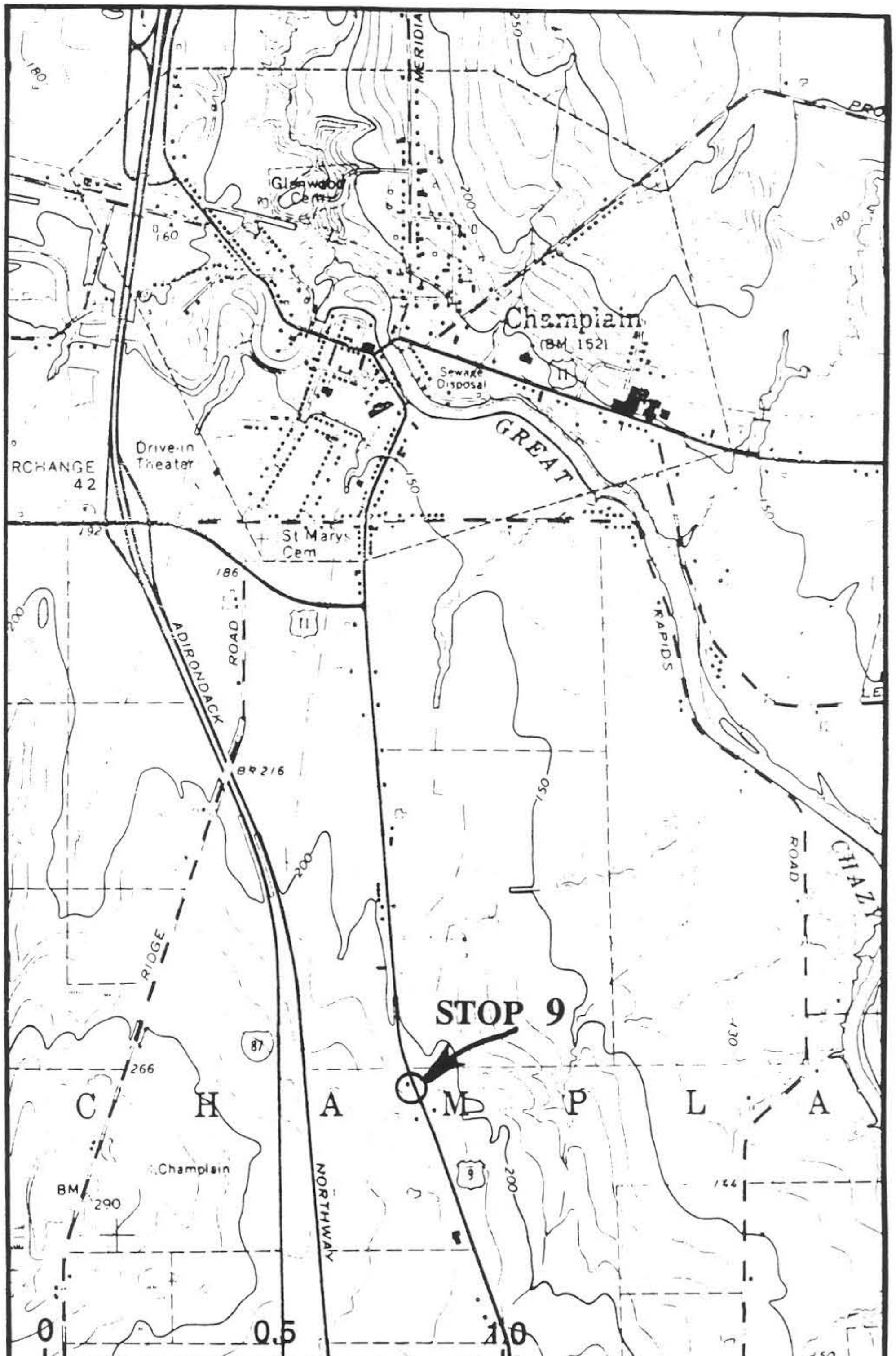

Figure 9: Topographic map illustrating location of Stop 9 from the Champlain, N. Y., Quadrangle.

197.7	3.9	Cross Great Chazy River once more on East edge of town of Champlain.
200.7	2.3	Cross I-87 ("Northway").
200.7	0.7	At stoplight, jct. with N.Y. Route 9. Turn Right (South) and travel on surface held up by Potsdam Sandstone which underlies glacial deposits.
201.8	1.1	Stop 9. Parking lot of Clinton Farm Supply.

STOP 9: CLINTON FARM SUPPLY. (Figure 9)

Exposed in the north half of the parking lot and surrounding an abandoned quarry to the rear of the Clinton Farm Supply are bedding planes of the Potsdam Sandstone. Rocks here are assumed to be in the upper or Keeseville Member. These are medium bedded, cream to white, plane and cross-stratified, quartz arenites (Fisher, 1968).

This outcrop affords an exceptional opportunity to examine bedding plane features, including large scale ripple marks, trough cross strata, sand waves and associated ichnofossils. The unusual and abundant trace fossils are of particular interest. Some aspects of these rocks are very similar to the white sandstone facies seen at Pleasant Valley Road (Stop 2) and within the Theresa Formation at Oak Point (Stop 3) this morning. At first look the traces fossils seem unique, however.

Some of the paleoenvironmental and ichnological features of this outcrop are described in the accompanying paper (Erickson, Connett, and Fetterman, this volume) resulting from studies by the 1993 St. Lawrence Paleoecology Class. Relationships here give insights into some questions raised at previous stops as well as spawning new questions such as:

1. What organism produced the large traces?
2. What controlled their distribution?
3. How deeply did they burrow?
4. What is the relationship, if any, of burrows to the ripples in both space and time?
5. How did the trace makers feed, on what, and where? (e.g., were they filter feeders, suspension feeders, or deposit feeders?)
6. Is this Potsdam Sandstone? What age is it?

Examine the outcrop thoroughly. Several bedding planes are preserved but the lowest surface, that displaying the ripples, is the most interesting. It continues on the south side of the road where some brittle-fracture patterns are present as well.

202.9	1.1	Jct. with Hwy 11. Turn LEFT (Westward) following Route 11 to Malone via Chateaugay. Proceed to west side of the village.
250.0	47.1	Jct. of Hwys. 11, 11B, 30 and 37 at stop light on the west side of village of Malone. Turn **LEFT** and proceed southward to next light - one long block.
250.3	0.3	Turn RIGHT (westward) at the light and follow Hwy. 11B to Potsdam.
250.4	0.1	Climb again onto rich soils of the Malone paleo-delta. Once a major potato-growing area, the presence of Popeye on the sign denotes that spinach is now the major crop. It is frozen on sight for sale to several brands. Rotation of fields assures nearly steady production during the growing season.
255.0	5.0	Bangor, N.Y., on the East Branch of the Little Salmon River.
257.3	2.3	W. Bangor
260.9	3.6	East Dickinson
270.9	10.0	Enter St. Lawrence County
273.9	3.0	Nicholville, NY on the St. Regis River. The Nicholville Conglomerate Member (Postel, et al., 1959) of the Potsdam Sandstone crops out along the river near here. It is overlain by orthoquartzite which occurs along Hopkinton Brook, .75 mi. N of Village of Hopkinton. Stratigraphic relationship between members are uncertain and distribution is local. As such, the Nicholville is one of three or four difficult-to-define, restricted, basal conglomeratic units that may be "Potsdam," but are more likely Precambrian. The Allens Falls Fanglomerate (Harris, 1988) is another such unit. Their examination is beyond the scope of this trip.
275.9	2.0	Hopkinton, N.Y.
284.1	8.2	Cross West Branch of St. Regis River.
289.6	5.5	Potsdam, N.Y.
289.7	0.1	Jct. with Hwy. 11. Turn LEFT at second stoplight. Proceed one block. Note use of true Potsdam Sandstone in buildings of Clarkson University and the village museum and library. This rock came from quarries south of the village that are no longer available for study.
289.8	0.1	Turn RIGHT and proceed "straight" through next light.
290.2	0.4	Cross the Raquette River. Note church of Potsdam Sandstone. on left when crossing island. Follow route 11 to Canton.
301.9	1.0	Intersection of Main and Park/Court Streets.

END OF TRIP. HOPE YOU HAVE ENJOYED IT.

ACKNOWLEDGMENTS

Numerous individuals and organizations have rendered assistance that has made this trip possible. If we should omit their mention here, it is due to our oversight rather than their lack of support.

Stratigraphy classes at St. Lawrence in the 1970's and early 1980's contributed much to my interest in the Cambro-Ordovician rocks of the St. Lawrence Valley. Marguerite Walsh, Catherine Goodmen,
Neil Sammis, Mark Klett, Charles Kerans and Michelle Judson made studies of particular note.

Dr. Thomas W. Bjerstedt was responsible for much of the descriptive work presented in conjunction with the second and third stops of this trip.

Mr. Michael Meagan alerted us to the Malone dubiofossil and expedited its removal for study at St. Lawrence. His efforts are commendable. Mr. Rick Scott of the St. Lawrence University Physical Plant and Witherbee and Whalen effected transport of the specimens to SLU and their return to Malone. Dr. H. Hoffmann gave the benefit of his experience with Cambrian dubiofossils.

Mr. and Mrs. Richard Laurin of Clinton Farm Supply graciously welcomed St. Lawrence students on several occasions during study of their outcrop. Dr. Richard Lindeman called our attention to the CFS outcrop originally.

Dr. Ellis Yochelson has discussed numerous points of stratigraphy and paleobiology with Erickson in the field and in laboratory on several occasions. The result has been an improved field trip.

Dr. Michael R. Owen and his students contributed to studies of the Potsdam Sandstone and Dr. James S. Street and his students have greatly contributed to studies of the history and process of deglaciation in the St. Lawrence-northern Adirondack region.

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TRIP A3 (2)

TRACES FOSSILS AND STRATIGRAPHY IN THE POTSDAM AND THERESA FORMATIONS OF THE ST. LAWRENCE LOWLAND, NEW YORK

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INTRODUCTION

Our field examination will focus primarily upon two composite exposures of clastic sediments that make up the Potsdam and Theresa Formations as they are recognized in the St. Lawrence Lowland (Figures 1 and 2). Other outcrops will present some interesting features of the Potsdam Sandstone.

Stratigraphy in the region is occasionally problematical as a result of discontinuous outcrop, Paleozoic deposition upon an undulating Precambrian surface, and possible post-depositional structural deformation in some portions of the outcrop belt. We will concentrate on two sections lying at, or close to, the Precambrian/Paleozoic unconformity on the east flank of the Frontenac Axis (Figure 1). The sections pass upward through the siliceous Potsdam Sandstone into muddy, carbonate-cemented sandstones of the Theresa Formation. It is in these transitions between lithologies, and the depositional environments which they imply, that we shall find distinctions in bioturbation style and ichnofaunal content. Depositional environments have been summarized by Selleck (1978, 1984) and will not be the focus of this trip.

Reasons for becoming familiar with the trace fossils and lithologic characteristics of the Potsdam - Theresa interval are more than academic. They have much to do with making continued progress in our understanding of the stratigraphic history of this margin of the Canadian Shield during the Precambrian - Paleozoic transition. The Potsdam Formation, described by Emmons (1838), is the oldest recognized stratigraphic name in the American geologic literature, yet the rocks in the type area contain facies that are discontinuous, are

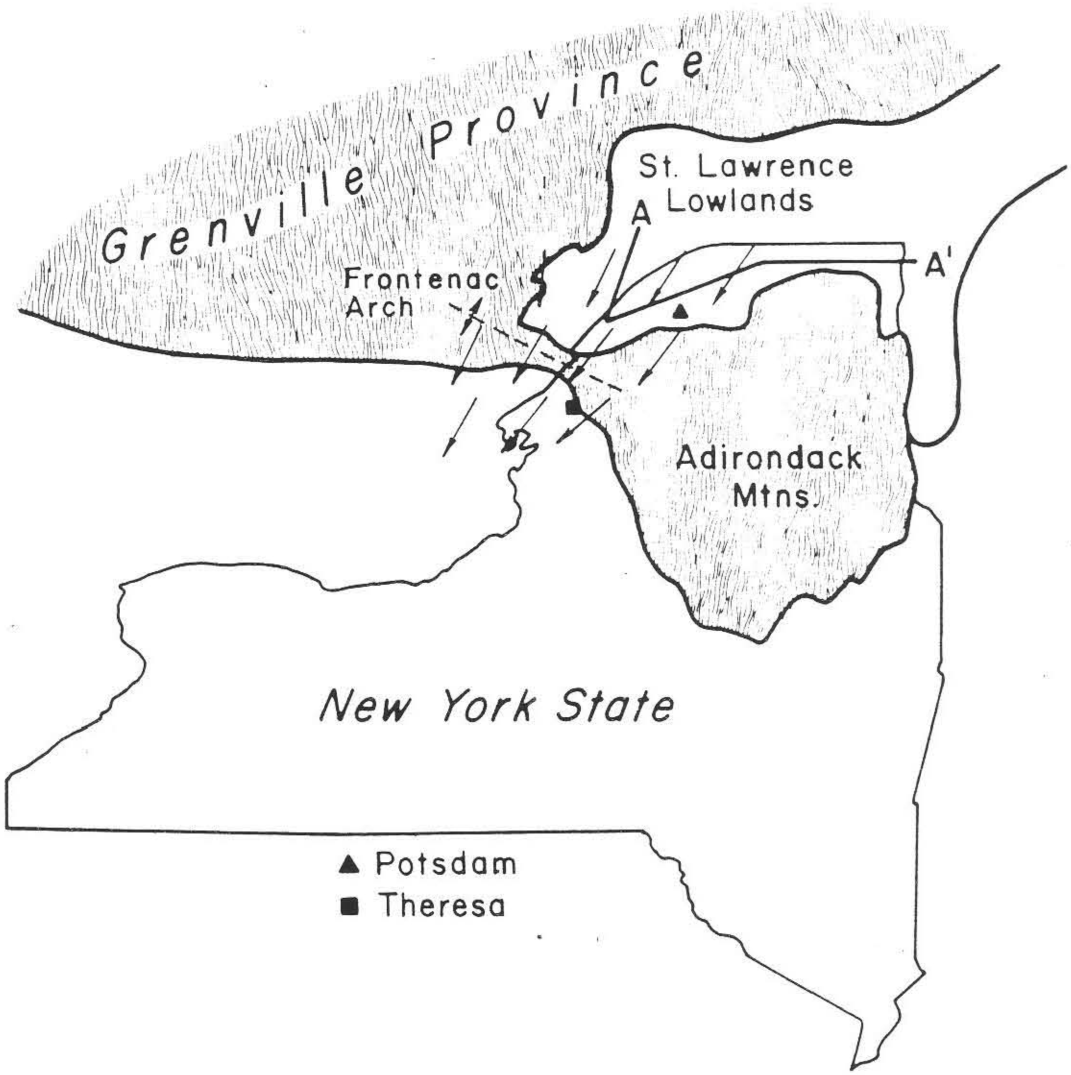

Figure 1: Regional map of the Adirondacks showing the St. Lawrence Lowlands and adjacent Precambrian Grenville terrane. Arrows show summarized vector means of moving averages for paleocurrent directions in the Thousand Island area after Lewis (1963).

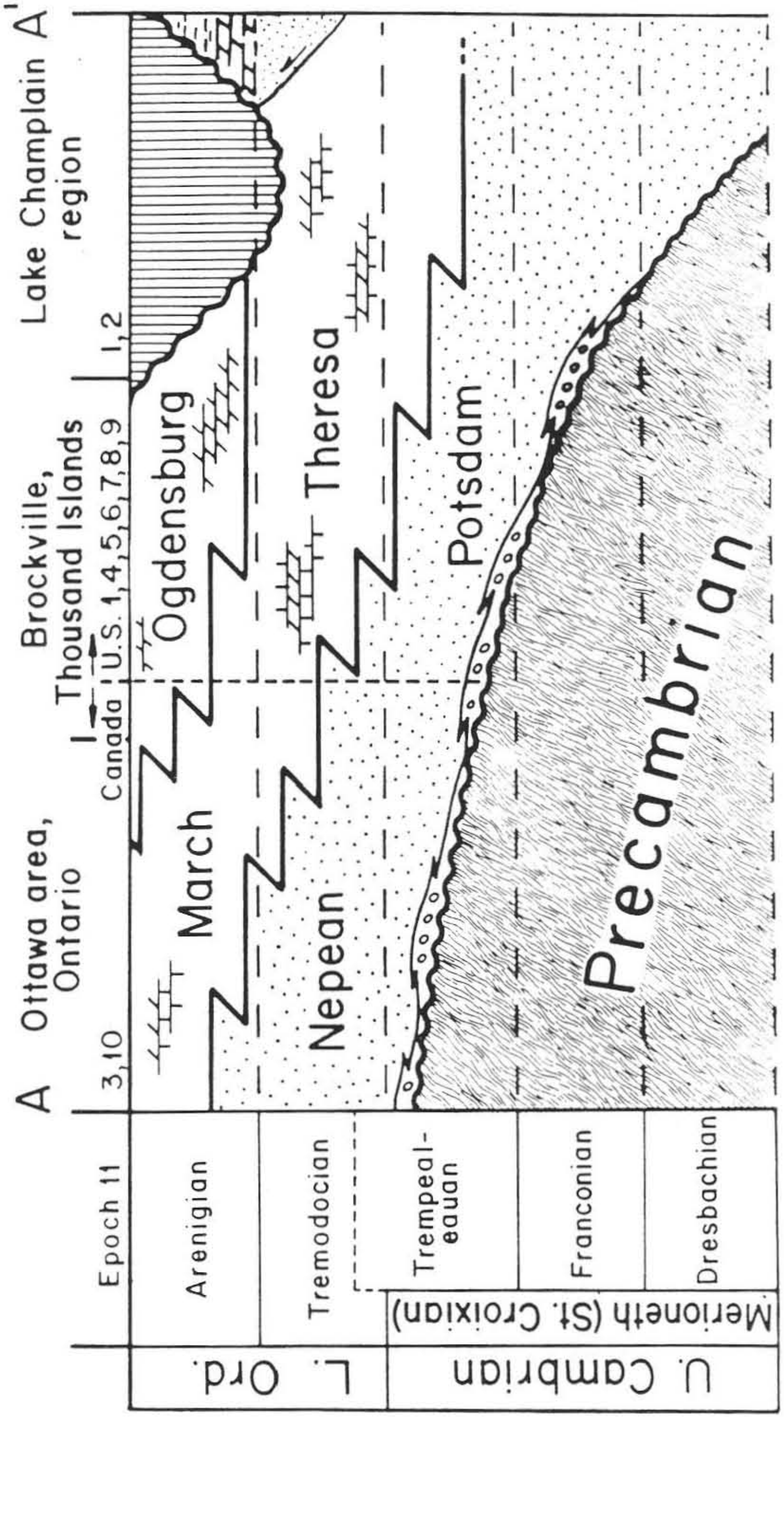

Figure 2: Lithostratigraphic and chronostratigraphic relationships for Cambrian-Ordovician rocks in the northern Adirondack region along line A-A' in Figure 1.

terrestrial in depositional origin (Chadwick, 1920), and may be Precambrian in age. They are notable for the absence of both body fossils and biogenic structures.

The Potsdam - Theresa interval in St. Lawrence County contains an assortment of facies and a number of formally named subdivisions, as well as a substantial list of locally-used names that make interpretation of the historical literature historical challenging. Readers are referred to Kirchgasser and Theokritoff (1971), Selleck (1984), and Kerans (1977) for discussions of the regional stratigraphic detail in the United States and to Greggs and Bond, (1971) for insights into the Canadian terminology.

Appearance of the first fossils, whether traces or shelled invertebrates, in this record is noteworthy as they offer the potential for correlation both locally and regionally, however tenuous those correlations may be. Fossils are reported from many outcrops in restricted numbers, and varying qualities of preservation. They only offer tantalizing pieces of biostratigraphic data for a puzzle that is complicated by presence of one, or more, disconformities among this suite of shallow water facies. Inarticulate brachiopods (*Lingulepis* sp.), flat-coiled gastropods, stromatolites, scarce conodonts, and the dendroid graptolite *Dictyonema potsdamense* which is known from only one locality, are the poor faunal elements we have to work with. Therefore, trace fossils have begun to attract more attention. Studying the appearance and development of the regional ichnofauna may add to our ability to recognize stratigraphic relationships.

POTSDAM SANDSTONE ALONG ROUTE 12

The Potsdam Sandstone in the region of Chippewa Bay and Oak Point (Figure 3) is a white, clean, medium- to poorly-sorted, fine- to coarse- grained, occasionally conglomeratic, cross-stratified or laminated, medium-bedded sandstone, generally having siliceous cement. Dolomitic cements occur upwards in some sections. Effects of high-energy depositional regimes are evidenced by presence of storm-generated conglomerate units interstratified with fine-grained sandstones. Intertidal lithotopes are suggested by herringbone cross-strata, current ripples and by mudcracks in interlayered Theresa facies. I reiterate that these Potsdam facies are significantly different from the "type Potsdam" as seen in Potsdam, N.Y. They are the Heuvelton Sandstone of Cushing (1916) or the "white Potsdam" of Chadwick (1920) which thicken southeastward from Chippewa Bay. They are at least in part Early Ordovician in age and locally there may be no Cambrian present.

The absence of Cambrian rocks is a "stretch" idea. It is suggested by the absence of the trace fossil *Climacticnites wilsoni* from the rocks here along the central St. Lawrence Lowland east of the Frontenac Axis. Yochelson and Fedonkin (1993) made a thorough study of this unusual trackway that occurs on tidal flat sandstone beds in Dresbachian formations from Missouri to the Champlain Valley. It is not an easy trace to miss and has been described from Perth, Ontario to the north and from Wellsley Island west of the Axis, yet it has not been recorded from a crescentic region flanking the Adirondacks from Chippewa Bay

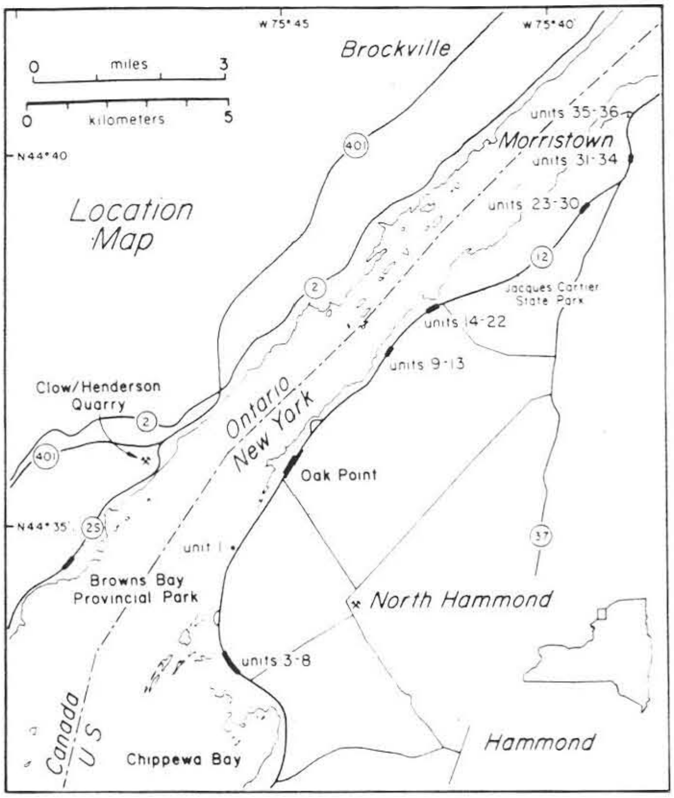

Figure 3: Detailed location map showing outcrops (by unit number) used to construct the composite section of the Potsdam and Theresa Formations along New York State Rt. 12 (Fig. 9). A Canadian reference section for the Nepean and March Formations in Ontario is along Rt. 2S in Browns Bay Provincial Park, and the Clow/Henderson quarry for the lower March Formation.

the Potsdam. Interbedded with this typical Theresa lithofacies are recurrent thin and medium beds of clean, white, cross-stratified siliceous sandstone recognizable as "Potsdam" lithofacies. The entire Theresa becomes thinner and undergoes a facies change to dominantly cross-stratified clean sandstone southeastward against the Adirondack Highlands (Selleck, 1984).

The Theresa thus contains mixed carbonate clastic lithologies in a shoaling-upward sequence. Conglomerates of Precambrian quartzite in the Theresa show that Precambrian topography continued to influence Theresa deposition (Selleck, 1984). The contact between the Potsdam and Theresa is well exposed at Chippewa Bay and Oak Point where grey, medium- to thick-bedded, calcareous and dolomitic, fine-grained sandstone overlies the Potsdam.

The lower 10 m of the Theresa is extremely bioturbated, and individual depositional events can be seen in decimeter-thick, scour-based, horizontally-laminated sandstones with burrowed tops. Most Theresa trace fossils figured herein were collected from these lower beds. Highly bioturbated lithologies in the lower Theresa represent protected, subtidal facies. Selleck (1984) noted that this facies is not everywhere preserved and occurs in areas of low Precambrian relief, indicating this facies filled topographic lows.

THERESA FORMATION

The Theresa Formation is a gray to bluish-gray, brown weathering, fine- to medium-grained, moderately-sorted thin- to thick-bedded, quartz sandstone often having calcitic or dolomitic cement and showing a greater amount of matrix residue after acid treatment than would the Potsdam. Interbedded with this typical Theresa lithofacies are recurrent thin and medium beds of clean, white, cross-stratified siliceous sandstone recognizable as "Potsdam" lithofacies. The entire Theresa becomes thinner and undergoes a facies change to dominantly cross-stratified clean sandstone southeastward against the Adirondack Highlands (Selleck, 1984).

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The middle and upper Theresa contain two sharply-defined lithofacies that alternate in vertical sequence. The first consists of grey, thick-bedded to massive, poorly sorted, medium- to coarse-grained, calcareous to dolomitic sandstone. This lithofacies is intensely bioturbated. Interbedded with grey sandstone is white, meter-thick, thin- to medium-bedded, fine- to medium-grained, siliceous to calcareous, planar and herringbone cross-bedded sandstone. This interbedding of distinct lithofacies does not occur in the lower Theresa. Upsection, the bioturbated sandstones are displaced by thicker, coarser, cross-bedded sandstones. Selleck (1984) reported increased dolomite content, mudcracks, intraformational conglomerates, and vuggy cryptalgal laminates in the upper Theresa.

The vertical alternation of lithologies described for the middle and upper Theresa resulted from migration of low intertidal sand flats. These were shoal-water areas that protected extensively burrowed interflat areas. The coarser, channeled, dolomitic sandstones in the upper Theresa represent high intertidal channel and overbank facies. The overlying Ogdensburg Dolostone contains stromatolites (Kerans, 1977; Selleck, 1987) and preserves supratidal facies.

ICHNOFACIES

The intertidal habitats preserved in the upper Potsdam Formation contain a *Skolithos* ichnofacies of low-level suspension-feeders dominated by *Diplocraterion. D. parallelum* is abundant, whereas *D. helmerseni* is rare. Escape burrows resembling *Monocraterion* (Hallam and Swett, 1966) are very common in one thick "white" Potsdam bed. Shallow *Skolithos* and *Monocraterion* burrows (3 to 6 cm) occur at most locations exposing the "white" Potsdam in the Thousand Island region, but *Skolithos* generally does not occur with *Diplocraterion.*

The principle of competitive exclusion (Dodd and Stanton, 1981) states that single-species dominated trophic groups are more common than not (Walker, 1972). The mutual exclusion of *Skolithos* and *Diplocraterion* in Cambrian tidal facies has been noted by many authors (in Cornish, 1987, p. 484). High population densities of *D. parallelum* in single thick-bedded sandstones are exposed at the "unit 1" location (Fig. 3). These beds approach "pipe-rock" density (Hallam and Swett, 1966; Swett et al., 1971), and indicate periods of relative substrate immobility, and probable diastems.

Bjerstedt and Erickson (1989) regard energy from tidal currents, rather than wave energy, as the predominant environmental parameter that was especially favorable for the trace-maker of *D. parallelum* in the upper Potsdam. The *D. parallelum* trace-maker was a tidalphile, and especially favored clean sands frequently mobilized by swift tidal currents on extensive low intertidal sand flats.

In the Potsdam, water motion due to swift tidal currents between, or on the margins of, Precambrian bedrock ridges (Selleck, 1984) provided an optimum habitat for *D. parallelum.* The Chippewa Bay outcrop exposes facies deposited in proximity to unusual relief on the Precambrian surface. The occurrence of abundant *D. parallelum* in the region appears ultimately due to anomalous Precambrian paleotopography in the Thousand Islands region.

The ichnofauna representing each component of the mixed ichnofacies occurs in a characteristic lithology. The distribution of trace fossils in the Theresa is attributable mainly to physical energy variation and persistence that is reflected in the grain-size and sorting of two distinct lithofacies. Grain-size and sorting is a reflection of the magnitude of environmental "energy levels", and also the degree of energy level persistence.

Paleodepth, *sensu stricto,* played no part in the distribution of the ichnofauna. Potsdam-Theresa facies were deposited entirely in peritidal to shallow subtidal facies where gross environmental energy levels, and the persistence of that energy, controlled the availability and type of trophic resources. Lithologic and textural criteria are important supporting evidence for interpretation of mixed trace-fossil assemblages. Otherwise, *a priori* assumptions may form the basis of recognizing ichnofacies when ichnotaxa assumed to be indicative of a particular ichnofacies are comingled.

The parameter of energy level persistence, rather than magnitude, represents the most important factor in the composition of the Theresa mixed *Skolithos-Cruziana* ichnofacies.

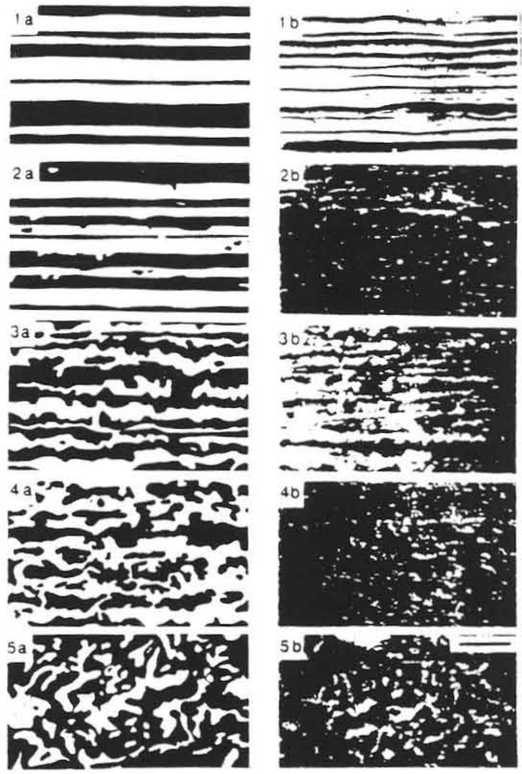

Figure 4: Schematic diagrams of ichnofabric indexes 1 through 5 with representative examples: schematic on the left; actual petrographic samples on the right. Scale bar is 10 cm. (From Drosser and Bottjer 1986, by permission)

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The parameter of energy level persistence, rather than magnitude, represents the most important factor in the composition of the Theresa mixed *Skolithos-Cruziana* ichnofacies. The consistency of environmental energy levels is a major influence on the adaptive strategy of trace-making animals. Equilibrium assemblages (Ekdale, 1985; Rhoads and Boyer, 1983; Vermeij, 1978) tend to represent specialists adapted to resource-limited environments where ecologic parameters are predictable over long periods of stasis. In contrast, opportunistic or pioneering assemblages tend to represent generalists adapted to fluctuating environments with ecologic parameters and food resources that can be near the edge of their tolerance (Ekdale, 1985). Vermeij (1978) recognized a third end-member strategy called stress-tolerant, that represents inhabitants of physiologically stressful environments such as the intertidal zone.

A characteristic ichnofauna occurs in each lithofacies. A *Cruziana* ichnofacies dominated by deposit-feeding burrows predominate in grey bioturbated sandstones. Among these deposit-

D. parallelum, *D. habichi*, *Skolithos*, and *Monocraterion* generally occur in horizontally-laminated, fine-grained, medium -bedded sandstone layers and lenses that occur in the lower 3 m of the Theresa. These beds represent sand splays washed into protected facies in Precambrian lows.

The *Skolithos* ichnofacies, dominated by suspension-feeders, is represented by shallow vertical tubes (3-6 cm) of *Monocraterion* and *Skolithos* in meter-thick, clean, cross-bedded sandstones. These sandstones record intertidal sand shoals or flats that successively migrated across bioturbated, low inter-flat facies.

ICHNOFABRIC CLASSIFICATION

In our studies we have used the concept of ichnofabric classification as developed by Drosser and Bottjer (1986, p. 558 & 559). The technique involves visual comparison of outcrops with a set of flashcards upon which a scaled area of bioturbation is portrayed. We have reproduced the illustration and description from their work for use in the field (see Figure 4).

Alternation of burrowed grey sandstones and white cross-bedded sandstones characterizes the Theresa from the middle to the top of the formation. The log of ichnofabric index for the Theresa in Figure 5 shows that in 43 m of total thickness, 24.8 m constitute grey, bioturbated sandstone, and 18.2 m constitute white, cross-bedded sandstone. For the grey sandstones, 100% of this thickness is Ichnofabric Index 4. For the white sandstones, 10.6 m (58%) are index 3, and 7.6 m (42%) are index 2. Thickness relationships among Bioturbation Units (BU's) are shown in Figure 6 (Bjerstedt and Erickson, 1989). A verbal descripttion of the Index of Droser and Bottjer (1986) is given below:

1) No bioturbation recorded; all original sedimentary structures preserved.
2) Discrete, isolated trace fossils; up to 10% of original bedding disturbed.
3) Approximately 10 to 40% of original bedding disturbed. Burrows are generally isolated, but locally overlap.
4) Last vestiges of bedding discernable; approximately 40 to 60% disturbed. Burrows overlap and are not always well defined.
5) Bedding is completely disturbed, but burrows are still discrete in places and the fabric is not mixed.
6) Bedding is nearly or totally homogenized.

(From Drosser and Bottjer, 1986)

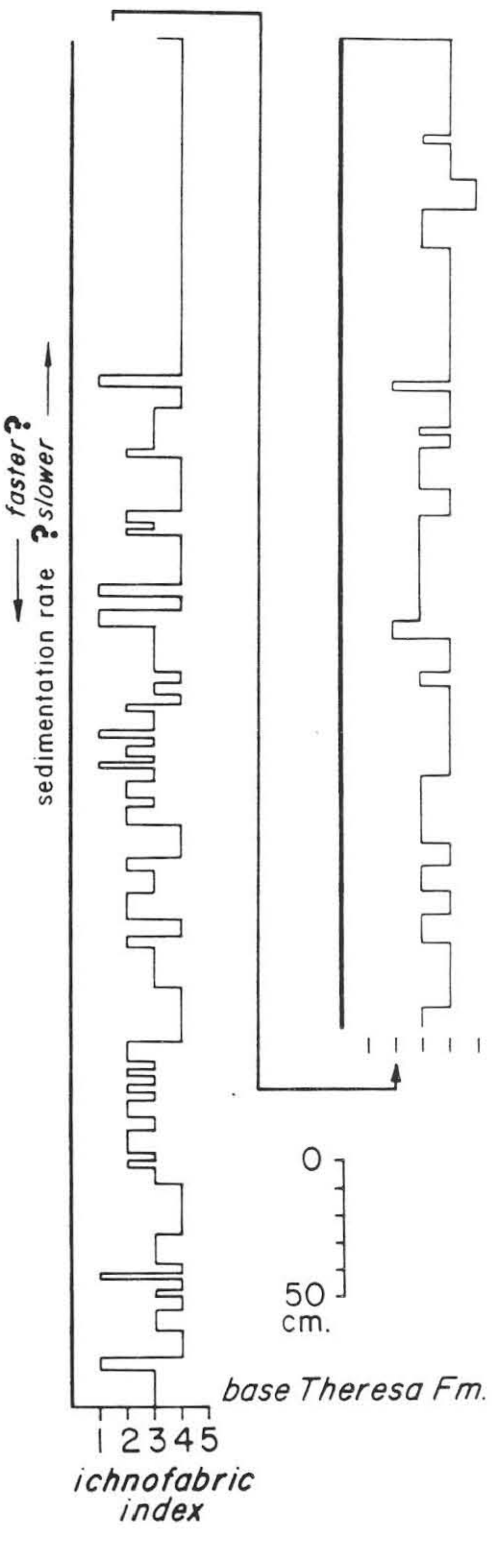

Figure 5: Detailed strip log showing bioturbation units defined by changes in ichnofabric index in unit 7 of the lower Theresa Formation at Chippewa Bay (Fig. 9). The relationship of ichnofabric to sedimentation rate is discussed in the text.

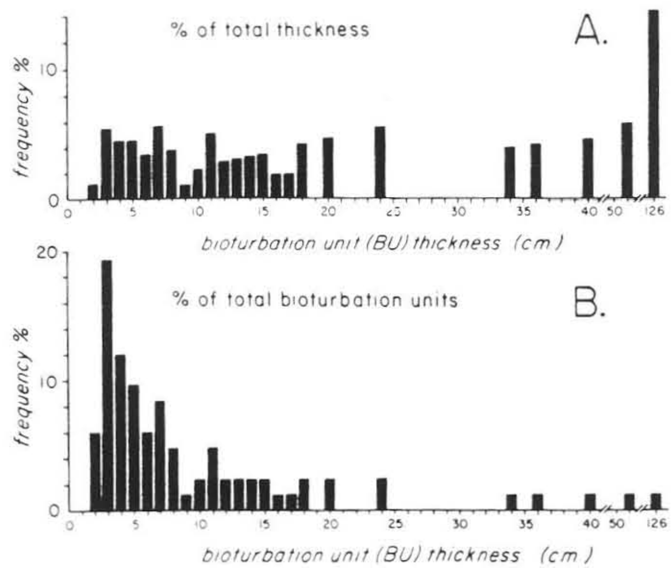

Figure 6: Thickness frequency of bioturbation units (BUs) from unit 7 of the Theresa Formation. **A**) Graph of BU cumulative thickness. **B**) Graph of BU thickness frequency. Endobenthic burrowing depth from -3 to -5 cm in shallow, protected subtidal facies is suggested by the BU thickness frequency cluster about 3-5 cm.

feeders are, *Fustiglyphus*?, *Gyrochorte*?, *Neonereites uniseralis*?, *Phycodes flabellum*, *Planolites beverlyensis*, *Rosselia socialis*, and *Teichichnus*. Suspension-feeders include *Diplocraterion habichi*, *D. parallelum*, *Monocraterion*, shallow-burrowing *Skolithos*, and possibly *Palaeophycus tubularis* (Pemberton and Frey, 1982). *Cruziana*? furrows are attributed to inferred scavenging or deposit-feeding trilobites.

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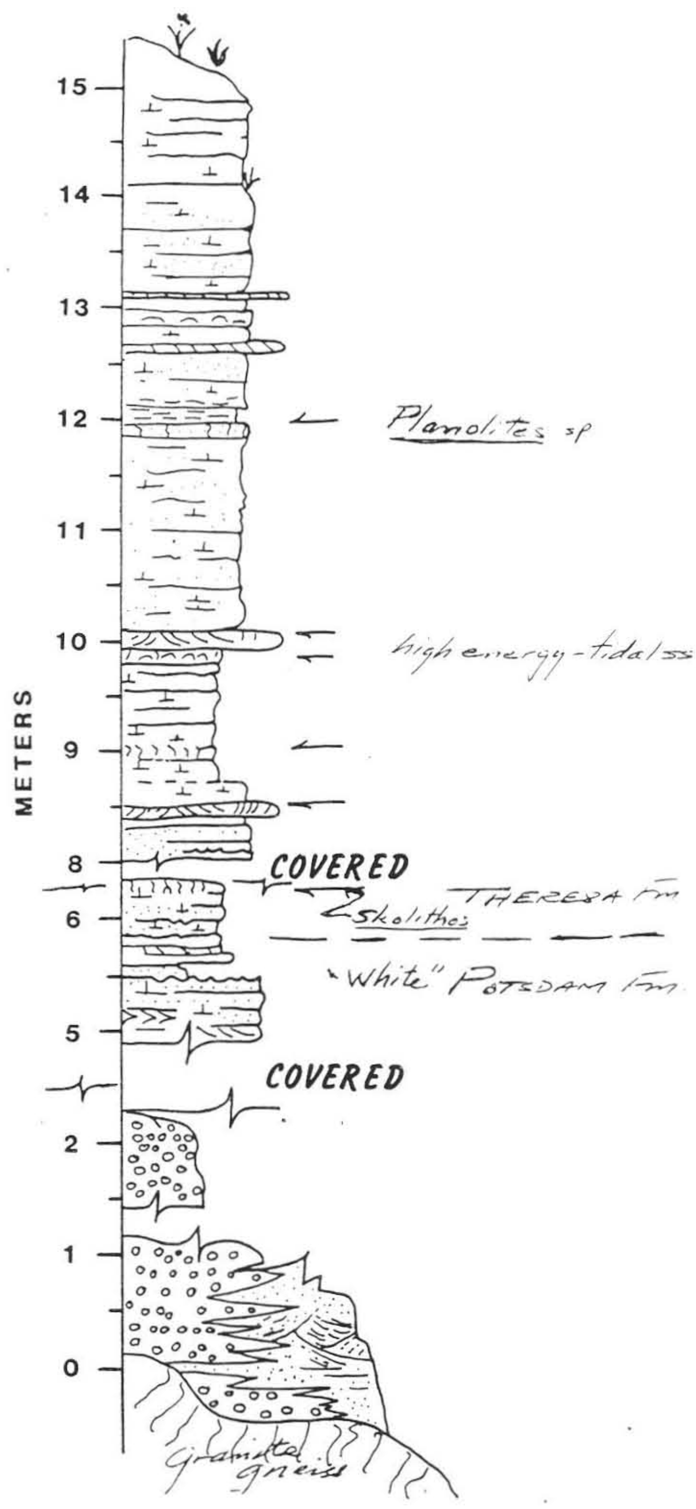

Figure 7: Stratigraphic column at Oak Point drawn to emphasize high energy "white Potsdam" sandstones that might have been a source for *Protichnites* (?) sp.

minimum burrow depths of *D. parallelum* were made through the upper Potsdam at this section. Data taken from Bjerstedt and Erickson (1989) are presented here in Figure 10.

SECTION AT OAK POINT

Although we will repeatedly see exposures of Theresa Formation along highway 12 there are few places where one can get a feeling for the stratigraphic relationships of the section without trespassing extensively on private property. The rocks at Oak Point provides a good opportunity to view much of the section without undue encroachment on private land. Please exercise care to stay on the roadway unless given the OK to explore, however.

At Oak Point more than 15.4 meters (50 feet) of strata are available for study, presenting most of the Potsdam-Theresa interval. The Potsdam lithologies at this locality are not fossiliferous in the lower portion and there may be some indications from the conglomerates and the magnitudes and types of cross-stratification that the Potsdam here is not marine in its base. A composite stratigraphic column made up the Oak Point Road is given in Figure 7.

The Potsdam-Theresa contact can be placed at this locality by the first occurrence of significantly-bioturbated, calcareous, grey sandstones. Potsdam lithologies re-enter the section at several positions in the adjoining outcrops, and most show some evidence of marine traces. Contrasting conditions of burrowing are easily seen in the upper half of this section.

When this outcrop was first created, approximately 25 years ago, St. Lawrence students collected from the outcrop debris the well preserved trackway of a large organism. This was illustrated by Bjerstedt and Erickson (1989) in Figure 15G but was not assigned. As we visit Oak Point it is appropriate to discuss the implications of the specimen further. The specimen (Figure 11A, 11B) is a meter-long slab of thin-bedded, dolomitic sandstone that reveals a trackway of evenly-spaced (evenly-paced) digit impressions from a multi-legged, bilaterally symmetrical, elongate organism that probably dragged a substantial posterior body element through the wet sand of a Theresa tidal flat. Width of the trackway is 23 cm, but width of the organism's body is not defineable from the trackway.

At least six legs are defineable from the pattern of impressions. One of these had trifid digitation (Figure 12A) and alternated with a bifid blade-like appendage in the walking sequence. Shorter, more centrally-located monodactylous walking legs (at least one pair) were shorter and were carried more centrally. These left much fainter impressions best seen on the whitened specimen (Figure 11B) and on a plaster cast made directly from the specimen (Figure 12B).

It seems certain that the trackway was the work of an arthropod moving at a steady gate through almost saturated sand. The identity of the organism is not recognizeable from the specimen. Assignment to an ichnogenus also would be premature in as much as the literature of Paleozoic arthropod trackways presently is confused and somewhat contradictory.

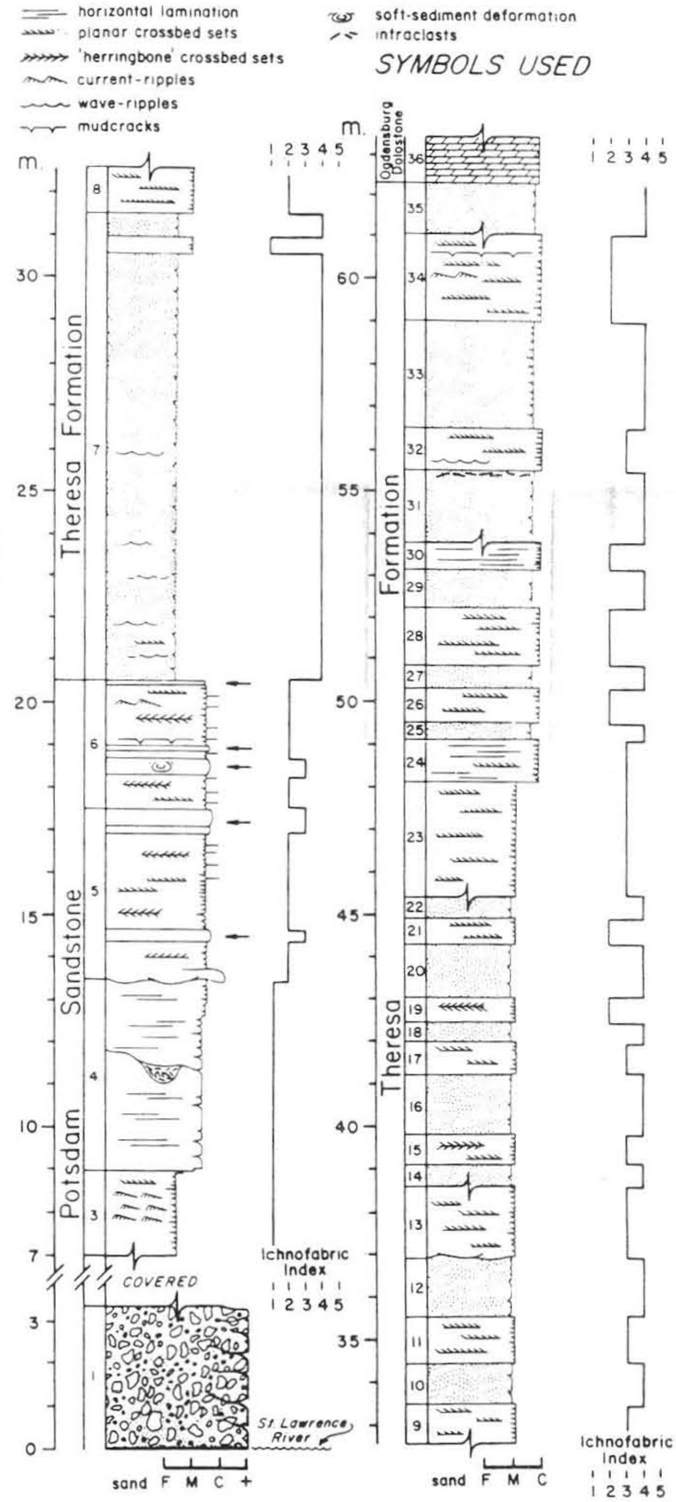

Figure 8: Composite section of the Potsdam and Theresa Formations in the Thousand Islands region showing bioturbation intensity based on ichnofabric indicies (Droser and Bottjer, 1988a). Arrows show medium- to thick-bedded sandstones containing *Diplocraterion* sp. in the "whit " Potsdam.

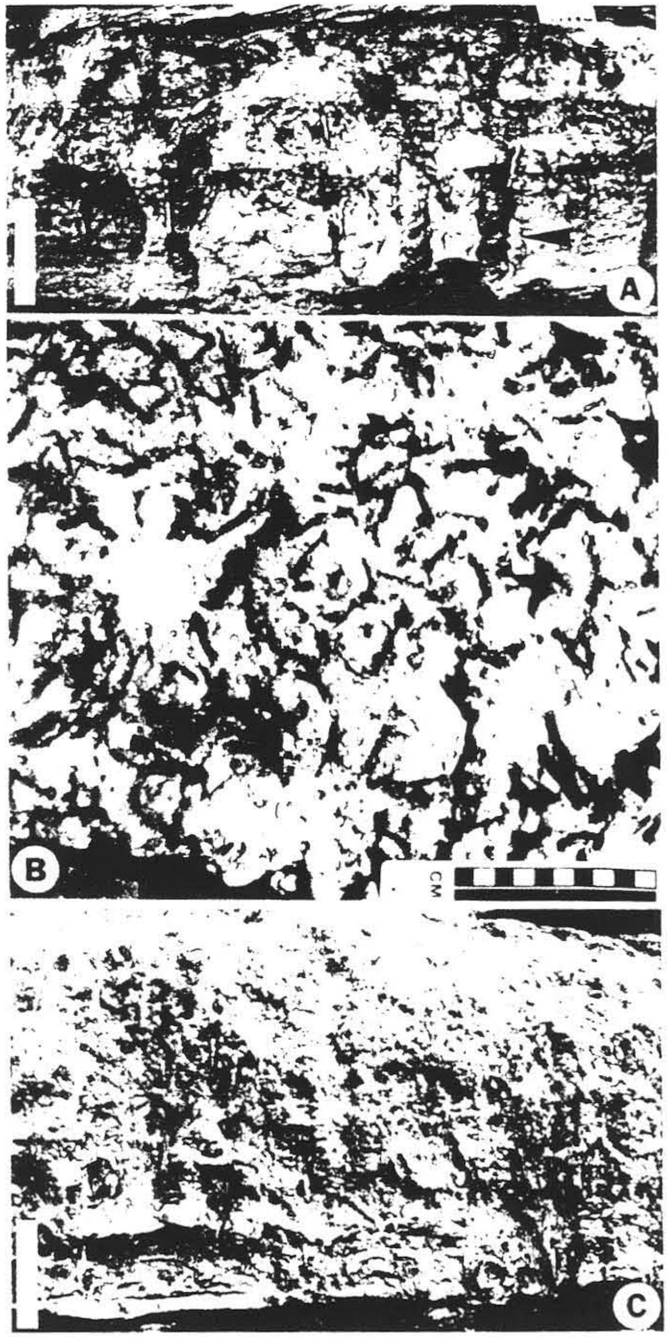

Figure 9: *Diplocraterion* sp. in the "white" Potsdam; bar scales = 10 cm. **A**) Clean sandstones containing *D. parallelum* at Chippewa Bay. Protrusive specimen at arrow may be escape shaft. **B**) Concave epireliefs on bedding plane exposure at unit 1 location. **C**) Thoroughly bioturbated sandstone bed in the "white" Potsdam at Chippewa Bay showing amalgamated *Diplocraterion* sp. tubes and escape burrows, some resembling nested funnels of *Monocraterion* sp.

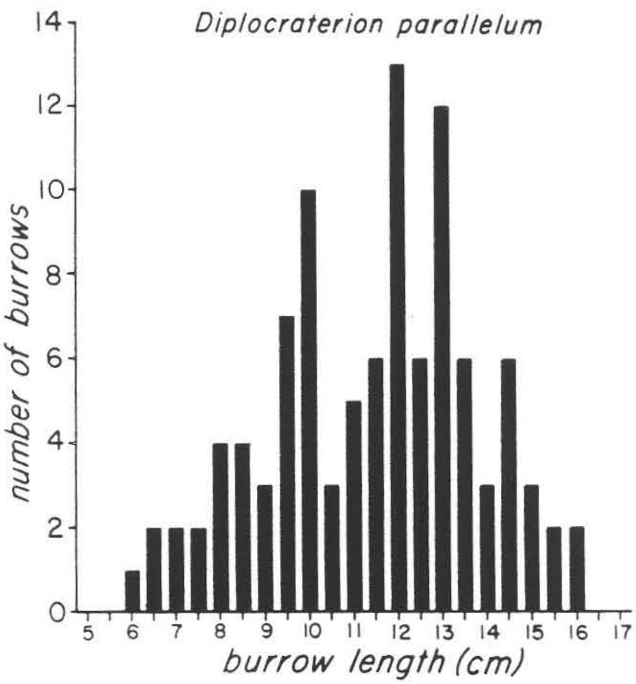

Figure 10: Histogram of minimum burrow depth for protrusive *Diplocraterion parallelum* from the "white" Potsdam Formation at Chippewa Bay, N = 102.

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Although we will repeatedly see exposures of Theresa Formation along highway 12 there are few places where one can get a feeling for the stratigraphic relationships of the section without trespassing extensively on private property. The rocks at Oak Point provides a good opportunity to view much of the section without undue encroachment on private land. Please exercise care to stay on the roadway unless given the OK to explore, however.

At Oak Point more than 15.4 meters (50 feet) of strata are available for study, presenting most of the Potsdam-Theresa interval. The Potsdam lithologies at this locality are not fossiliferous in the lower portion and there may be some indications from the conglomerates and the magnitudes and types of cross-stratification that the Potsdam here is not marine in its base. A composite stratigraphic column made up the Oak Point Road is given in Figure 7.

The Potsdam-Theresa contact can be placed at this locality by the first occurrence of significantly-bioturbated, calcareous, grey sandstones. Potsdam lithologies re-enter the section at several positions in the adjoining outcrops, and most show some evidence of marine traces. Contrasting conditions of burrowing are easily seen in the upper half of this section.

When this outcrop was first created, approximately 25 years ago, St. Lawrence students collected from the outcrop debris the well preserved trackway of a large organism. This was illustrated by Bjerstedt and Erickson (1989) in Figure 15G but was not assigned. As we visit Oak Point it is appropriate to discuss the implications of the specimen further. The specimen

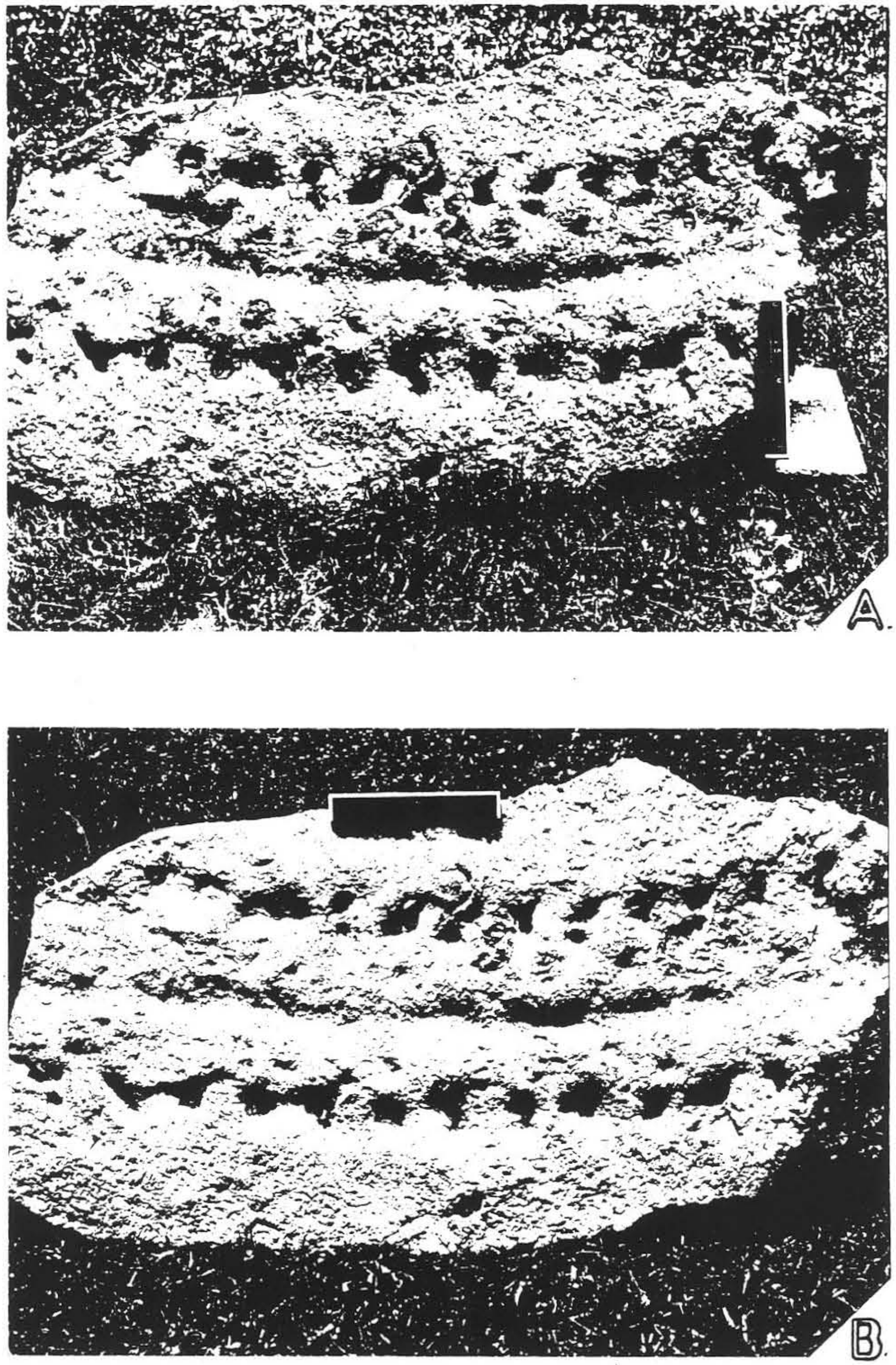

Figure 11: Slab of white sandstone from Oak Point showing trackway of *Protichnites*(?)-like ichnofossil. Scale bar = 15 cm. **A**) Specimen unwhitened. **B**) Specimen whitened to accentuate the depressions of short walking legs adjacent to groove.

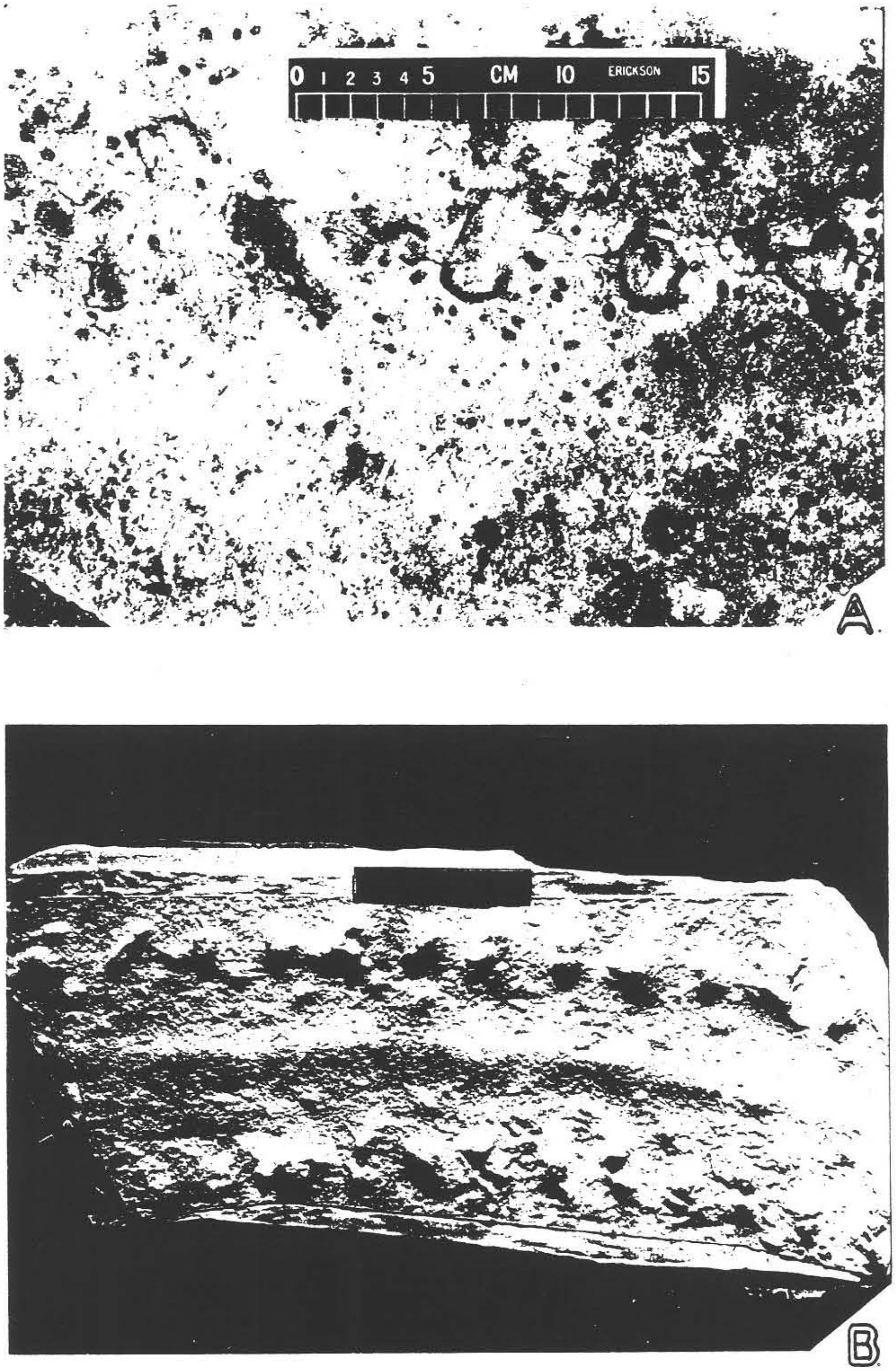

Figure 12: *Protichnites*?)-like ichnofossil. **A**) Note the trifid impressions alternatng with bladeed bifid impressions on this region seen just below scale bar in 12B. **B**) Plaster cast of the tracks presenting them in epirielf as if on the base of the bed above. Scale bar is 15 cm.

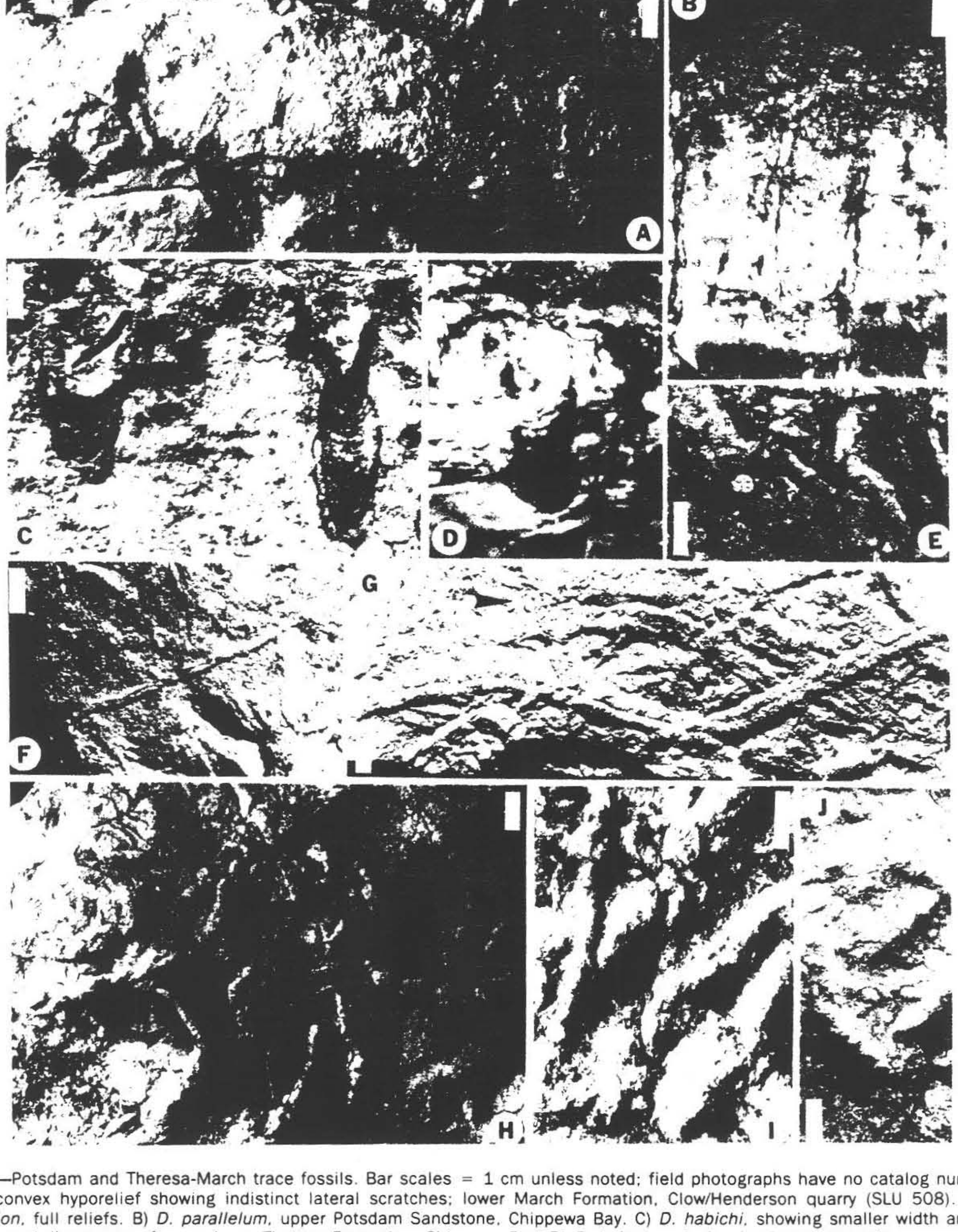

FIGURE 13—Potsdam and Theresa-March trace fossils. Bar scales = 1 cm unless noted; field photographs have no catalog numbers. **A)** *Cruziana*? convex hyporelief showing indistinct lateral scratches; lower March Formation, Clow/Henderson quarry (SLU 508). **B, C, D)** *Diplocraterion*, full reliefs. B) *D. parallelum*, upper Potsdam Sandstone, Chippewa Bay. C) *D. habichi*, showing smaller width and inward pinching/upward divergence of arms; lower Theresa Formation, Chippewa Bay. D) *D. helmerseni*, showing basal expansion; upper Potsdam Sandstone, Chippewa Bay. **E)** *Fustiglyphus*?, full relief in float block, black arrow shows single swelling, Theresa Formation, Oak Point. **F, G)** *Gyrochorte*?, convex epireliefs. F) Small specimen showing no plait structure, lower Theresa Formation, Chippewa Bay (SLU 509). G) Large specimen showing subtle plait structure, black arrow shows *Monocraterion*; lower Theresa Formation, North Hammond quarry (SLU 510). **H, I, J)** *Neonereites uniserialis*?, convex hyporeliefs, lower March Formation, Clow/Henderson quarry. H) (SLU 511), black arrow shows *Teichichnus*. I) (SLU 512). J) (SLU 513).

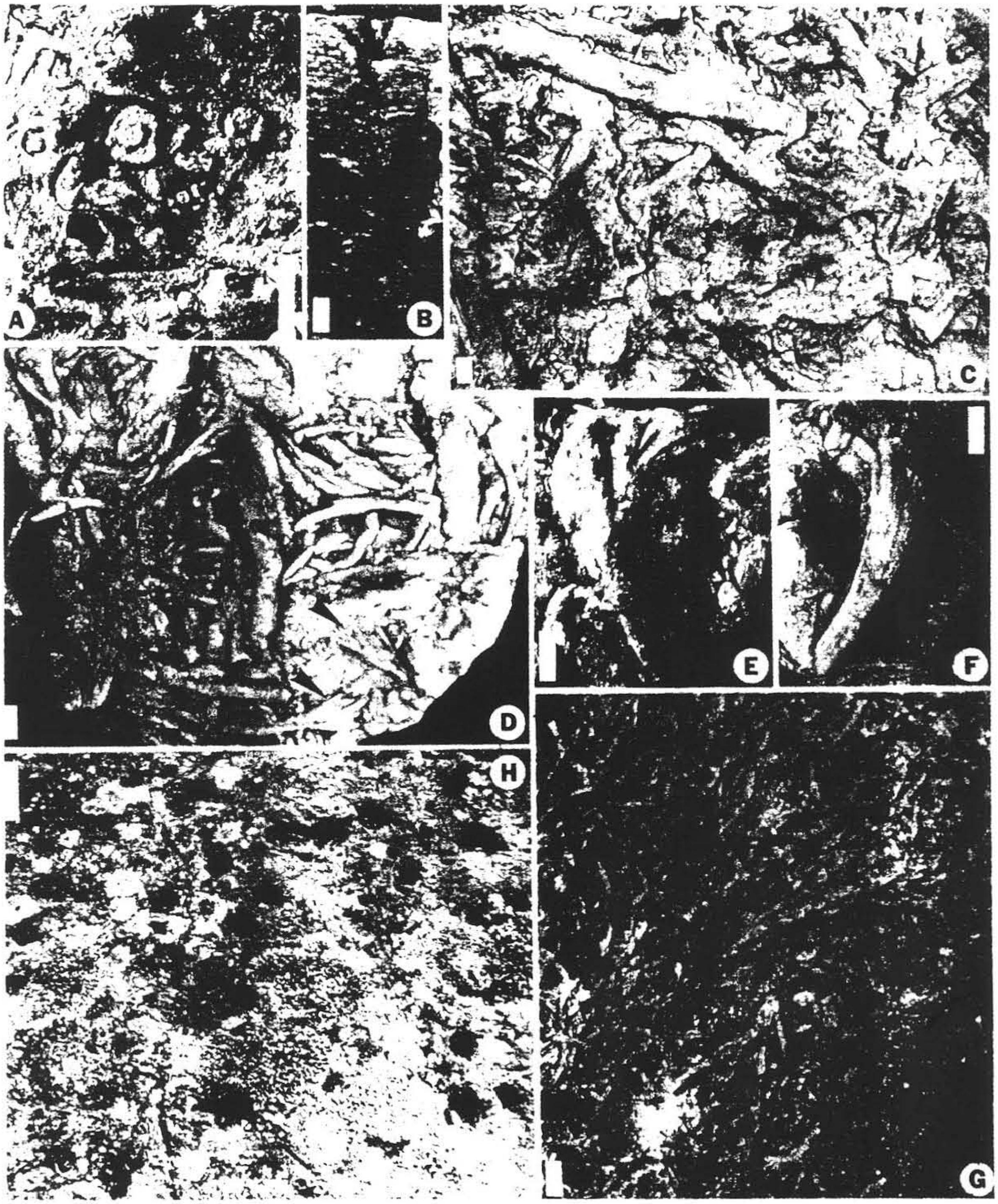

FIGURE 14—Theresa-March trace fossils. Bar scales = 1 cm unless noted; field photographs have no catalog numbers. **A, B)** *Monocraterion.* A) convex epirelief, lower Theresa Formation, Chippewa Bay (SLU 514). B) full relief from thin-bedded sandstone at the top of the Chippewa Bay outcrop (Fig. 5), Theresa Formation. **C)** *Palaeophycus tubularis,* full relief, lower Theresa Formation, Chippewa Bay (SLU 515). **D, E, F)** *Phycodes flabellum,* convex hyporeliefs, lower March Formation, Clow/Henderson quarry D) Slab with branching *P. flabellum* (black arrows), and abundant unbranched *Teichichnus,* (SLU 516). E) (SLU 517). F) (SLU 518). **G)** *Planolites beverlyensis,* convex hyporelief, lower Theresa Formation, North Hammond quarry (SLU 519). **H)** *Rosselia socialis,* concave epireliefs on a lichen-covered bench surface, bar scale = 5 cm, lower Theresa Formation, North Hammond quarry.

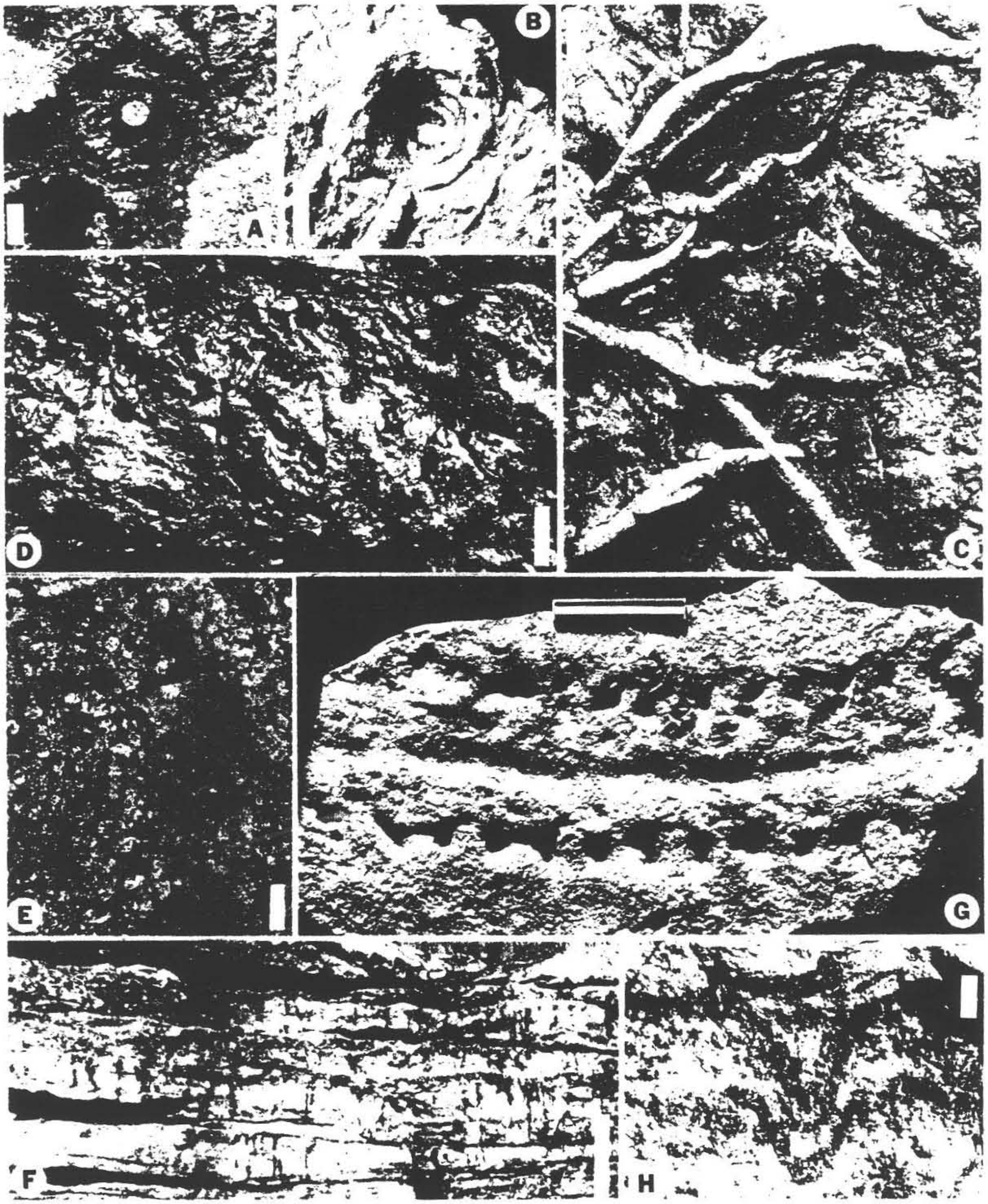

FIGURE 15—Theresa-March trace fossils. Bar scales = 1 cm unless noted; field photographs have no catalog numbers. **A, B)** *Rosselia socialis*. A) Concave epirelief showing basal tube on same bench surface in Fig. 14H, lower Theresa Formation, North Hammond quarry. B) Concave epirelief, lower Theresa Formation, Chippewa Bay (SLU 521). **C, D)** *Teichichnus*. C) convex hyporeliefs of discrete burrows, lower March Formation, Clow/Henderson quarry (SLU 524). D) weathered full reliefs on outcrop showing spreite, lower Theresa Formation, Chippewa Bay **E, F)** *Skolithos*. E) Epireliefs of sediment filled vertical tubes, lower Theresa Formation, Chippewa Bay (SLU 522). F) Shallow vertical burrows in meter-thick, thin-bedded sandstone at the top of the Chippewa Bay outcrop (Fig. 5), Theresa Formation, bar scale = 15 cm. **G)** Arthropod trail, concave epirelief from meter-thick, thin-bedded sandstone in the lower-middle Theresa Formation, Oak Point, bar scale = 15 cm (SLU 520). **H)** Escape burrow showing down-bent laminae, upper Theresa Formation, (unit 34; Figs. 3, 4).

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TRIP A3 (3)

A PRELIMINARY EVALUATION OF DUBIOFOSSILS FROM THE POTSDAM SANDSTONE

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ABSTRACT

Preliminary results of study of a meter-long block of Potsdam Sandstone (Cambrian) from Malone, New York, indicate that the contained structures are not body fossils although the sub-cylindrical and sub-triangular morphologies of the 22-24 cm - long dubiofossils are very suggestive of once-living structures. Wall structure included four layered units, two of which are known only from the void that remains after their dissolution(?). The medial layer is of clay and silt whereas the outer layer is of medium and coarse quartz sand grains. All structures are curved and preserved in the stable, concave down position. It appears that the structures are mud curls, or clay rolls - "tonrollen," that have included two thin algal layers in their fabric. The algal layers were likely responsible for the quality of preservation of these unusual specimens.

INTRODUCTION

In 1992, during excavation at a residence in Malone, New York, a boulder of Potsdam Sandstone was brought to the surface and dropped, whereupon it parted on a bedding plane revealing remarkable structures of uncertain origin. Appearance of the boulder, and its proximity to another, coupled with local geomorphic setting, suggest that Potsdam Sandstone bedrock lies close by and that the boulder has not been glacially transported.

Local residents recognized the unique character of the specimen and alerted members of the St. Lawrence Geology Department. The owners subsequently permitted removal of the slabs to the laboratory in Canton for study. They have since been returned to the owner in Malone (Figure 1).

THE BOULDER

The boulder is a meter-long, trapezohedral slab of Potsdam Sandstone with length determined by joint or fracture plane and thickness by parting on parallel bedding planes

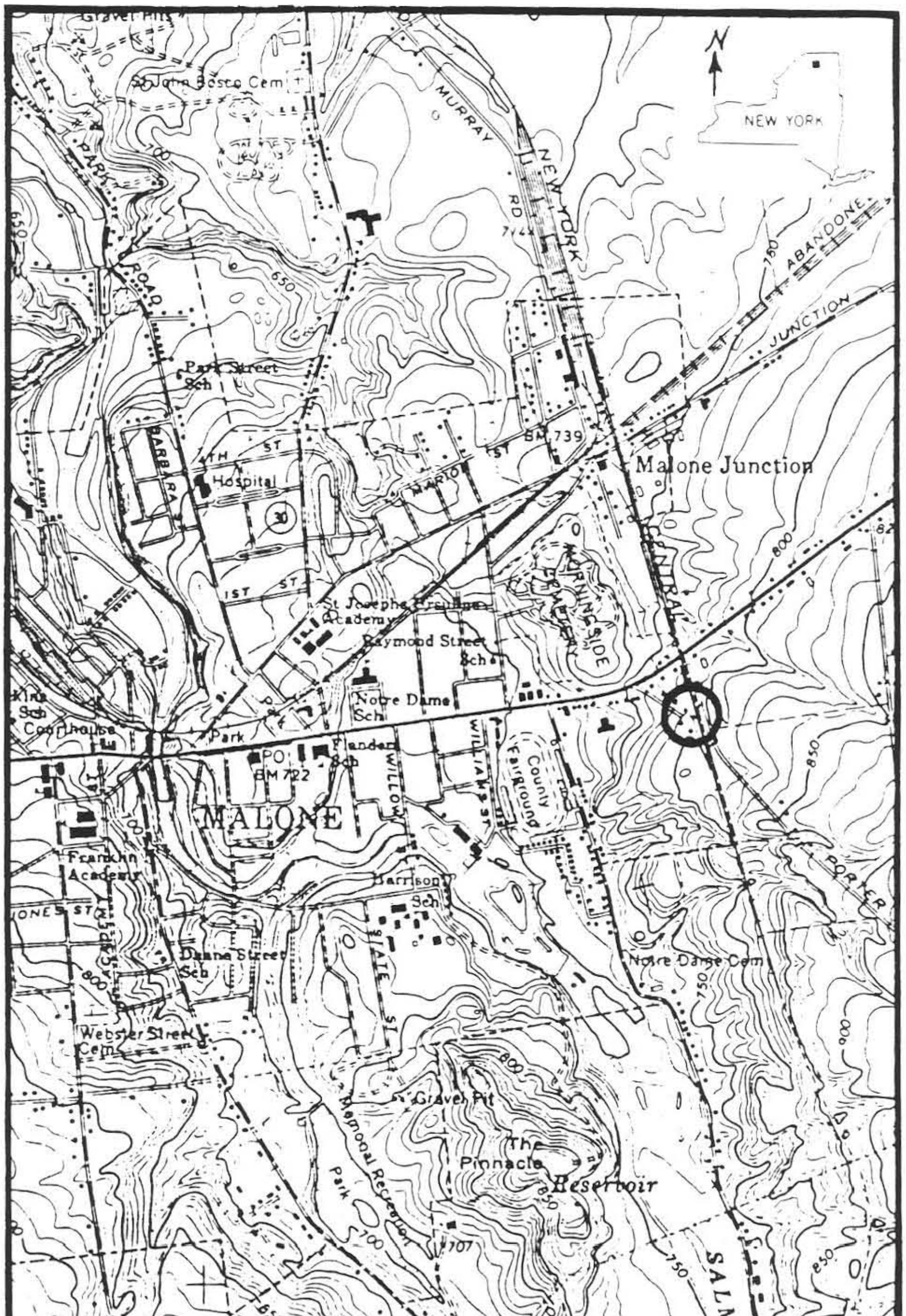

Figure 1: Portion of the U.S.G.S. Malone 71/2 minute topographic quadrangle indication the approximate location of discovery of the unusual Potsdam Sandstone specimen.

Figure 2: Lower parting surface of Potsdam Sandstone slab exposing enigmatic structures which occur as raised elements (or convex epireliefs) on this surface. Knife is 9.0 cm long.

Figure 3: Upper parting surface preserving structures as concave hyporeliefs in Potsdam Sandstone. Knife is 9.0 cm long.

approximately 45 cm apart. Edges are gently-rounded angles. It is composed of fine- and medium-grained quartz sandstone that is plane-laminated on a millimeter scale. Some laminae are colored by dark maroon hues presumed to result from hematite coatings on grains. The principal cement is silica.

Examination of cross-sectional surfaces indicates no cross-stratification present. Stratigraphic up is suggested only by some very tentative channeling that seems to cut out a few laminations. There are no trace fossils present. If correctly interpreted, the rock split on a bedding plane lying approximately 7.5 cm from the stratigraphic top of the slab when it was dropped. It parted to expose 35 concave, curved, sub-cylindrical to sub-triangular, dark maroon features on the lower surface (Figure 2). The counterpart, on the upper surface of the bedding plane, exposed the convex replica of the same features (Figure 3) with a few exceptions due to differences of parting fracture. All seem to have lain concave down.

THE STRUCTURES

The structures themselves show as much as 2.5 cm of relief when seen in cross-section on sides of the slab. Curved relief of approximately 1 cm is common for most bedding surface elements. They are gently curved throughout or relatively flat centrally with curved margins. Among the 35 individual elements no two are precisely the same, however, there are two or three general shapes that recur. Most obvious is the sub-triangular form (Figure 4) and varieties of it that seem to result from bending or breaking of triangular structures during deposition as seen in Figure 5. Elongate, sub-cylindrical forms (Figures 2 & 3) occur in several lengths and widths and some, likely incomplete, are nearly equidimensional. Forms are evenly distributed across the bedding plane and obviously extended beyond the confines of the slab at hand. Greatest length for both forms is consistantly between 22 and 24 cm. "Width" of sub-triangular types is between 12 and 15 cm based upon three specimens, one of which was a reconstruction.

Each individual displays, on part and counter part, a smooth, maroon-colored clay(?) surface, patterned with transverse grooves, or channels, that are presently filled with white, very fine-to-fine-grained sandstone matrix. Grooves are arrayed perpendicular to the long axis of each specimen and occur at rather regular intervals of 1.5 to 2 cm. On the sub-triangular forms these are intersected centrally by an axial pattern as well. Margins are smooth and all termini are rounded. There are no angular features excepting those that can be assigned to breakage during transport or burial.

Encasing walls, if wall is an appropriate term to apply to a feature that separates two deposits, do not close upon themselves, although some nearly meet and in one instance a wall rolls "inside" itself. The nature of the wall structure itself is of interest. When the parts were separated the wall material was lost. Consequently the features preserved are molds in most instances. On some specimens it appears the wall structure was missing on burial so that matrix adheres to matrix with no wall between.

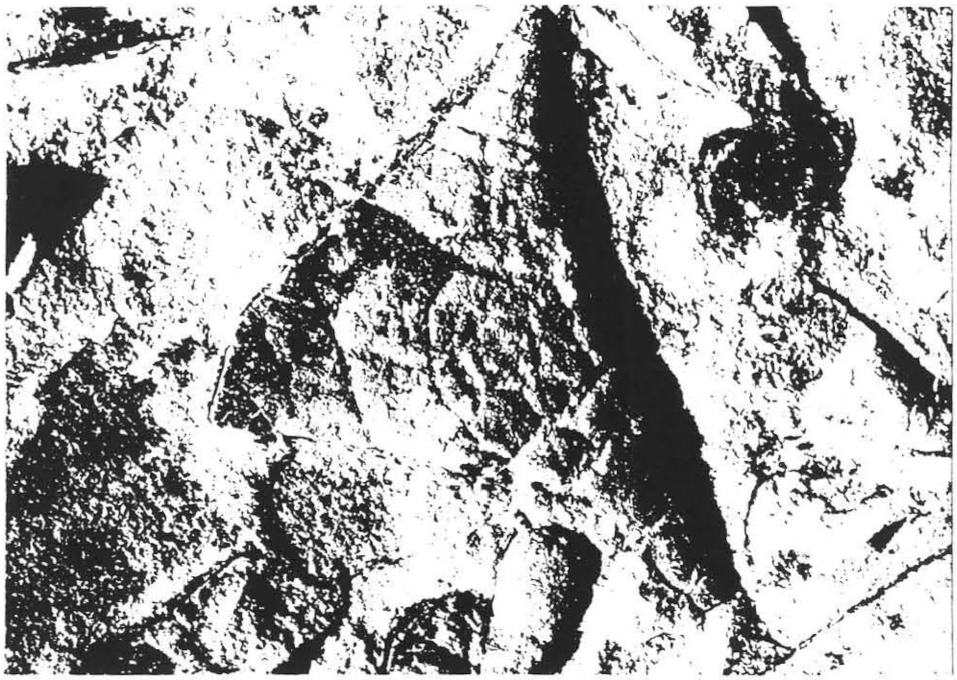

Figure 4: Sub-triangular element preserved as convex epirelief. Longest edge is 22 cm long. Note rounded tips and pattern of matrix-filled grooves resembling mudcracks.

Figure 5: Sub-triangular structure that has been centrally crushed and bent during deposition. Length of straightest margin, tip to bend, is 14 cm. Note spaces where wall layers were not preserved on this and adjacent structures.

THE WALL

Although not present over the surface, wall material is available where the structures are embedded in the sandstone. Specimens shown in Figure 5 display many of the pertinent relationships. Walls were approximately 1.5 mm thick and layered as indicated in Figure 6. Preservation shows the walls to have been three-layered, more probably four-layered, units, two of which are not preserved but appear

as voids (Fig. 5, 6). Between the missing layers was a layer of silt in a clay matrix (Figures 7 and 8). The outer surface of the structures seems to have been a layer of medium and coarse quartz sand one or two grains in thickness. These grains are embedded in a clay matrix and seem to show pressure solution on the portion of the grain directed toward the outer wall (Figure 9).

Perhaps the most puzzling, and potentially important, observation is that two layers have not been preserved. Each was approximately 0.5 mm thick, perhaps greater when fresh, and they seem to have been uniformly distributed over the surfaces of all forms because they are found universally around all intact margins. Complete coverage is, obviously, an inference. Voids imply the absence of layers due to non-preservation. Dissolution after induration is the probable cause because there seems to have been no collapse of matrix into the voids. What was the composition of these lost wall layers? SEM examination of surfaces of the preserved midwall did not reveal any structure or surface texture that might shed light on the nature of this missing material. Might it have been organic in composition?

COMPARISONS

The aspect of these structures makes them likely candidates for interpretation as arthropod exuvia. Discussions with Dr. Ellis Yochelson (oral communication, 1992), and Dr. Hans Hofmann (written communication, 1992) have pointed out that similar objects have been described as arthropod carapace fragments Tillyard (1936), "fossil-like" objects (Elston and Scott, 1972), and mud curls, or clay rolls (Voigt, 1972). Both Cloud (1973) and Hofmann (1971) have warned of similar structures that may easily be confused for fossils, but should be regarded as pseudofossils or dubiofossils (Hofmann, 1972) because they preserve no biogenic structure.

Preliminary comparisons with the literature show that pseudofossils from the Precambrian Troy Quartzite of Arizona illustrated on the cover of Geotimes by Elston and Scott (1972) are quite similar in form. These are re-illustrated in a discussion of pseudofossils in Häntzschel (1975, p. W169). No triangular elements are defined in that specimen, but sub-cylindrical forms are quite similar to those on the Malone specimen.

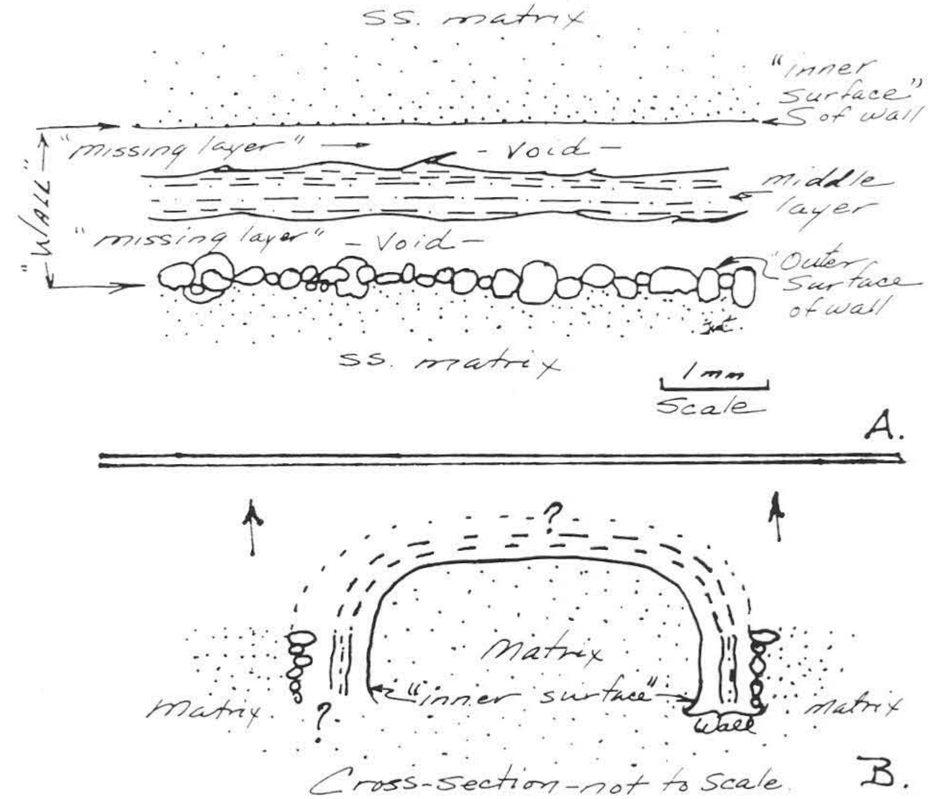

Figure 6: Sketch of wall structure (A) and relationships of wall layers with reference to stratigraphic up (arrows) on a cross-section of Malone dubiofossil specimens (B). "Depositional up" of the proto-wall mud layers was likely coarse-layer down.

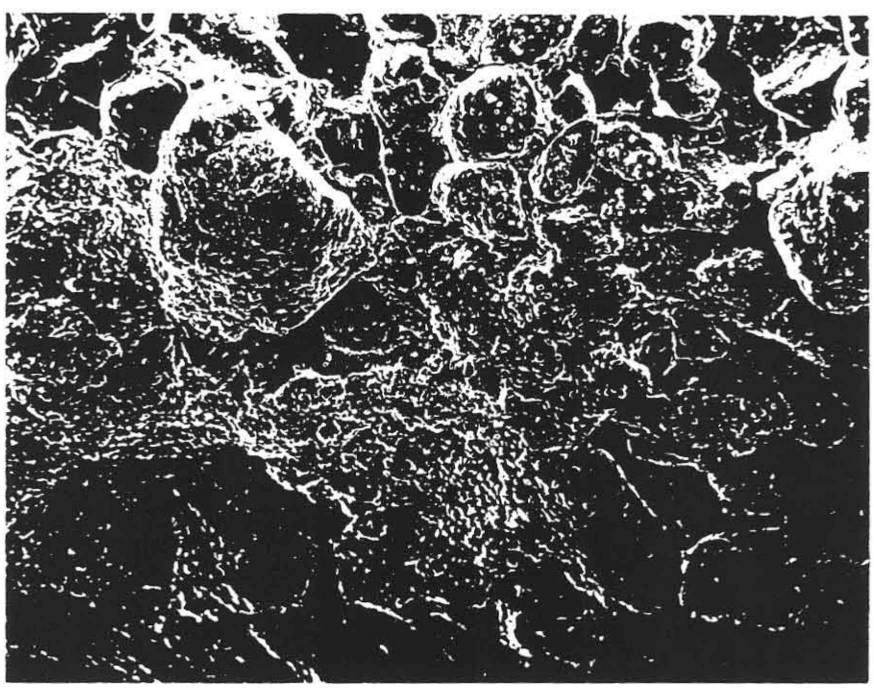

Figure 7: SEM photomicrograph of the outer (depositionally upper) surface of the middle layer of clay-covered silt grains. Magnification 470X.

Figure 8: SEM photomicrograph of clay from the inner surface of the middle layer. Magnification approximately 7,500X.

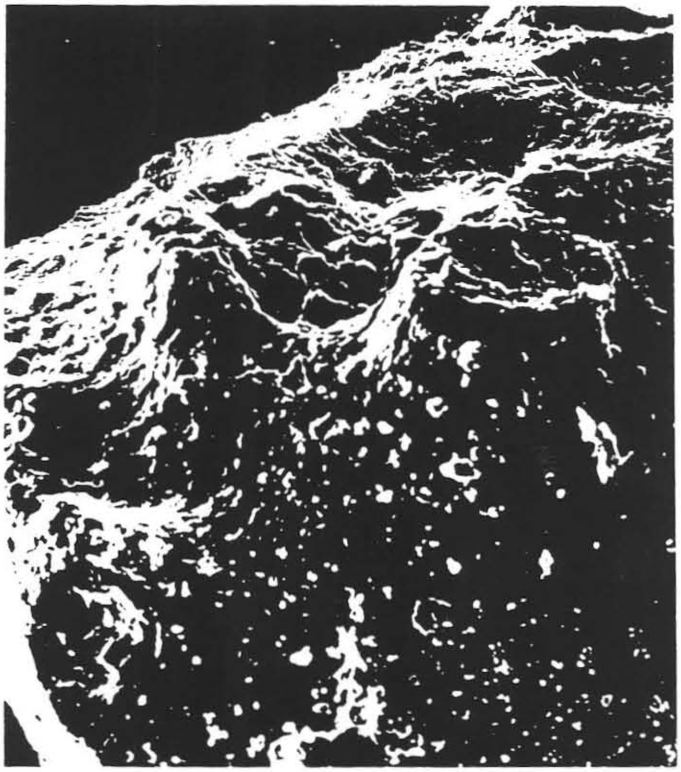

Figure 9: SEM photomicrograph of portion of the surface of a single grain of coarse quartz sand plucked from the outer edge of the wall. Note elongations and grooves. Although interpreted as a pressure solution surface, this may represent secondary silica cement coating the grain. Magnification approximately 1,300X.

INTERPRETATIONS

Repetition of general forms, rugosity and continuity of surfaces, even the degree of separation on the bedding plane, all contribute to the impression that these structures are organic remains. Their study was approached as a test of the hypothesis of an organic origin. Observations were made as if an investigation of skeletal elements and organismal morphology were being conducted.

No morphological features assignable to mesozoan or metazoan invertebrates were recognized. Specific shapes were not repeated, although general shapes recur as noted above. Wall structure preserves only inorganic materials and structures. A medial clay layer seems to verify that clay rolls, or "tonrollen", are the objects in question. It appears that a multi-layered clay drape was desiccated on a Potsdam Sandstone tidal plain. An incoming tide, or some similar rising water event such as a storm surge, brought water to gently mobilize portions of a clay layer that had rolled as it dried.

In Spring of 1993, Yochelson and I performed a simple experiment using exact morphological replicas of individual specimens from the Malone slab in a recirculating flume. Aluminum foil replicas were picked up and carried easily when concave up, but were forced onto the bottom by the current when they rolled into the concave-down position. In this position they were very stable and were readily covered by migrating ripples. It appears that the specimens occur in the rock in the stable position which is in accord with earlier observations about orientation of the boulder. Their interpretation as clay rolls seems a possibility.

Presently, the best interpretation is that they are interesting, non-biogenic structures. This is not to say that further examination, or better yet an alternate interpretation, may provide other insights in the future. Certainly, some questions remain unanswered. Why were two layers of the curls missing? What has been lost from those layers of the specimen? Why are there no desiccation cracks in the termini of any specimens? How far have they been moved before deposition?

Elston (1975, *in* Häntzschel) apparently noted a relationship between algae and the rolled specimens in the Arizona samples. It seems very likely that algae actively bound the layers together in the Malone specimen as well. After close examination it appears that the two layers missing from the framework of the wall were biogenic. These layers were probably made of closely "woven" algal mats. Such structures aided the clay to be flexible yet sturdy and thus to withstand transport more readily. Precise paleoenvironment of origin and deposition have not been determined as yet.

Presence of similar structures in Late Precambrian and Cambrian rocks in several parts of the world implies that similar conditions existed in those regions. One may wonder why no occurrences of these inorganic(?) structures are described from younger rocks? Could it be that conditions for forming these structures were constrained temporally? Perhaps bioturbation has been responsible for not permitting algal mats to be preserved extensively, and perhaps bioturbating organisms destroy algal mats by either eating them, or simply by breaking them up a great deal so they are destroyed by current action. Has such biologic action kept these dubiofossils from appearing in the rock record after the Cambrian? Might their absence from post-Cambrian rocks hold information about the evolution of, or developments in, feeding styles?

More discoveries of these unique structures will have to be made in other units before their exact relationships become understood. Presently, they must be considered curled algal mats with interlayered mud built up on a bed of sand lamina. Hofmann's (1972) term "dubiofossil" is an appropriate assignment.

ACKNOWLEDGMENTS

The author is grateful to Mr. Michael Meagan, owner of the specimen, for bringing it to his attention and permitting its removal for study. Mr. and Mrs. Lahart graceously allowed truck access across their property. Witherbee and Whalen transported the specimens skillfully. Drs. Ellis Yochelson, Hans Hofmann, and Michael R. Owen made valuable suggestions regarding the nature of the specimens. Lance Erickson helped with the initial observations and Michael Whitton with lab study. I appreciate every assistance.

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TRIP A3 (4)

DISTRIBUTION OF TRACE FOSSILS PRESERVED IN HIGH ENERGY DEPOSITS OF THE POTSDAM SANDSTONE, CHAMPLAIN, NEW YORK

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ABSTRACT

An unusual bedding plane exposure of the Potsdam Sandstone near Champlain, New York, displays sedimentological and paleontological features that are seldom available for study in this formation. Strong marine currents having three vectors of orientation produced trough cross-strata and asymmetrical ripple marks that are indicative of complex tidal settings associated with am inlet or gut between large sand bodies or barrier bars.

Associated with these high energy deposits are the trace fossils *Diplocraterion*(?) sp. and *Phycodes*(?) sp. which colonized the cross-stratified deposits after deposition. *Diplocraterion*(?) sp., a suspension-feeder may have occupied local patches of bottom in the most active current whereas distribution of *Phycodes*(?) sp., a deposit feeder or predator, was controlled by food content of the sediments. In cross-stratified units there are some indications that numbers of *Phycodes*(?) sp. and *Diplocraterion*(?) sp. were inversely related suggesting that there was habitat competition between them. Theoretical considerations regarding the amount of detrital organic matter that one may expect to have associated with cross-stratified sands also suggest that this species of *Phycodes* may have been a predator of microorganisms rather than simply a detrital deposit feeder.

INTRODUCTION

The region immediately south of Champlain, N.Y., is underlain by white or cream-colored, fine and medium-grained, quartz sandstone that is assigned to the Keesville Member of the Potsdam Formation (Fisher, 1982). It can be viewed in local road cuts along N.Y. Route 9, in some farm fields, and in small abandoned quarries exposing 15 feet, or so, of section. One such quarry behind the Clinton Farm Supply Company, 1.1 miles south from Champlain on highway 9, includes a bedding plane exposure that reveals both biologic and sedimentologic properties of the formation that are seldom seen to this degree (Figure 1). In this paper we describe conditions of environment and biota revealed by this exposure and make comparisons with the Potsdam Sandstone in St. Lawrence County.

METHODS

Study of this exposure proceeded by developing a grid of 1-meter squares over the outcrop to serve as a base for mapping both sedimentary structures and ichnofauna. Vectoral data of cross-strata axes and ripple crests were collected within the grid. The resulting map (Figure 2) is drawn with respect to magnetic north. Trace densities within each square were counted to provide a frame of reference for population densities of the trace makers.

SITE DESCRIPTION AND OBSERVATIONS

A continuous plane exposure forms the surface along the north side of the parking lot in front of the Clinton Farm Supply building. The entire surface has been glaciated as evidenced by presence of chatter marks, striae, and a polished surface on the outcrop. This glacially-eroded surface continues on the east side of Route 9, as well, but it is much less dramatic in both ichnofossils and sedimentary structures there.

To the northwest and west of the exposed bedding plane, the overlying bed, an 18 inch-thick, white quartz arenite can be seen. It too, has a rippled surface in places, but is generally not unusual. To the rear of the CFS building lies a small quarry which has been cut into the units being examined. It does not afford any meaningful exposures because it is presently drowned.

STRATIGRAPHY

This site does not offer much stratigraphic data. Reference sections for "Potsdam" strata in the Champlain Valley lie in the Ausable River valley. The lower portion of the formation is reddish, arkosic, subangular to subround, fine-to medium-grained sandstone. Overlying this, as noted above, is a white, subround-to-round, fine-to medium-grained quartz sandstone, the Keesville Member of the Potsdam. The outcrop in question lies in the medial (?) portion of the member.

Stratigraphic relationships with "Potsdam" rocks to the west are not clear, nor are age relationships within the formation. The best regional biostratigraphic index available seems to be the trace fossil *Climacticnites wilsoni* Logan, 1860. Yochelson and Fedonkin (1993) have recently produced an exhaustive study of the paleobiology and paleobiogeography of this strange trace which occurs in the form of a trail made during feeding. *Climacticnites wilsoni* seems to be restricted to Dresbachian rocks in Missouri and Wisconsin. The same trace fossil was first described from sandstones of northern New York, north of Mooers, and Canada

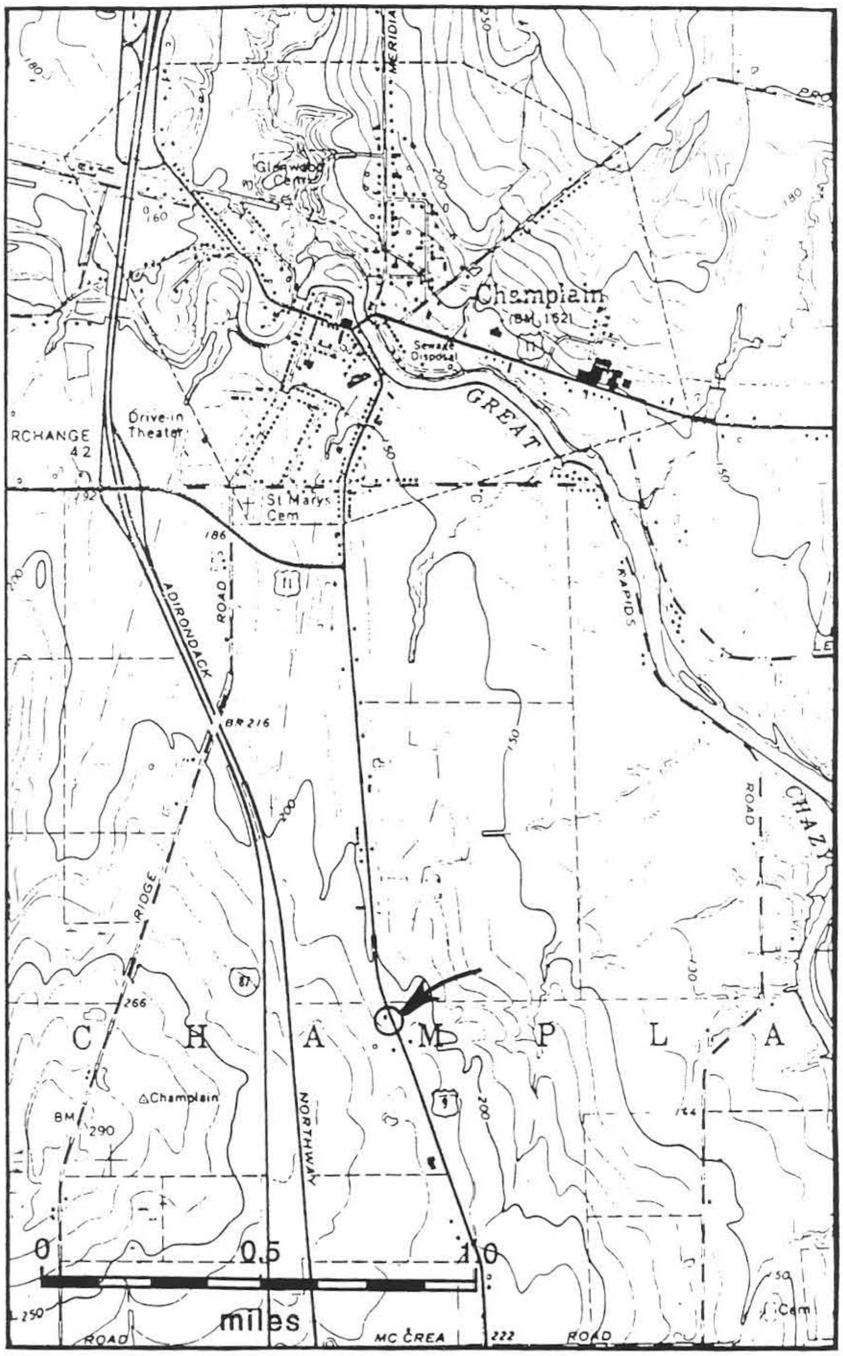

Figure 1: Portion of the Champlain, New York 71/2 minute topographic quadrangle indicating location of the outcrop under discussion herein.

near Perth, Ontario. It is known to occur at Ausable Chasm in the arkosic, Ausable Member and on Wellesley Island where it is found in white, clean, plane-bedded sandstone, but is not known from the type Potsdam nor the surrounding region in St. Lawrence County. In Quebec, Canada, *C. wilsoni* occurs in the Cairnside Formation in the Potsdam Group overlying more than 500 meters of the Covey Hill Formation which in turn rests on Grenvillian basement (Yochelson and Fedonkin, 1993: Hofmann, 1972). Probably the outcrop we describe, with its contained ichnofauna, is post-Dresbachian and pre-Ordovician in age. Local absence of *Climacticnites wilsoni* is most likely due to an absence of rocks representing its preferred intertidal to supratidal flat habitat (Yochelson and Fedonkin, 1993). It would seem that our site was not more than a few meters paleodepth below this. We would welcome discovery of strata containing both *Diplocraterion sp.* and *Climacticnites wilsoni* if such exist.

SEDIMENTARY STRUCTURES

Reference to Figure 2 will permit identification of regions of the outcrop that display large-scale ripple marks and sets of trough cross-strata that have been examined in this study. Cross-cutting relationships record four episodes of large-scale trough migration through a very restricted area. Axial orientations of these sets are designated by episode (Do, D1, D2, D3) on the map in Figure 3. A portion of the complex of sets, D1, D2, and D3 is illustrated by sketch (Figure 4) to point out the three primary axial orientations and their age relationships. Data on axial vectors for 17 axes are presented in Figure 5. Four episodes of trough production resolve into three vector identities that are not the obvious result of a simple, alternating (bi-directional), or tidal, current flow.

Adjacent to the trough cross-strata are examples of large scale, asymmetrical ripples. Although crests of some have been removed by glacial scour, many clearly show steep-sided stoss faces whose direction of migration could be measured perpendicular to the ripple crest. Migration directions again indicate three principal current vectors generally keeping with orientations of cross-strata axes. The stronger tractive current appears to have been associated with ripples designated R1 on Figure 2 and Figure 6 as their wavelength of 38 to 40 cm is almost one-third greater than that of sets R2 and R3. It appears that none were sediment starved.

ICHNOFAUNA

Perhaps the most striking characteristic of the outcrop is the variety of trace fossils preserved in this cross-stratified quartz arenite. The horizon contains a densely burrowed region to the west of the area mapped in Figure 2. That region contains plane bedding which shows *Planolites beverlyensis, Phycodes* sp., *Teichichnus*(?) sp., and possible *Skolithos* sp. Most noteworthy are the ichnotaxa associated with trough cross-strata.

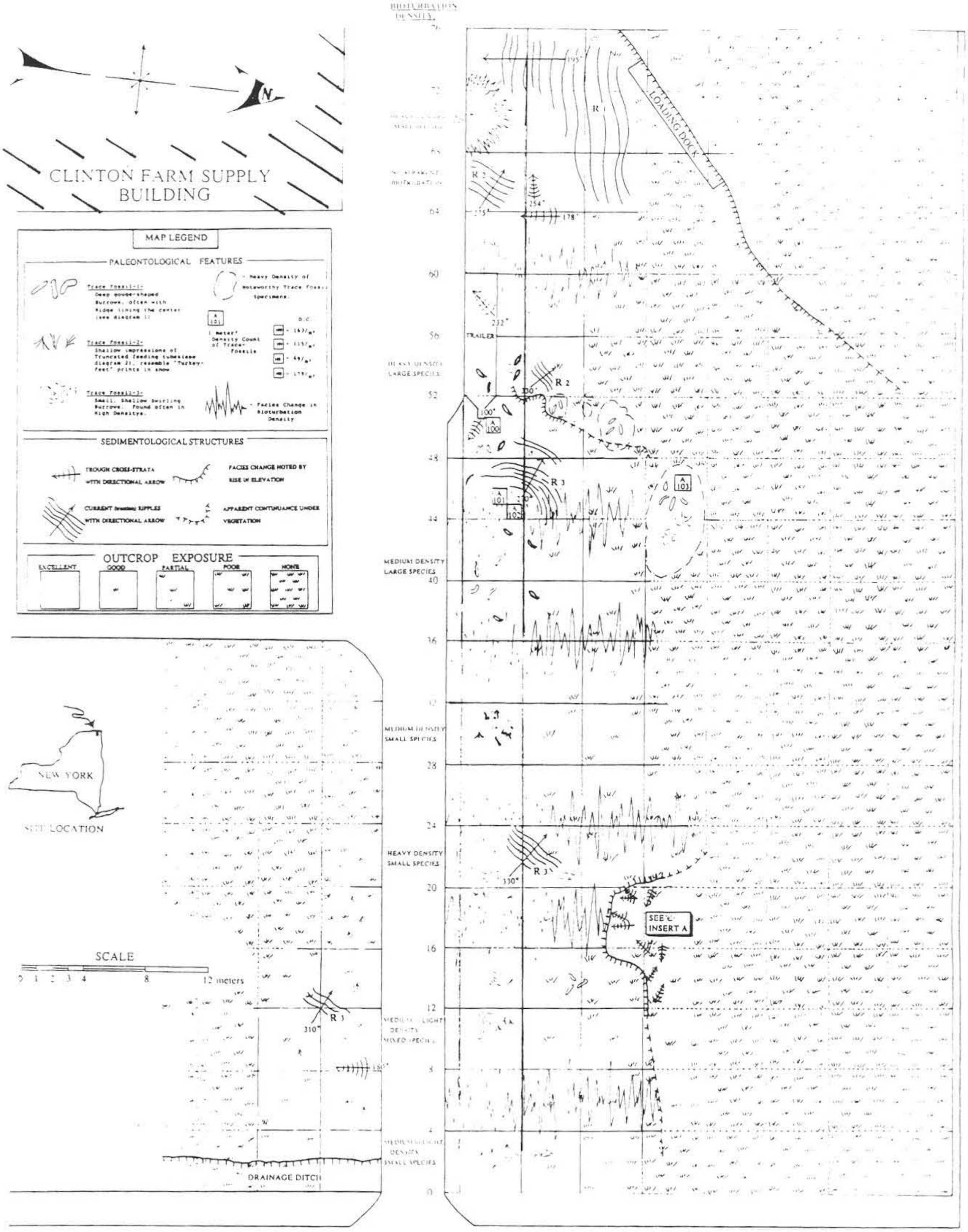

Figure 2: Map of sedimentary structures and ichnofossil distributions on outcrops adjacent to the Clinton Farm Supply building, 1.1 miles south from Champlain, N. Y.

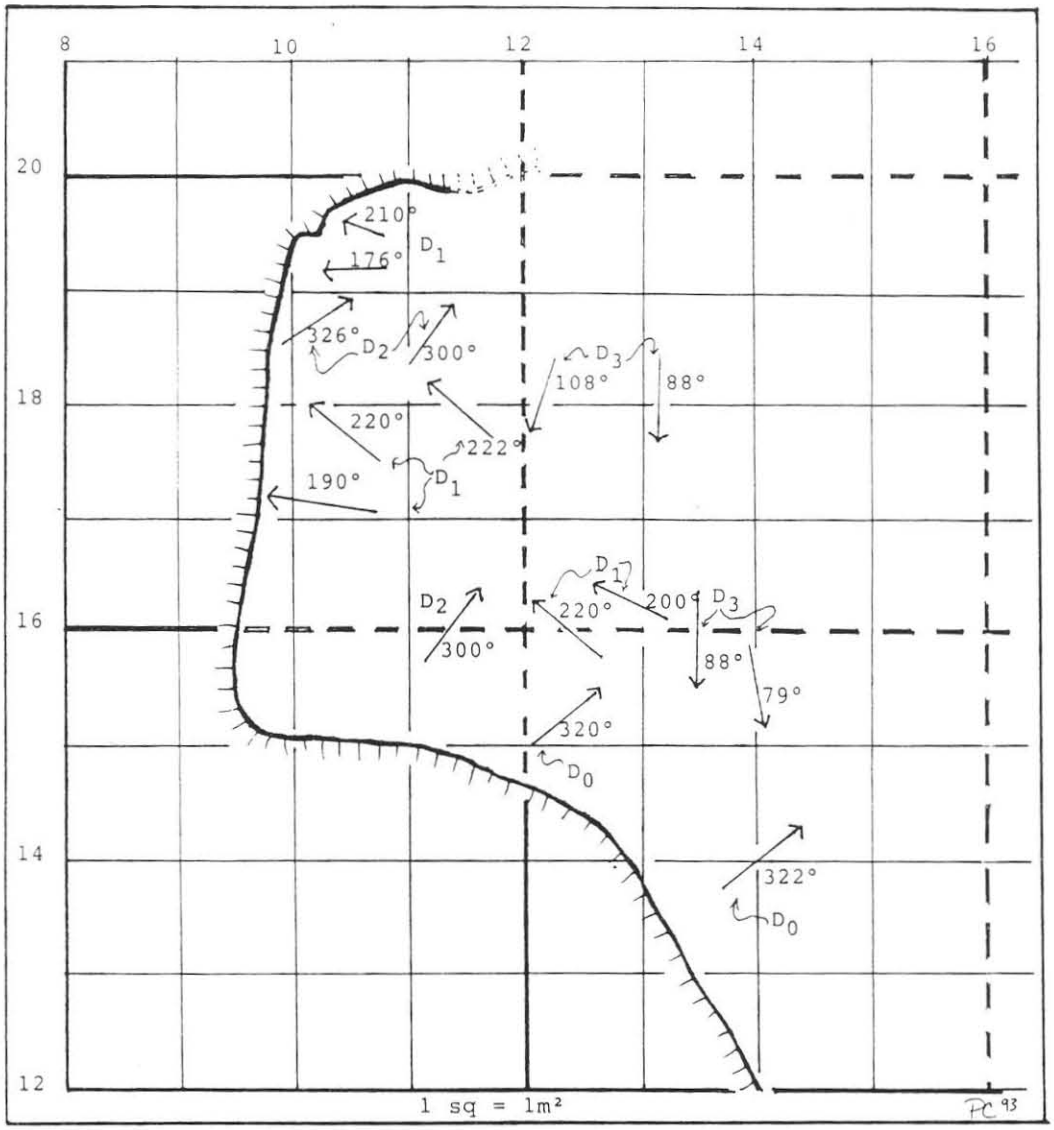

Figure 3: Detail of trough-crossbed axes in area marked "insert A" on Figure 2. Axial orientations are to magnetic north.

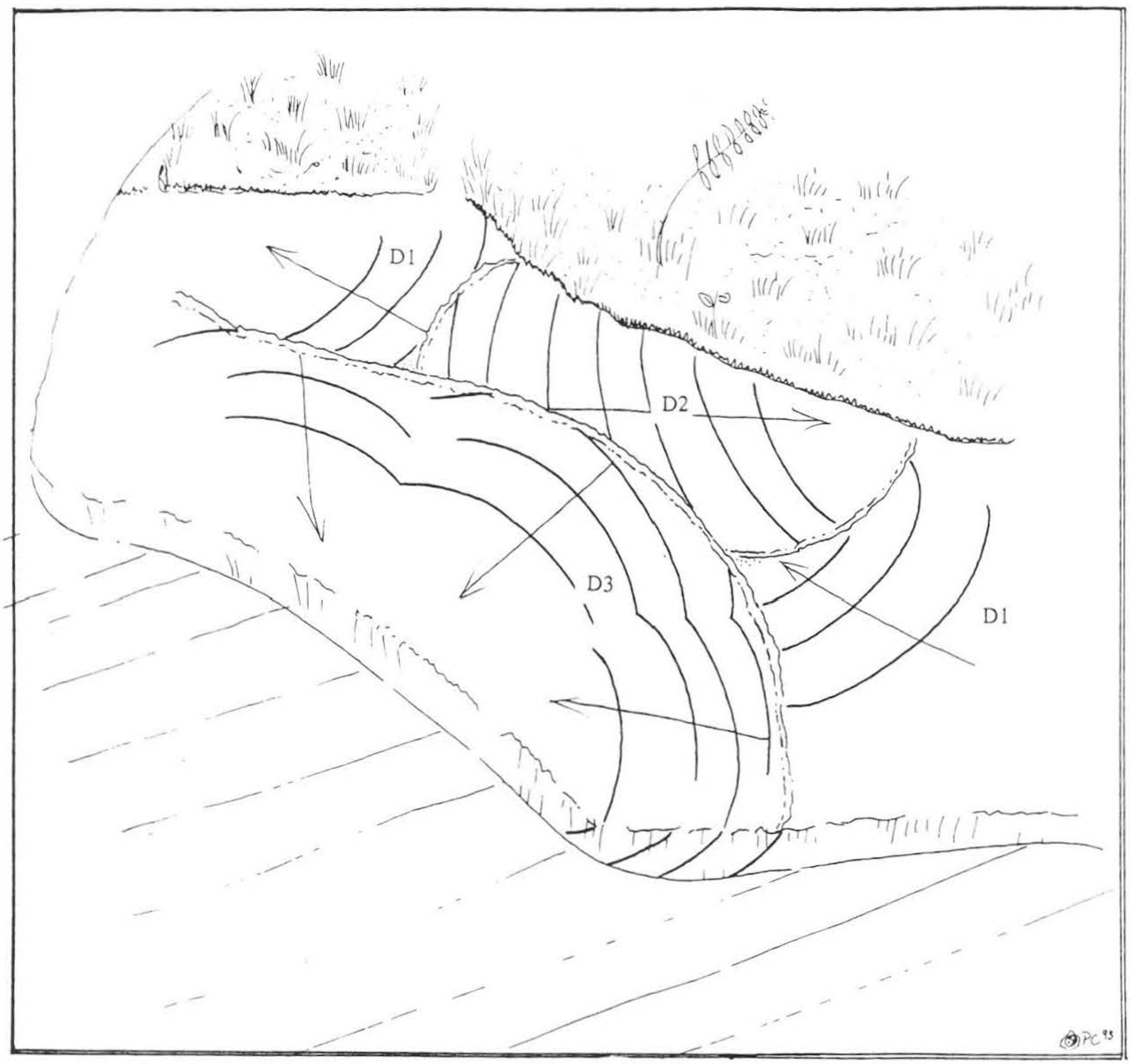

Figure 4: Outcrop sketch of cross-cutting relationships between sets of cross-strata.

Much of the outcrop reveals a gouge, or groove-shaped, trace that appears to be the bottom portion of a u-shaped burrow that has been truncated by glacial scour. Grooves are 1-2 cm wide and 6-8 cm long. Their original height can't be determined. They are widest at the ends reflecting position of the vertical element of the trace (Figure 7). Careful inspection reveals that burrows contain a central ridge interpreted to be spreiten, or a casting, that settled to the floor of the burrow. This relationship was only exposed to the camera in one example (Figure 8) although it was noted many times. In that instance the relationship indicates the body wall of the trace-maker was approximately 4-5 mm in thickness and the gut produced a casting of similar diameter.

The identity of this trace has not been determined with certainty. Evidence suggests that these are the bases of u-shaped tubes of *Diplocraterion* sp. whose appearance on this outcrop is very similar to the "turkey tracks" seen on bedding planes in the Kope Formation of the Cincinnati Arch (Osgood, 1977). Turkey tracks are basal portions of u-shaped tubes assigned to *Diplocraterion biclavatum*, but they are developed in gray shales rather than cross-stratified sandstone. Assignment of these traces to *Diplocraterion*(?) sp. is made with caution.

Population densities of *Diplocraterion*(?) sp. were determined and distributions mapped (Figure 9) across the exposed portions of the outcrops using the grid. Three density levels were established, >115/sq.m., 70-115/sq.m., and <70/sq.m.

Occurring separately from the most dense areas of *Diplocraterion*(?) sp. is another taxon resembling the distal portions of a diminutive species of *Phycodes*. Like the larger tracks these have been truncated by erosion so that they are always seen well below the former sediment-water interface. They appear as empty tubes, 1 to 2 mm in diameter, entering the bed downward at a steep angle and splaying outward from the locus of entry while gradually flattening their angle of penetration to become nearly horizontal. Before splaying, the tubes present a horse-shoe-shaped array of between six and nine openings when crossed by the plane of glacial scour. Definition of the array is lost as tubes change angle of attack toward the horizontal and splay apart.

Diplocraterion(?) sp. and *Phycodes*(?) sp. occur in inverse proportion to each other. Also notably, *Planolites beverlyensis* is absent from the cross-stratified portion of the outcrop as is *Skolithos* sp. Because of the unusual, glacially-eroded view of a high energy depositional regime, this exposure permits some interpretations not easily made from other Potsdam exposures.

INTERPRETATIONS

Outcrop relationships suggest that the local area was an intertidal sandy and muddy flat, somewhat protected, when the flat-bedded, heavily-burrowed *P. beverlyensis* beds west of the CFS building were deposited. Sediments incorporated a richness of organic matter

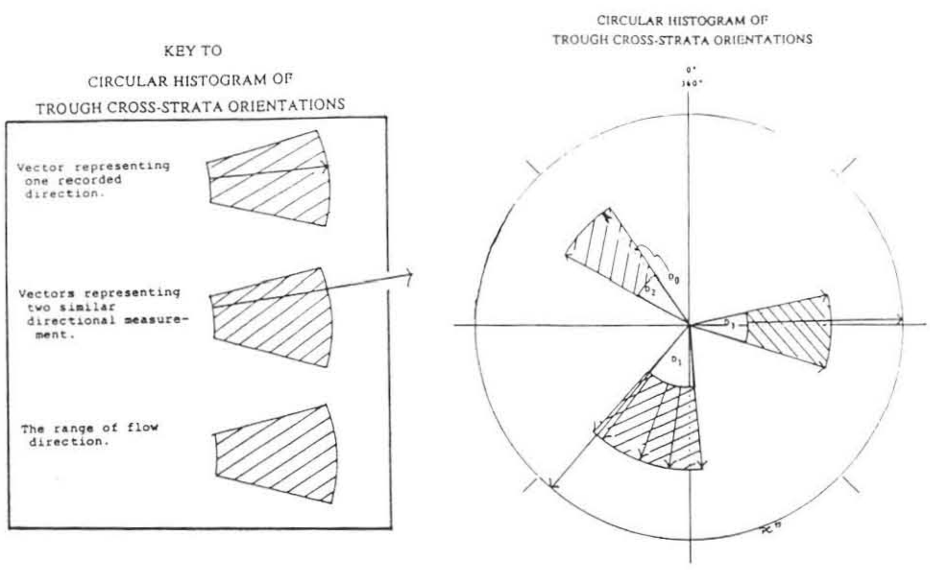

Figure 5: Circular histogram of directional data for axes of trough cross-strata based on magnetic north. N = 14.

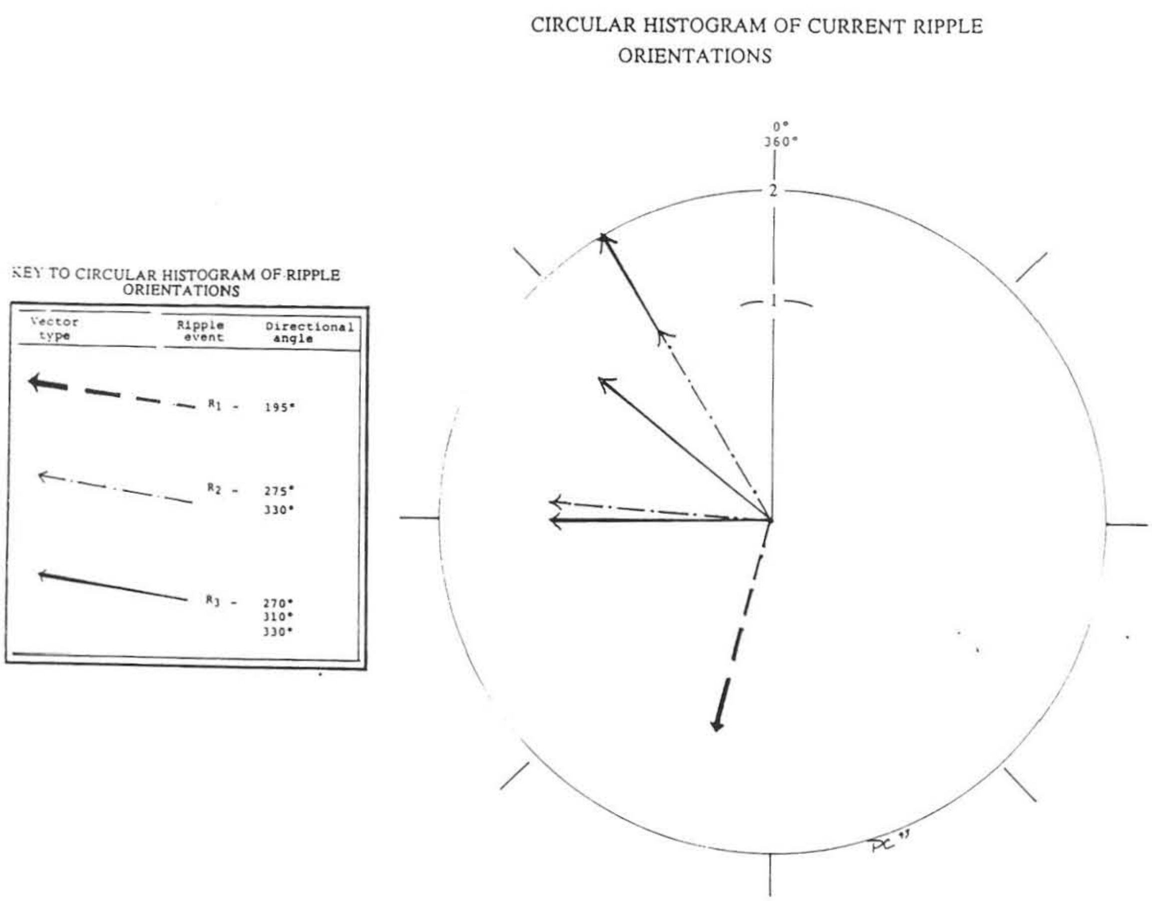

Figure 6: Current directions of asymmetrical ripples based on magnetic north. N = 6.

Figure 7: Bottoms of U-shaped burrows of *Diplocraterion* (?) sp. in grid square A101 (Fig. 2) revealed by glacial erosion of bed. Knife is 9 cm long.

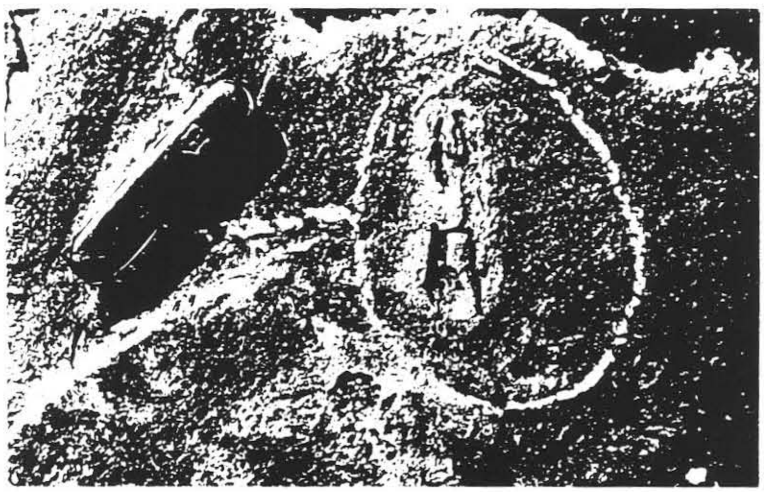

Figure 8: *Diplocraterion* (?) burrow and spreiten remaining in place. Knife is 9 cm long.

Figure 9: View northwestward over part of outcrop at Clinton Farm Supply showing process of laying out 1-meter mapping grids.

exploited by the deposit-feeding burrowers working along preferred horizons. Populations of burrowers were dense. Yochelson and Fedonkin (1993) have suggested that agglutinated foraminiferids may have been numerous in tidal flat sediments. They may have formed some of the food source being utilized here. Other sources for detrital organics must be considered including the possibility that terrestrial organics may have been available for delivery to the tidal flat.

Flat bedded units were dissected by strong currents when the local area was transgressed by a low barrier bar or a spit. Complex current patterns represented by the large-scale three-directional trough cross strata. The complicated current pattern is reminiscent of those developed at the south end of Plum Island on the Massachusetts coast today. There strong tidal flows are occasionally enhanced, or interfered with, by fluvial additions from the estuary behind the barrier island.

In the Potsdam, the strong currents apparently supplied a welcomed habitat for the *Diplocraterion*(?) sp. seen at Champlain. Highest concentrations occur nearest to the most prominent cross-strata. The trace-makers were apparently suspension-feeding organisms which benefitted from current activity. They appear to have been able to burrow deeply enough to prevent being exhumed by the strong currents bringing food.

The *Phycodes*(?) sp. were deposit feeders. They provide some interesting interpretations. It seems they did not prefer to compete with *Diplocraterion*(?) sp. for habitat, perhaps because one simply interfered with the burrowing of the other, making deposit feeding more difficult. Never-the-less presence of the deposit feeder implies presence of organic matter deposited <u>within</u> the cross-stratified sandstones. What form did this food source take? Was it detrital, or may it have been composed of living organisms such as the agglutinated forams mentioned earlier? Obviously an affirmative response would imply that the trace makers were predators, living somewhat like modern scaphopod molluscs, rather than deposit feeders. The narrow diameter of the *Phycodes*(?) sp. burrows suggests that the trace makers themselves were not large organisms, implying that the available food supply was not generous.

Relationships here seem unique in the experience of these authors when compared with the Potsdam Sandstone in St. Lawrence County (Bjerstedt and Erickson, 1989). Those large-scale ripple-cross-stratified sandstones contained *Diplocraterion parallelum* and *Skolithos* sp. without the narrow *Phycodes*(?) sp. seen here. in addition *D. parallelum* in St. Lawrence County rocks is only half the size of the large traces seen in the trough cross-strata at Champlain.

CONCLUSIONS

In conclusion, it appears that conditions of deposition represented by the Clinton Farm Supply outcrop were of unusually high energy when compared with many outcrops of

Potsdam Sandstone that seem to be of the same stratigraphic position to the west. The CFS rocks contain a trace fossil assemblage indicative of this unusual, high-energy, yet food-containing, depositional system. A large species of *Diplocraterion*(?) sp. dominates the suspension-feeder guild whereas a diminutive form of *Phycodes*(?) is the dominant deposit feeder (or predator?) in this habitat-partitioned assemblage.

ACKNOWLEDGMENTS

The authors are grateful to Dr. Richard Lindemann, Skidmore College and to Dr. Ellis Yochelson for calling this outcrop to our attention. We appreciate the permission to study the deposit granted by Mr. and Mrs. Richard Laurin. Lance Erickson assisted with initial observations. Mapping of trace distributions was begun by the St. Lawrence University 1993 Paleoecology class, particularly Myron Getman, Meghan Suanders, Eric Detweiler, Corey Reid, and Tim Montford. The authors thank all the above for their various efforts.

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Paleoecology students developing ideas about trace fossil trophic and habitat relationships at the Clinton Farm Supply outcrop.

(Photos by J.M.Erickson)

TRIP A4

THE LATE GLACIAL ORIGIN OF THE CLINTON COUNTY FLATROCKS

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INTRODUCTION

The Clinton County Flat Rocks comprise a discontinuous, 5-kilometer wide belt of bare sandstone areas that extend approximately 30 km southeastward into the Champlain Valley from Covey Hill, near Hemmingford, Quebec (Figure 1). Created by catastrophic floods from the drainage of glacial Lake Iroquois and younger post-Iroquois proglacial lakes in the St. Lawrence Lowland more than 12,000 years ago (Denny, 1974; Clark and Karrow, 1984; Pair et al., 1988), the exposed sandstone today provides a nutrient-poor, drought-prone habitat that is often covered by jack pine (*Pinus banksiana*) barrens. The relatively low-diversity jack pine community is maintained by fire, which has an important role in ecosystem regeneration.

We will examine the deglacial events leading to the origin of the Flat Rocks by following the path of late glacial meltwater drainage from the divide between the St. Lawrence and Champlain drainage basins. We will also address on-going efforts to understand the linkages between the hydrogeology and ecology of the jack pine barrens and document the recent history of anthropogenic development in the Flat Rocks region. The trip will feature a visit to several sites in the southeastern portion of Altona Flat Rock on property owned by the William H. Miner Agricultural Institute. The area contains the remains of the "Million-Dollar Dam", part of a failed hydroelectric project begun by William Miner in 1910. Additional stops will examine the deposits and landforms that record deglacial events leading to the catastrophic drainage of Lake Iroquois and subsequent formation of the Flat Rocks. The text of this field guide is an expanded version of a previous field guide (Franzi and Adams, 1993) that was concerned exclusively with the geology and ecology of Altona Flat Rock.

GEOLOGICAL SETTING

The Clinton County Flat Rocks lie in the upper reaches of the English, Chazy, and Little Chazy river drainage basins in the northwestern Champlain Lowland, New York (Figure 1). The bare rock areas are entirely underlain by flat-lying Potsdam Sandstone (Cambrian) that ranges from cross-laminated, orange-pink to pale red, very coarse to medium-grained arkose with quartzitic green shale and conglomeratic interbeds to pinkish gray to very pale orange, well sorted, fine to medium-grained quartz sandstone (Fisher, 1968).

The large areas of exposed sandstone were created more than 12,000 years before present by the erosional effects of ice-marginal streams related to catastrophic drainage of glacial Lake Iroquois and younger post-Iroquois lakes (Woodworth, 1905a, 1905b; Coleman, 1937; Denny, 1974; Clark and Karrow, 1984; Pair et al., 1988). Lake Iroquois occupied the Ontario Lowland and drained eastward across an outlet threshold near Rome in the western Mohawk Lowland (Coleman, 1937). Lake Iroquois expanded northeastward into the St. Lawrence Lowland during deglaciation between the Adirondack Uplands to the south and the waning Laurentide Ice Sheet margin to the north. The former water level probably stood at a present elevation between 329 and 332 meters a.s.l. (above sea level) near Covey Hill, Quebec (Figure 1) (Denny, 1974; Clark and Karrow, 1984; Pair et al., 1988).

Northward recession of the ice front into Chateaugay region diverted glacial meltwater westward along the ice margin and created the well-developed Chateaugay Channels (MacClintock and Stewart, 1965). Drainage through the channels emptied sequentially into the northeastwardly expanding Lake Iroquois as ice recession continued. Westward drainage ended when the ice front in the Champlain Lowland receded from the vicinity of the Ellenburg Moraine. Subsequently, eastward drainage of Lake Iroquois began as lower outlets were exhumed along the drainage divide between the Champlain and St. Lawrence drainage systems southwest of Covey Hill. The initial drainage may have occurred through a channel approximately 1 km north of Clinton Mills that was controlled by a threshold between 329 and 332 meters a.s.l. (Clark and Karrow, 1984). The falling levels of proglacial lakes in the St. Lawrence and Ontario lowlands temporarily stabilized at the glacial Lake Frontenac level (Clark and Karrow, 1984; Pair et al., 1988) as the ice margin receded northward and the col at The Gulf (308-311 meters a.s.l.) was uncovered. Outflow from these lakes was directed southeastward along the ice margin where it crossed the English, North Branch and Great Chazy watersheds before eventually emptying into Lake Fort Ann which occupied the Champlain Lowland at an elevation between 225 and 228 meters a.s.l. (Denny, 1974). The outflow streams stripped large areas of their surficial cover and cut deep bedrock channels and plunge pools (e.g. The Gulf (MacClintock and Terasme, 1960) and the Dead Sea (Woodworth, 1905a; Denny, 1974)) (Figure 1) into the Potsdam Sandstone. The most intense scour (e.g. Stafford Rock, Blackman Rock, and Altona Flat Rock) generally occurred on major watershed divides. Cobblestone Hill (Figures 1) is an accumulation of bouldery debris washed from the exposed rock areas by glacial lake outflow floods (Woodworth, 1905a; Denny, 1974).

The scour of the areas southeast of the St. Lawrence-Champlain divide continued as ice recession caused the drainage of Lake Frontenac around the northern flank of Covey Hill. Denny (1974) suggested that the ice margin may have oscillated in the area around Covey Hill causing the lakes in the eastern St. Lawrence Lowland to refill and empty several times. The lake-drainage episodes ended when the ice front receded from the northern flank of Covey Hill for the last time and the proglacial lake in the St. Lawrence Lowland was lowered to the level of Lake Fort Ann in the Champlain Lowland (Pair, et. al., 1988).

ALTONA FLAT ROCK

Physiography

Altona Flat Rock, with an area of approximately 32 km^2, is the largest of the Clinton County Flat Rocks (Figure 1). The exposed rock surface slopes north and east from an elevation of more than 300 meters a.s.l. (above sea level) to below 200 meters a.s.l. where it passes beneath surficial deposits in the Champlain Lowland (Denny, 1974). The sloping surface is broken into a series of stair-like bedrock treads separated by risers that range from a few decimeters to tens of meters in height (Figure 2). The tread surfaces have little local relief except near stream channels and risers. The eroded edges of truncated trough cross-beds, ripple marks, and solution pits are common minor surface features. Shoreline deposits from the highstand of glacial Lake Vermont (Fort Ann Stage) (Chapman, 1937; Denny, 1970, 1974) lap onto the northern and eastern margins of Flat Rock.

The central portion of Altona Flat Rock is drained by Cold Brook, a principal headwater tributary of the Little Chazy River that originates near the Dead Sea (Figure 1). Cold Brook is an underfit stream that occupies a bedrock channel that may locally be more than 200 meters wide and 25 meters deep. The greatest channel incision generally occurs where the stream cuts across prominent southeast-facing bedrock risers. The generally southeastward drainage of Cold Brook is characterized by a subtle rectangular channel pattern that is probably related to bedrock fracture patterns.

Cobblestone Hill forms a conspicuous, elongate ridge on the northern flank of Cold Brook at the southeastern margin of Flat Rock. The ridge is more than 15 meters high, 500 m wide, and 2.5 kilometers long and is composed of angular boulders, almost exclusively Potsdam Sandstone, that range from 0.5 to 3 meters in diameter. The average size of surface boulders decreases to the southeast. Boulder and gravel terraces on the northeast flank of Cobblestone Hill represent beach ridges formed in Lake Vermont (Woodworth, 1905a; Chapman, 1937; Denny, 1974).

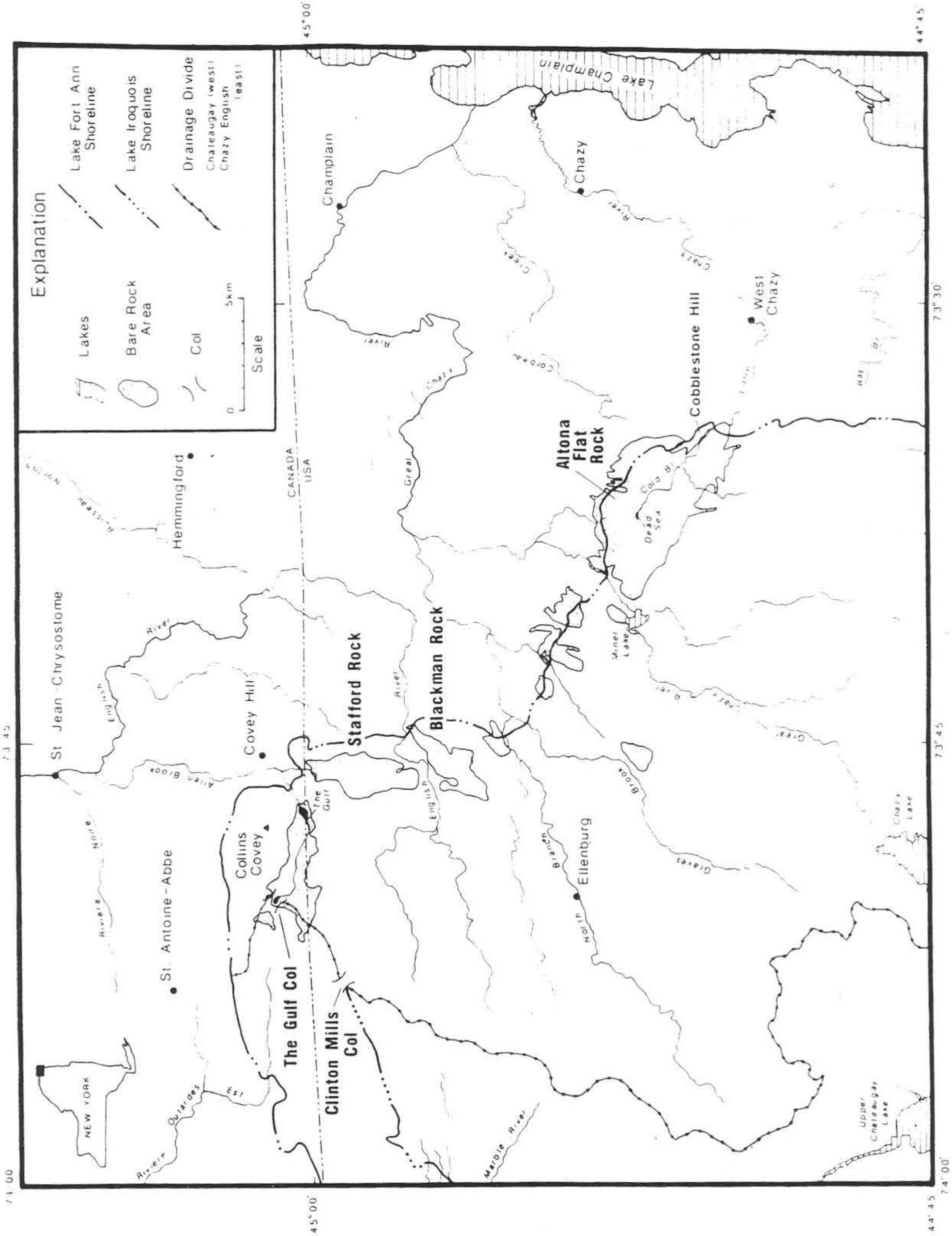

Figure 1. Location map showing the principal bare rock areas east of the divide between the Chateaugay (west) and Chazy and English (east) river watersheds in northeastern New York and adjacent parts of Canada (from Woodworth, 1905a; Denny, 1974; LaSalle, 1985).

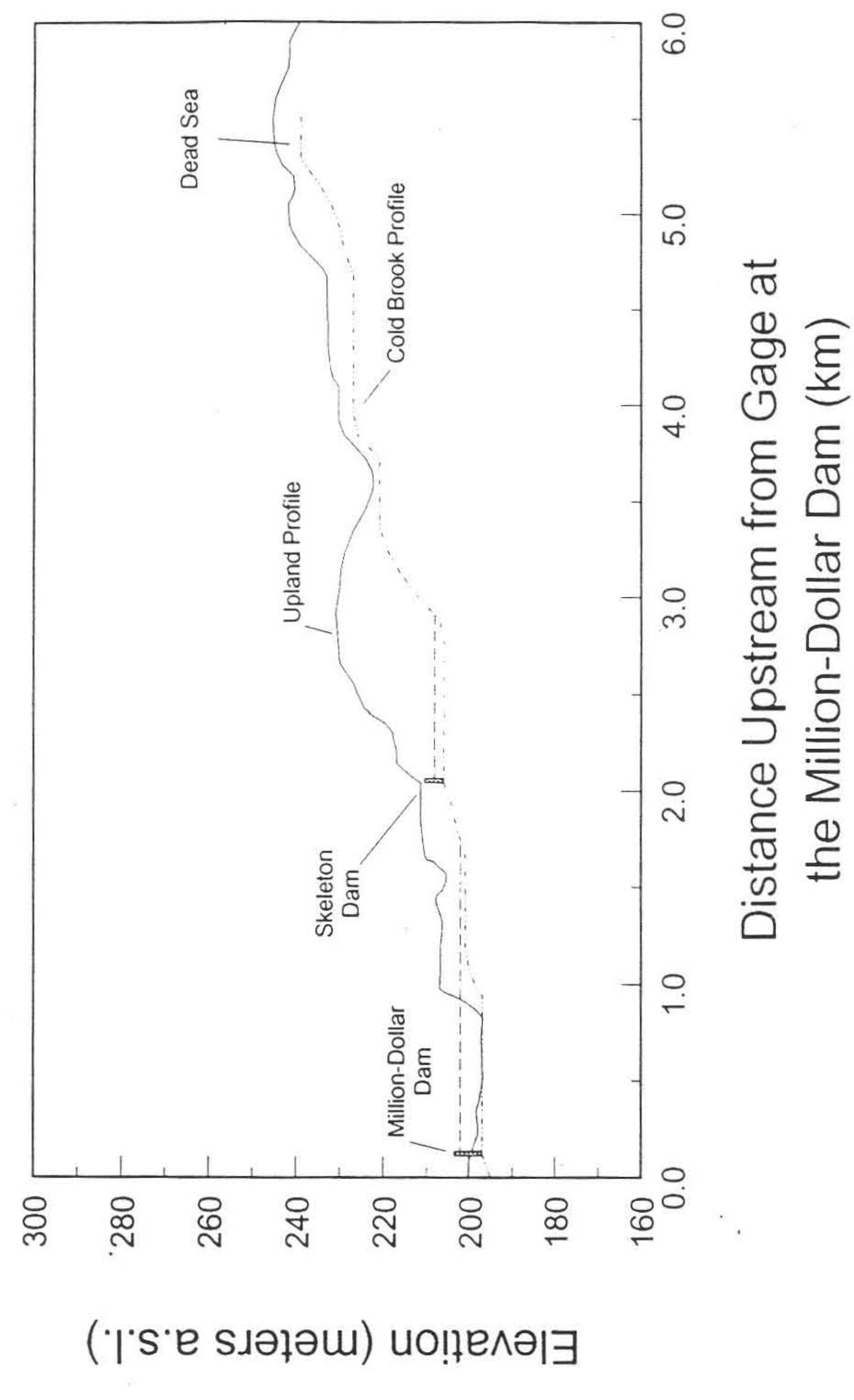

Figure 2. Topographic profile of Cold Brook and adjacent uplands on Altona Flat Rock showing the location of the Million-Dollar and Skeleton Dams and the approximate design pool elevations of their respective reservoirs. The upland profile represents the maximum land surface elevation within 0.5 kilometers of a line oriented N40^{0}W through the Cold Brook Valley.

Figure 3. Photograph of "typical" jack pine barrens on Altona Flat Rock.

Jack pine is a relatively short-lived (<150 years), shade-intolerant, boreal species that has maintained a relic community at Altona Flat Rock because of its adaptations to fire and ability to survive in an area with thin (or absent), nutrient-poor soils (Figure 3). The Altona Flat Rock pine barrens is near the southern limit of the present natural range of jack pine (Burns and Honkala, 1990; Harlow, et al., 1991).

The relatively low species diversity in the barrens reflects low seasonal water availability and the thin, nutrient-poor soils on Flat Rock. The barrens consists essentially of a single tree species, jack pine, with virtually no subcanopy or understory trees. The understory shrubs are predominantly lowbush blueberry (*Vaccinium angustifolium*), black huckleberry (*Gaylussacia baccata*), black chokeberry (*Pyrus melanocarpa*), sweetfern (*Comptonia peregrina*), and sheep laurel (*Kalmia angustifolia*). Ground cover is primarily reindeer lichen (*Cladonia rangiferina*), haircap moss (*Polytrichum commune*), bracken fern (*Pteridium aquilinum*), and *Sphagnum spp*. (Stergas and Adams, 1989).

Jack pine requires periodic crown fires for successful regeneration to occur (Ahlgren and Ahlgren, 1960; Cayford, 1971; Rowe and Scotter, 1973; Cayford and McRae, 1983; Rouse, 1986). Fire releases seeds from serotinous cones stored in the jack pine canopy, prepares a nutrient-rich ash seedbed, and reduces competition for the young seedlings. Since this barrens is a fire-dependent ecosystem, fire exclusion will ultimately cause the local extinction of jack

(*Gaylussacia baccata*), black chokeberry (*Pyrus melanocarpa*), sweetfern (*Comptonia peregrina*), and sheep laurel (*Kalmia angustifolia*). Ground cover is primarily reindeer lichen (*Cladonia rangiferina*), haircap moss (*Polytrichum commune*), bracken fern (*Pteridium aquilinum*), and *Sphagnum spp*. (Stergas and Adams, 1989).

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The combined effects of anomalously high summer air temperature, low seasonal water availability, and flammable foliage produce a fire-prone environment for which the pine barrens community is well adapted. Mean annual precipitation from meteorological records for a 27-year period between July, 1963 to August, 1992 at the Miner Institute in Chazy, New York is approximately 80 cm. Mean monthly air temperature ranges from -11°C in January to 20°C in July (Stergas and Adams, 1989). Summer air temperature in bare rock areas, however, may be as much as 16°C higher than in the surrounding areas, and midday temperatures commonly exceed 38°C (Woehr, 1980). Preliminary data from observation wells on Flat Rock indicate that, in many places, the water table lies well below the depth of root penetration.

There have been four stand-replacing wildfires at Flat Rock during this century (1919, 1940, 1957 and 1965). The oldest jack pine stand at Flat Rock (ca. 73 years) is beginning to show signs of decline. Nearly 40 percent of the trees in this stand (1919 burn area) are dead (Hawver, 1992). The accumulation of dead tree biomass increases the probability of another fire in this stand. A fire management plan, that includes both planned-ignition and natural-ignition fires, is needed for the entire barrens.

The Flat Rock Hydroelectric Project

In the summer of 1910, William Miner, ignoring the advice of his engineers, began construction of a hydroelectric dam and generating station on southeastern margin of Altona Flat Rock (Gooley, 1980). By the time of its completion in March, 1913, the concrete dam, known locally as the "Million-Dollar Dam", had a maximum height of over 10 meters and stretched more than 700 meters across the Cold Brook valley (Figures 4 and 5). The design capacity of the reservoir was more than 3.5 million cubic meters. A second dam, the Skeleton Dam (Gooley, 1980), was constructed upstream to provide supplemental flow to the main impoundment.

The dam and generating station were completed in 1913 but it took almost two years to fill the reservoir to near capacity. The inadequate flow of Cold Brook and ground water seepage through Cobblestone Hill, which formed the eastern flank of the reservoir, proved to be major design flaws. At one point, seepage beneath the dam was so great that it caused severe damage at the Stephen LaPierre residence, approximately 600 meters east of the dam (Gooley, 1980). A 10 to 15 cm layer of concrete grout was spread over more than 100,000 m^2 along the southwestern flank of Cobblestone Hill to mitigate the seepage loss (Figures 4 and 6). A deep trench was excavated at the base of Cobblestone Hill behind the dam for the purpose of pouring a grout curtain to the underlying sandstone and thereby, presumably, sealing the northeastern flank of the reservoir. The grouting effort was partially successful and the power generating plant began operation on January 21, 1915, more than four years from the beginning of the project (Gooley, 1980). The power plant produced electricity intermittently for seven years before mechanical problems forced the abandonment of the project.

SUMMARY

The Clinton County Flat Rocks illustrate the impact of glacial and post-glacial processes on landscape development and contemporary ecosystem-level processes. The region contains a unique record of meltwater drainage related to the retreat of the Laurentide Ice Sheet. The Flat Rocks sandstone pavement, created by erosion associated with late glacial lake-outflow floods, provides an environment characterized by extreme deficiencies in nutrients and soil moisture. Jack pine and its associated heath plants are among the few native species that can survive in this hostile setting. The combined effects of the harsh physical environment and its associated vegetation create an ecosystem that is adapted to and maintained by periodic fire. A fire-management program, based upon a detailed study of ecosystem dynamics and function, is needed if the uniqueness of the Flat Rock jack pine barrens is to be preserved.

ACKNOWLEDGMENTS

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ROAD LOG

The road log begins at the Hudson Hall parking lot on the SUNY Plattsburgh campus, Plattsburgh, New York. Road log distances are presented in English units. All other measurement are in SI units.

Persons using this log in the future should be aware that the Altona Flat Rock field trip stops are located on private property that is owned and patrolled by the William H. Miner Agricultural Institute. A permit must be obtained from the Miner Institute to access this property.

Cumulative mileage	Miles from last point	Route description
Start		Assemble in the Hudson Hall parking lot on the SUNY parking area, turn right at entrance, and proceed northwestward on Broad St.
0.1	0.1	Traffic light, continue northwestward on Broad St.
0.2	0.1	Traffic light, continue northwestward on Broad St.
0.4	0.2	Traffic light at the corner of Broad and Cornelia (US Route 3) streets. Bear left onto Cornelia St. and proceed westward through the next two traffic lights.
1.1	0.7	Junction I-87 North. Turn right onto entrance and proceed northward to Interchange 40 in Beekmantown.
8.1	7.0	Exit ramp at Interchange 40 (Spellman Road). Exit right and proceed to Spellman Road.
8.3	0.2	Intersection of Northway exit ramp and Spellman Road. Turn left and proceed west to Beekmantown Corners.
11.0	2.7	Intersection of Spellman Road and U.S. Route 22. Turn right and proceed north on U.S. Route 22.
14.4	3.4	Intersection of U.S. Route 22, N.Y. Route 348, and West Church Street in West Chazy. Turn left and proceed west on West Church Street.
15.1	0.7	Intersection of West Church Street, Parker Road, and O'Neil Road. Bear left then right to remain on West Church Street.
15.9	0.8	Intersection of West Church Street and Barnaby Road. Turn right and proceed north on Barnaby Road.

16.9	1.0	Barnaby Road changes to a gravel surface at the farm just north of Slosson Road intersection.
17.9	1.0	STOP 1.

STOP 1. LAKE FORT ANN BEACH RIDGES.

Park at the gate at the entrance of the Miner Institute property and continue northward on foot along Barnaby Road approximately 100 meters (320 ft). Turn left into woods and proceed west for 150 to 200 meters (500-750 ft) up the eastern flank of Cobblestone Hill. The beach ridges occur at elevations between 175 and 205 meters (580 and 670 ft) above sea level (Denny, 1974).

The beach ridges on Cobblestone Hill were first described by Woodworth (1905a) and later by Denny (1974). The beaches consist predominantly of moderately rounded to well rounded, pebble to cobble gravel that is deposited in multiple, elongate, low-relief ridges that extend along the northern and eastern flanks of Cobblestone Hill between 175 and 205 meters a.s.l. (Figure 7). Individual deposits are typically as much as 1 meter high and 30 meters wide, and often extend laterally for more than 400 meters (Denny, 1974). The gravel is almost exclusively composed of Potsdam Sandstone that was presumably derived from the alluvial cobble to boulder gravel that composes Cobblestone Hill.

Return to the vehicles at the gate after the discussion at this stop.

17.9		Proceed through the entrance gate. Low roadside excavations approximately 75 meters (250 ft) west of the gate expose the cobble gravel that comprises the Lake Fort Ann beach ridges. Near the crest of the ridge the angular, 0.3 to 1.2 meter (1 to 4 ft) diameter boulders that comprise the core of Cobblestone Hill can be observed at the surface. The remains of the "Million-Dollar Dam" can be seen on the right just southwest of the hill crest.
18.2	0.3	STOP 2.

STOP 2. THE "MILLION-DOLLAR DAM".

The "Million-Dollar Dam" and hydroelectric generation plant was completed on 11 March, 1913 and operated intermittently from 21 January, 1915 until its closure in 1922. A large hole was blasted in the dam shortly after William Miner's death in 1930 to permit Cold Brook to drain freely through the former reservoir. The Flat Rock sandstone pavement is exposed southwest of Cold Brook. The change from mixed deciduous, primarily oak, forest on Cobblestone Hill to jack pine barrens on Flat Rock is characteristically sharp at this location.

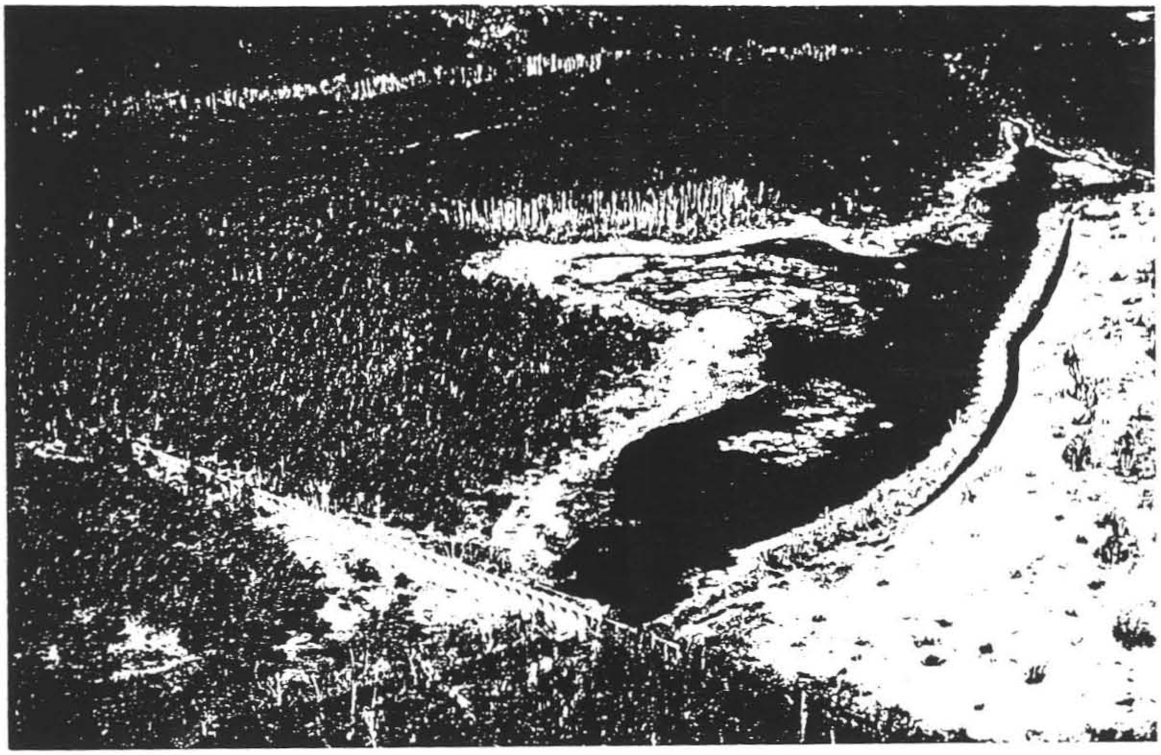

Figure 4. Oblique aerial photograph showing the Million-Dollar Dam and the northwestern flank of Cobblestone Hill (lower right) where it was covered with a concrete veneer to reduce seepage from the former reservoir. The grout-curtain trench can be seen on the right side of the photo.

Figure 5. The Million-Dollar Dam looking northwest from the reservoir outlet.

Figure 6. The Scarpit.

establish an instrumented field station for undergraduate research and instruction in geology and environmental science at the Miner Dam site. A monitoring-well network, consisting of nine wells ranging in depth from 10 to 25 meters, was completed in May, 1992 between the northeastern portion of the former Million-Dollar dam reservoir and the Skeleton Dam (Figure 7). Water-level measurements were begun in late July, 1992 (Figure 8). Future plans include the installation of a weather station, an inflow stream gaging station, and expansion of the monitoring-well network. The field station will provide an important linkage between traditional and applied educational opportunities that addresses some of the unique geological and ecological aspects of the Flat Rock region.

Return to the vechicles following the discussions at this stop and proceed eastward toward Cobblestone Hill. Turn left onto a small road near the crest of the hill that leads northwestward along the flank of the former reservoir.

18.4	0.2	STOP 3.

STOP 3. "THE SCARPIT".

The "scarpit" is the local name given to the desolate landscape created by efforts to grout the porous boulder gravel slope of Cobblestone Hill (Figures 6 and 7). The surface consists

18.4	0.2	STOP 3.

STOP 3. "THE SCARPIT".

The "scarpit" is the local name given to the desolate landscape created by efforts to grout the porous boulder gravel slope of Cobblestone Hill (Figures 6 and 7). The surface consists of a thin (1.2 to 2.5 cm) layer of cement that was poured and raked between large boulders composed predominantly of Potsdam Sandstone. The trench that was dug for the grout curtain (Figure 4) can be observed approximately 100 meters west of the concrete road that parallels the former shoreline of the reservoir.

Return to the vehicles following the discussions at this stop and proceed northwestward on the concrete road.

18.9	0.5	The first of nine observation wells drilled in May 1992 can be observed to the left near the treeline at the edge of the grout surface. The wooded area beyond the well is part of minor southeast-facing bedrock riser. The slope of Cobblestone Hill steepens and the boulder size increases to the northwest.
19.4	0.5	The concrete road ends and the access road bears sharply northeast and continues on the bedrock surface through the jack pine barrens.
19.5	0.1	The road crosses a surface-water supported wetland. The road bed is deeply rutted where it crosses a wetland that contains 0.2 to 1.0 meters of organic soil. Observation wells located approximately 50 meters northeast and 70 meters southwest of the wetland indicate that the water table is usually more than 7.5 meters below the surface.
19.7	0.2	The road crosses a small channel that contains a large wetland. A concrete wall on the left (south) side of the road was constructed to prevent water impounded behind the "Million-Dollar Dam" to escape northward through this channel.
		The access road forks immediately west of the channel. The right fork leads to an abandoned fire tower on the top of Pine Ridge. Bear left and proceed southward.
19.9	0.2	STOP 4.

STOP 4. THE "SKELETON DAM".

The partially completed "Skeleton Dam" was designed to augment flow to the reservoir impounded behind the "Million-Dollar Dam" (Figure 7). The dam impounds "Chasm Lake"

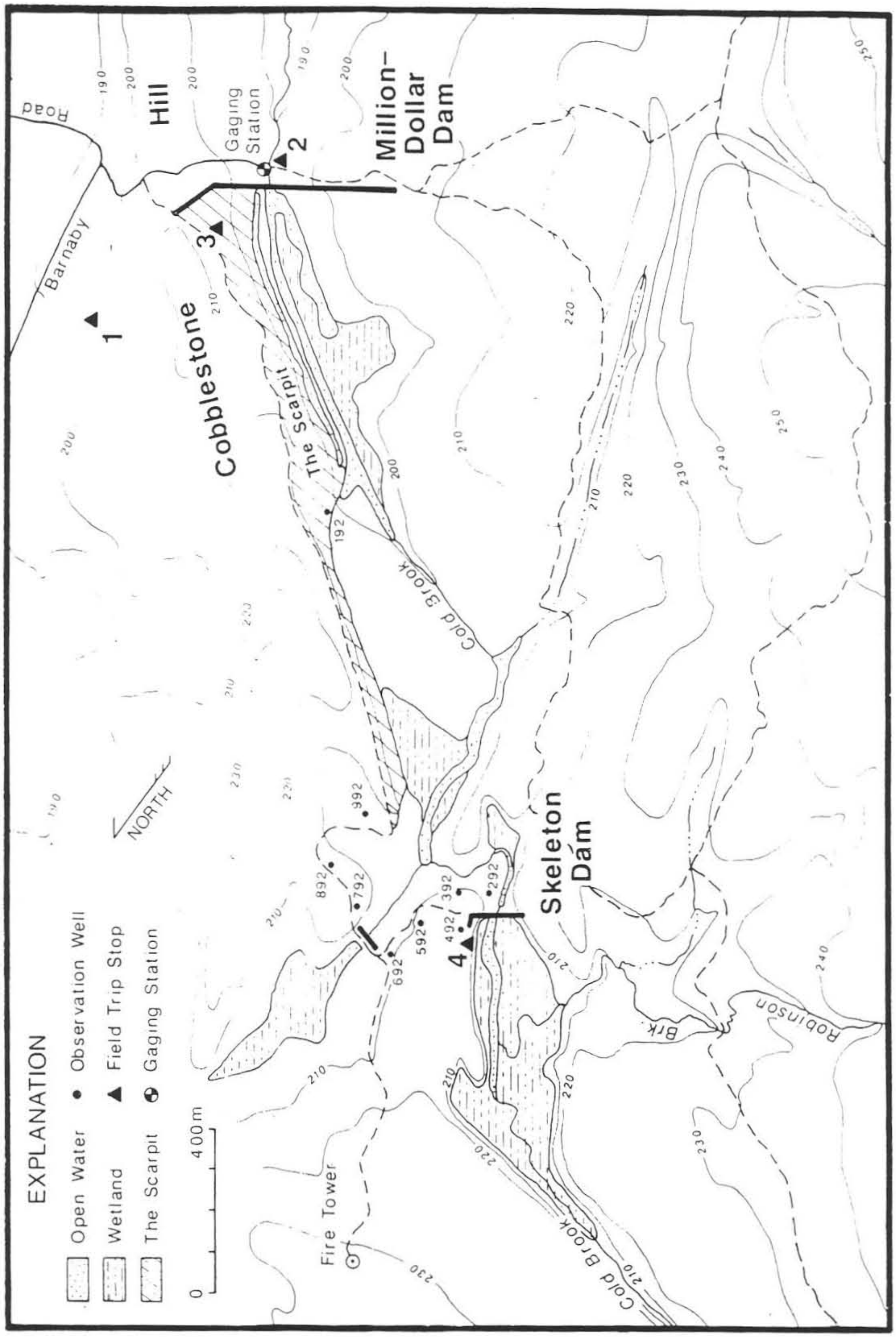

Figure 7. Topographic map of the southeastern portion of Altona Flat Rock showing locations referred to in text. (Topographic base from West Chazy Quadrangle, U.S. Geological Survey 7.5-Minute Series)

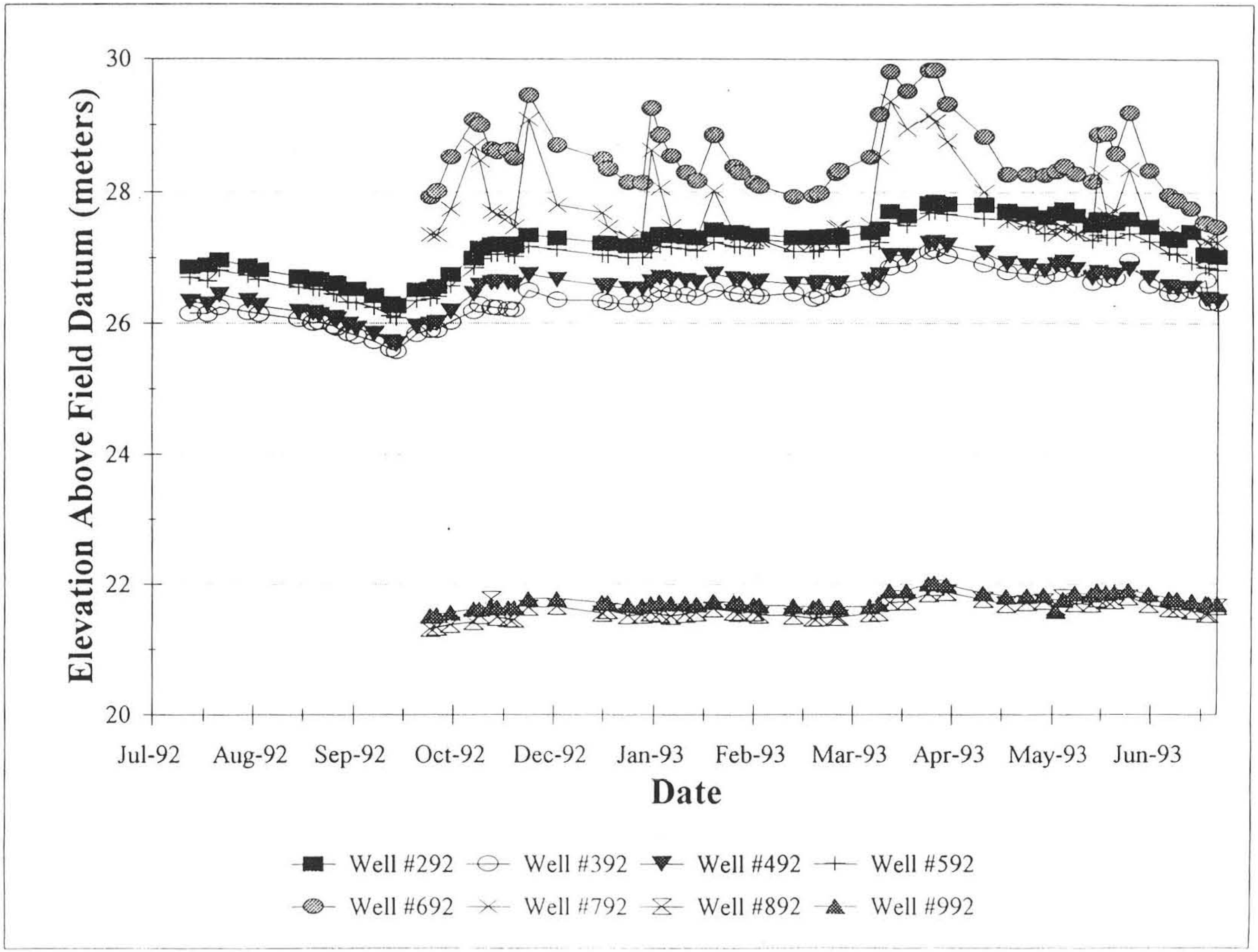

Figure 8. Hydrographs for monitoring wells near the Skeleton Dam.

(Gooley, 1980), presumably named for the deep gorge cut into a prominent sandstone riser at its northwestern edge.

The water level of Chasm Lake dropped more than 2 meters below the spillway of the Skeleton Dam during the summers of 1991 and 1992. What little surface flow reached the basin during the summer months was lost by evaporation and ground-water seepage from the basin. Water level measurements from nearby observation wells since late July, 1992 indicate that steep, eastwardly directed hydraulic gradients exist at the southeastern flank of Chasm Lake, providing support for the hypothesis that some water is being lost from the reservoir by groundwater seepage.

Return to vehicles after the discussion at this stop and follow the rod log in reverse order to the gate at the entrance of the W.H. Miner Institute Property on Barnaby Road.

21.7	1.8	Miner Institute gate on Barnaby Road. Proceed south on Barnaby Road to the West Church Street intersection.
23.7	2.0	Intersection of Barnaby Road and West Church Street. Proceed west on West Church Street.
23.8	0.1	West Church Street forks, bear left (southwest) onto Recore Road.
26.0	2.2	Intersection of Recore Road and Old Military Turnpike. Turn right onto Old Military Turnpike and proceed northwestward toward Ellenburg.
37.1	11.1	Blinking light at the Intersection of Old Military Turnpike and Plank Road. Turn right onto Plank Road and proceed northward.
37.8	0.7	STOP 5.

STOP 5. THE ELLENBURG MORAINE.

The gravel pit at this stop is excavated into the eastern (ice-proximal) side of the moraine. The pit contains approximately 10 to 12 meters of interbedded sand, gravel and diamicton that overlies Potsdam Sandstone. Bedset thickness generally ranges from about a decimeter to just over a meter. The maximum elevation of the upper surface of the moraine at this location ranges between 290 and 297 meters a.s.l..

Return to vehicles after the discussion at this stop and continue northward on Plank Road.

38.8	1.0	Intersection of Plank Road and U.S. Route 11. Turn left onto Route 11 and proceed westward to Ellenburg Depot.
39.2	0.4	Turn left and proceed southward on lake road.

39.7	0.5	STOP 5a.

STOP 5a. THE ELLENBURG MORAINE.

The exposure at this location is on the west (ice-distal) side of the moraine. The exposure contains approximately 10 to 12 meters of interbedded fine to medium sand with minor gravel and silt interbeds. Bedsets range from a centimeter to a few decimeters thick and are commonly horizontally laminated or ripple-cross laminated. Thin silt or silty fine sand deposits occur locally as draped laminae. Ripple azimuths and the gentle dip of the strata indicate a westerly paleocurrent. The moraine deposits were probably deposited in a proglacial lake west of the moraine in the upper North Branch valley. A small sandplain at an elevation of about 290 meters a.s.l. at Ellenburg may represent a delta that was built by the North Branch into the western end of the proglacial lake.

Return to the vehicles after the discussions at this site and return to Ellenburg Depot.

40.2	0.5	Intersection of lake road and U.S. Route 11. Turn left onto Route 11 and proceed westward to Chateaugay.
47.0	6.8	Route 11 crosses the Chazy-Chateaugay drainage divide. The most easterly of the Chateaugay Channels can be seen adjacent to and crossing the road over the next few miles.
55.2	8.2	Intersection of U.S. Route 11 and State Route 374. Turn left and proceed south on Route 374.
56.3	1.1	STOP 7. Turn right into Chateaugay Park.

STOP 7. DISCUSSION OF THE CHATEAUGAY CHANNEL SYSTEM.

Return to the vehicles after the discussions at this stop and continue southward on Route 374.

56.5	0.2	Turn right onto Pulpmill Road.
57.3	0.8	Cross the Chateaugay River and bear left along the river. The Chateaugay channels are on the right.
58.3	1.0	Turn right onto Hartnett Road and proceed west .4 miles.
62.3	4.0	STOP 8

STOP 8. CHATEAUGAY CHANNELS.

This stop is alongside one of a series of underfit stream channels referred to as the Chateaugay Channels by MacClintock and Stewart (1965). These channels are eroded in till and bedrock and trend east-west across the low relief of the south slope of the St. Lawrence Valley. The channels are present from elevations of about 400 m (1300') to 315 m (1040') and are graded to the west.

MacClintock and Stewart (1965), Denny (1974), and Clark and Street (1984) concluded that the channels were eroded along a retreating ice front and emptied into Lake Iroquois to the west. These channels do appear to be graded to the highest level of Lake Iroquois in the vicinity of Malone (Pair and Rodrigues, 1993).

Discussion at this stop will focus on the channel morphology, nature of the ice-marginal drainage, and the possible duration of the drainage events.

Return to vehicles and continue westward on Hartnett Road.

61.3	3.0	Turn right onto Montgomery Road and proceed north.
61.7	0.4	Bear left towards the village of Burke.
62.4	0.7	Village of Burke. Continue north towards Burke Center.
63.6	1.2	Turn left onto Route 11 towards Malone.
71.7	8.1	Cross the Salmon River in downtown Malone.
72.1	0.4	Follow signs to Route 11B. Turn left and proceed south.
72.3	0.2	Bear left (west) on stay on 11B.
88.7	16.4	Village of Dickinson. 11B passes through morainal topography identified as the Fort Covington Moraine by MacClintock and Stewart (1965).
91.9	3.2	Turn left onto Savage Road.
94.1	2.2	Turn right onto Ploof Road.
95.7	1.6	STOP 9

STOP 9. THE ST. REGIS ESKER- FAN COMPLEX AND NICHOLVILLE CHANNELS.

This stop will illustrate the sequence of deglacial landforms characteristic of ice margins in this part of the St. Lawrence Lowland. Moraines and an esker-fan complex in the St. Regis River valley indicate that a lobate ice margin extended southward and ended in a local proglacial lake. The esker ridge attached to the subaqueous fan complex is 80 to 90 feet high and extends northward 2 miles. Esker-fan complexes are present in many of the north-south valleys along the northwestern flank of the Adirondacks. Moraines in the region were attributed by MacClintock and Stewart (1965) to their Fort Covington readvance. They suggested that this ice advance was equivalent to the Port Huron stade. Our studies suggest that the ice margin here is probably recessional in nature and post-dates the Port Huron stade.

Continued ice retreat along the Adirondack flank uncovered the north end of the St. Regis River valley and the water level of the local proglacial lake dropped. Northwest of St. Regis Falls at Nicholville, the ice margin was still grounded and a series of progressively lower ice-marginal channels (from 1050-950') carried drainage westward and emptied into Lake Iroquois. We are standing in the bottom of one of these channels. Subsequently, a northward-flowing stream constructed the Iroquois delta at Nicholville.

Discussion at this stop will center around the time-transgressive nature of ice retreat, the proglacial water bodies in this region, and the morphology of the characteristic landforms. Return to vehicles and continue west on Ploof Road.

95.8	0.1	Turn left (south) onto Fisk Road.
96.0	0.2	Turn left onto Port Kent Road and proceed across the sand plain of the Iroquois delta towards Nicholville.
96.2	0.2	Turn left back onto 11B and proceed west towards Potsdam.
112.8	16.6	Potsdam. Follow signs through town for Route 11.
121.8	9.0	Outskirts of Canton. Best Western/University Inn on left.
122.3	0.5	Romoda Drive entrance to St. Lawrence University.

END OF ROAD LOG

RETHINKING GRENVILLE-AGE DEFORMATION - DUCTILE SHEAR IN GRANITIC GNEISSES OF THE LOWLANDS

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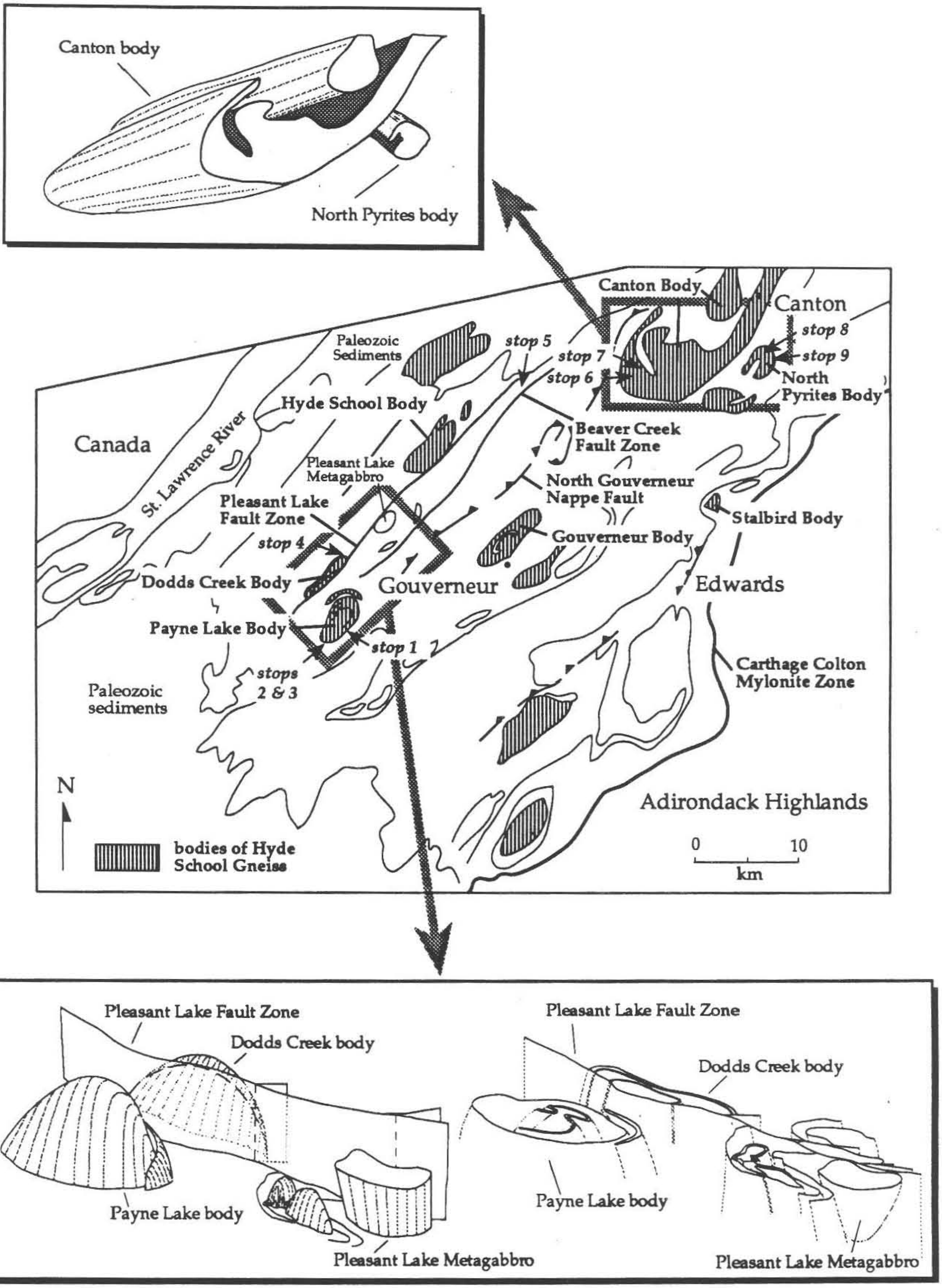
INTRODUCTION

Grenville-age metamorphic rocks exposed in New York State are customarily divided into two regions, the Northwest Lowlands and the Adirondack Highlands (figure 1). The Lowlands are dominated by upper amphibolite to lower granulite facies metasedimentary lithologies; the Highlands are dominated by lower granulite to granulite facies metaigneous units. The Highlands-Lowlands Boundary lies along the Carthage-Colton Mylonite Zone, a major shear zone that juxtaposes the two terranes and exhibits evidence of a complex motion history. The youngest motion along the Carthage-Colton Zone has recently been dated at approximately 1098 Ma by Mezger *et al.* (1992).

Despite the dominance of metasedimentary lithologies in the Lowlands, a number of metaigneous units have figured prominently in various models for the structural history of the Lowlands. Chief among these are the 14 large bodies of leucogranitic gneiss collectively referred to in the recent literature as the Hyde School Gneiss, after exposures in the Hyde School body (Carl *et al.*, 1990, Whitney *et al.*, 1989; McLelland *et al.*, 1992) (figure 1). The Hyde School Gneiss bodies are complex domical structures scattered throughout the Lowlands. U/Pb zircon geochronology indicates that protoliths for the Hyde School Gneiss in a majority of these bodies formed *ca.* 1225-1230 Ma (McLelland *et al.*, 1992).

Traditional views portray the structural evolution of Proterozoic rocks in the Northwest Lowlands as a series of folding events at scales ranging from microscopic to regional as part of a series of contractional orogenic events stretching from Texas to Labrador over a time period from some time between 1300 and 1200 Ma to about 1050 Ma. These models attribute complex patterns in the distribution of lithologic units to the interference of fold wave forms of different ages and orientations. Shearing, when recognized at all, has been



Canton body
North Pyrites body
Canton Body
Canton
stop 8
stop 9
North Pyrites Body
Paleozoic Sediments
stop 5
stop 7
stop 6
Hyde School Body
Beaver Creek Fault Zone
Canada
St. Lawrence River
Pleasant Lake Metagabbro
North Gouverneur Nappe Fault
Pleasant Lake Fault Zone
Stalbird Body
stop 4
Gouverneur Body
Dodds Creek Body
Gouverneur
Edwards
Payne Lake Body
stop 1
stops 2 & 3
Carthage Colton Mylonite Zone
Paleozoic sediments
N
Adirondack Highlands
bodies of Hyde School Gneiss
0
10
km
Pleasant Lake Fault Zone
Dodds Creek body
Payne Lake body
Pleasant Lake Metagabbro
Pleasant Lake Fault Zone
Dodds Creek body
Payne Lake body
Pleasant Lake Metagabbro

relegated to discreet and narrow zones and has not been previously thought of as a major kinematic element in evolution of Lowlands structures.

Over the past several years, I have evolved a very different view of deformation in the Lowlands, one in which regional ductile shear played a major role, and have argued that models for the formation of structures in the Lowlands must be consistent with the manner in which microscopic through macroscopic structures evolve when rocks undergo deformation involving a large component of ductile shear. Evidence for the importance of regional ductile shear has come from examining several bodies of Hyde School Gneiss, as well as a series of younger granitic gneisses in the Lowlands. The introductory paper for this field trip will first compare published models for the evolution of Lowlands structures, then briefly outline an alternative model consistent with structural features of the Hyde School Gneiss, and finally explore the evidence compelling us to think seriously about regional shear.

TWO DIFFERENT MODELS FOR THE EVOLUTION OF MAJOR STRUCTURES IN THE LOWLANDS

The Cross-Folding Model

Many workers over the years have proposed similar models for the evolution of structures in the Lowlands (*e.g.*, Brown, 1989; Foose and Carl, 1977; Whitney *et al.*, 1989; Wiener *et al.*, 1984). Each of these models portrays successive episodes of folding as responsible for reorientation of foliations and lineations. In particular, these models suggest that the quasi-circular, ovoidal, and hook-shaped outcrop patterns of the Hyde School Gneiss (map, figure 1) were formed by repeated co-axial isoclinal folding (producing Ramsay Type III refolds) followed by cross-folding (producing Type I refolds in earlier complex structures). The model is illustrated nicely in 3 dimensions in a figure published by Foose and Carl (1977) portraying a set of F_1 isoclines refolded co-

Figure 1 (previous page). Generalized geologic map of the Northwest Lowlands showing bodies of the Hyde School Gneiss (map modified from Carl *et al.*, 1990). 3-dimensional reconstruction at the top of the page shows the subsurface geometry of the westsouthwest plunging Canton sheath fold and the northwest plunging North Pyrites sheath fold as viewed obliquely from the south. The 3-dimensional reconstructions at the bottom of the page show subsurface and above-ground geometries of the Payne Lake and Dodds Creek sheath folds and several other adjacent sheath-shaped structures as viewed looking obliquely down to the southwest.

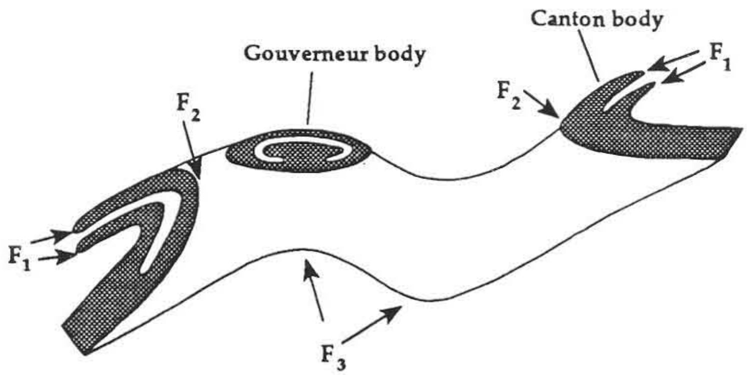


Figure 2. Outcrop patterns and subsurface geometry of the Canton and Gouverneur bodies according to the widely-published cross-folding model for deformation of the Lowlands (from Foose and Carl, 1977).

axially by northeast-trending F_2 isoclines and cross-folded into domes and basins by northwest-trending F_3 folds (figure 2). In this model, what can be visualized as the humps of a Loch Ness Monster breach the surface in the domical culminations of the Hyde School Gneiss bodies of the Lowlands.

While other models suggest three generations of early northeast-trending folds instead of two, all view deformation across the Lowlands as having proceeded largely by folding and re-folding. Shear, when recognized at all, is portrayed as occurring along discreet shear zones such as the North Gouverneur Nappe Fault (Brown, 1989), the "tectonic slides" in the southeastern Lowlands, and other very localized structures. Only Heyn's work (1990) brings large-scale ductile shear into evolution of a major zone in the Lowlands, the Carthage Colton Mylonite Zone that separates the Lowlands from the Highlands.

The Sheath Fold Model

Work by Tewksbury and co-workers (1991, 1992, 1993, in review, and in progress) has shown that shear fabrics are extensively developed, both in the Hyde School Gneiss and in younger granitic gneisses of the Beaver Creek Region. We have argued that the refolding/cross folding model can explain the *geometries* of the Hyde School Gneiss bodies, but that it cannot successfully explain the extensive development of shear fabrics in the Hyde School Gneiss and in other granitic rocks in the Lowlands. Furthermore, we have argued that

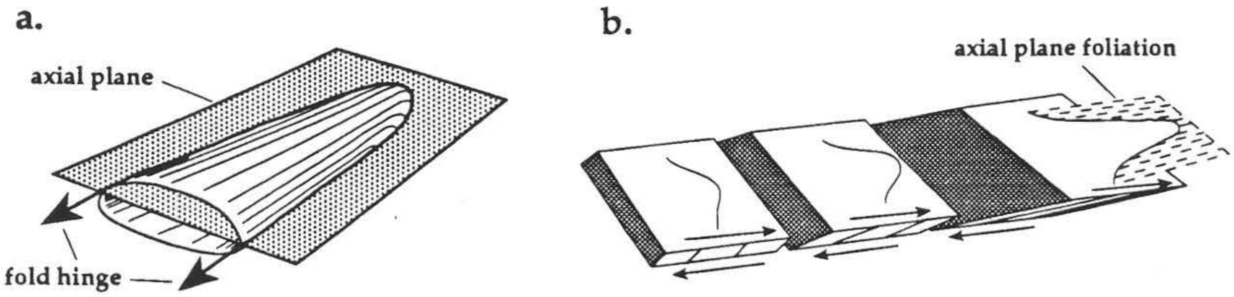


Figure 3. a) The anatomy of a sheath fold, showing the re-curved nature of the fold hinge; b) progressive development of a sheath fold during simple shear by passive amplification of an irregularity that extends across the shear planes (modified from Malavielle, 1987).

any model for the development of large-scale structures in the Lowlands must be consistent with the way that structures evolve when rocks undergo deformation involving a large component of ductile shear at a regional scale.

We have proposed that both the geometries and fabrics of the Hyde School Gneiss are more consistent with the development of sheath folds than with refolding followed by cross folding. A sheath fold is a fold shaped like a sword sheath or the hand of a mitten. Rather than having a straight hinge, as a cylindrical fold does, a sheath fold has a highly recurved hinge (figure 3a). A simple sheath fold has a circular to ovoidal cross section, and map areas dominated by sheath folds show many closed outcrop traces. Sheath fold development does not require folding and orthogonal cross-folding in order to create recurved hinges and ovoidal outcrop patterns. Rather, sheath folds are commonly thought to develop progressively in environments of simple shear by passive amplification of irregularities that extend upward across shear planes (Cobbold and Quinquis, 1980) (figure 3b). A structural irregularity (*e.g.*, a local culmination in an older fold hinge), a bolus of magma, or any other type of irregularity can be streaked out into a long finger by progressive simple shear. The higher the shear strain, the more elongate the sheath. Remarkably complex patterns can be produced by clustered irregularities (Skjernaa, 1989).

Our model contends that the Hyde School Gneiss bodies are sheath folds, produced in an environment of regional ductile shear. Furthermore, we have established the importance of ductile shear in rocks in the Lowlands other than the Hyde School Gneiss. Our model eliminates the D_4 cross folding of many previous authors (*e.g.*, Wiener *et al.*, 1984) (D_3 of Brown, 1989) and suggests instead that circular outcrop patterns and highly recurved hinges developed by sheath folding during their D_2 (or Brown's D_1). Such a model was proposed in passing by Whitney *et al.* (1989), but they cited no supporting evidence. Our work over the last several years provides strong evidence in support of a sheath fold model, and the following sections briefly outline that evidence.

EVIDENCE FOR REGIONAL DUCTILE SHEAR AND THE DEVELOPMENT OF SHEATH FOLDS

Circular or ovoidal outcrop patterns can be produced by dome-and-basin refolds resulting from two successive folding episodes of different orientations. Alternatively, circular or ovoidal outcrop patterns can be produced by sheath folds resulting from one episode of deformation dominated by ductile shear. Distinguishing between the two possibilities requires an understanding of both the 3-dimensional geometry of the structures and the related microfabrics.

Our model is based on evidence collected during detailed studies of fabrics and structures in the Hyde School Gneiss of the Payne Lake and Dodds Creek bodies, during reconnaissance work in several different granitic bodies in the Beaver Creek Region, and during work currently in progress on the Canton, North Pyrites, and Stalbird bodies of the Hyde School Gneiss (HSG) (figure 1).

Geometries of Macroscopic and Mesoscopic Structures

The first piece of evidence in favor of a sheath fold model for the formation of bodies of the HSG is that the ones we have studied are sheath-shaped. In the Payne Lake body, both the mass of HSG and the main regional foliation define an elongate domical structure with subvertical margins (figure 1). The Dodds Creek body was undoubtedly similar in geometry but was truncated nearly down the middle by slip along the late, brittle Pleasant Lake Fault Zone. It is now a steep-sided half dome. Interlayers of metasediment within both bodies mimic the sheath shape of the outer contacts and reveal a complex double "tip" in the sheath shape of the Payne Lake body (figure 1).

While both the Payne Lake and Dodds Creek bodies are nearly vertical sheaths, the Canton, North Pyrites, and Stalbird bodies have shapes consistent with interpretation as shallowly-plunging sheaths. Figure 1 shows the geometry of the complex, westsouthwest-plunging Canton body and the simpler northwest-plunging North Pyrites body. Reconnaissance work on the Stalbird body has revealed intriguing closed foliation patterns within the body itself, suggesting that the Stalbird body may consist of a series of "glove fingers".

If the HSG bodies did, indeed, form as sheath folds, one might expect to find evidence for other sheath-shaped structures in nearby areas in rocks other than the Hyde School Gneiss. Such structures do exist immediately north of the Payne Lake body and east of the Pleasant Lake Fault Zone in an area mapped by Lewis (1969). Here, upward closing sheath-shaped structures occur in metasedimentary gneisses and marbles, and a downward closing sheath-shaped structure occurs in the Pleasant Lake Metagabbro. Figure 1 shows how similar in geometry these structures are to the Payne Lake and Dodds Creek bodies.

One might also hope to find mesoscopic sheath folds in the region. Rare mesoscopic sheath folds in metasediments south of the Payne Lake body (stop 3) have long axes parallel to the Payne Lake sheath axis and provide additional evidence for the presence of macroscopic sheath folds. Rare mesoscopic sheath folds also exist in metasediments adjacent to mylonitized granitic gneiss above what Brown (1989) has termed the North Gouverneur Nappe Fault.

Microfabrics in the Hyde School Gneiss

One of the primary results of our work has been the discovery that what has been referred to for many years as "the main regional foliation" has the characteristics of a ductile shear fabric in many portions of the Hyde School Gneiss in all of the bodies we have examined. Features include porphyroclasts with core-and-mantle structure and asymmetric tails (sigma grains and rare delta grains), abundant matrix material consisting of dynamically-recrystallized grains, quartz ribbons (some many centimeters long), rare mica fish, and rare grain shape preferred orientation in quartz.

A second important result of our work is that the Hyde School Gneiss shows evidence of multiple shear fabrics. In the Payne Lake and Dodds Creek bodies, one shear fabric (the main regional foliation S_m) wraps around the margins of the bodies and defines the shapes of the sheaths (figures 1 and 4a). A second shear fabric (S_{d2}) lies parallel to the axial planes of the sheaths (figure 4b). This younger fabric lies parallel to S_m on the sheath fold limbs and cuts discordantly across both the main foliation and the margins of the body at the northeastern and southeastern hinge regions and in the cores of the bodies. Both S_m and S_{d1} in the Payne Lake body show east side up sense of shear. In the Dodds Creek body, S_m shows clear east side up sense of shear. Sparse data on S_{d1} in the

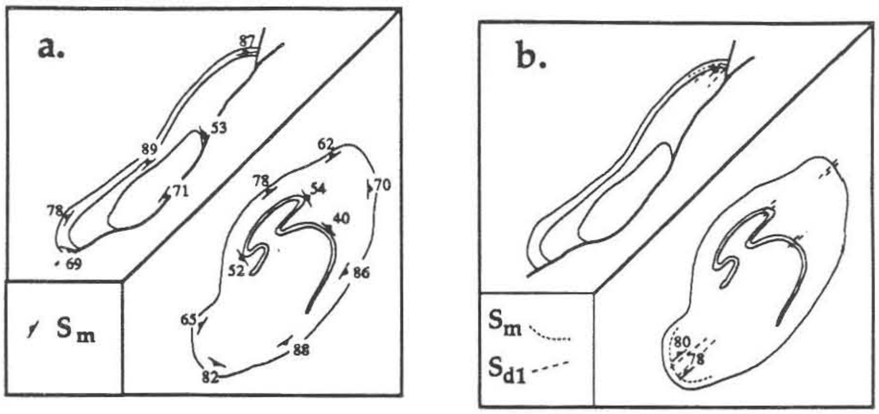
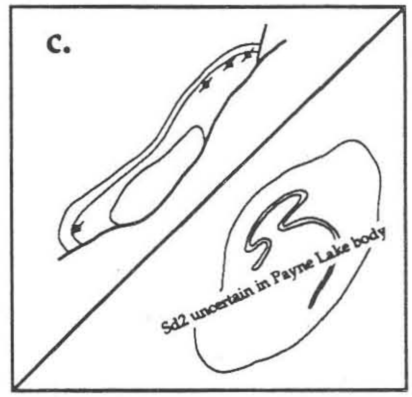


Figure 4. Attitudes of foliations in the Dodds Creek (upper left) and Payne Lake (lower right) bodies. 2a) Attitudes of the main shear fabric, S_m. 2b) Attitudes of the main shear fabric, S_m, and the older discordant shear fabric, S_{d1}. 2c) Attitudes of the younger discordant shear fabric (S_{d2}) from discreet ductile shear zones in the Dodds Creek body.



Dodds Creek body suggest east side up sense of shear as well, but data are too limited for us to confidently confirm the shear sense. Preliminary work in the Canton and North Pyrites bodies suggest that S_m and S_{d1} are clearly present at least in the North Pyrites body as well, with similar concordant and discordant relationships to the body as a whole.

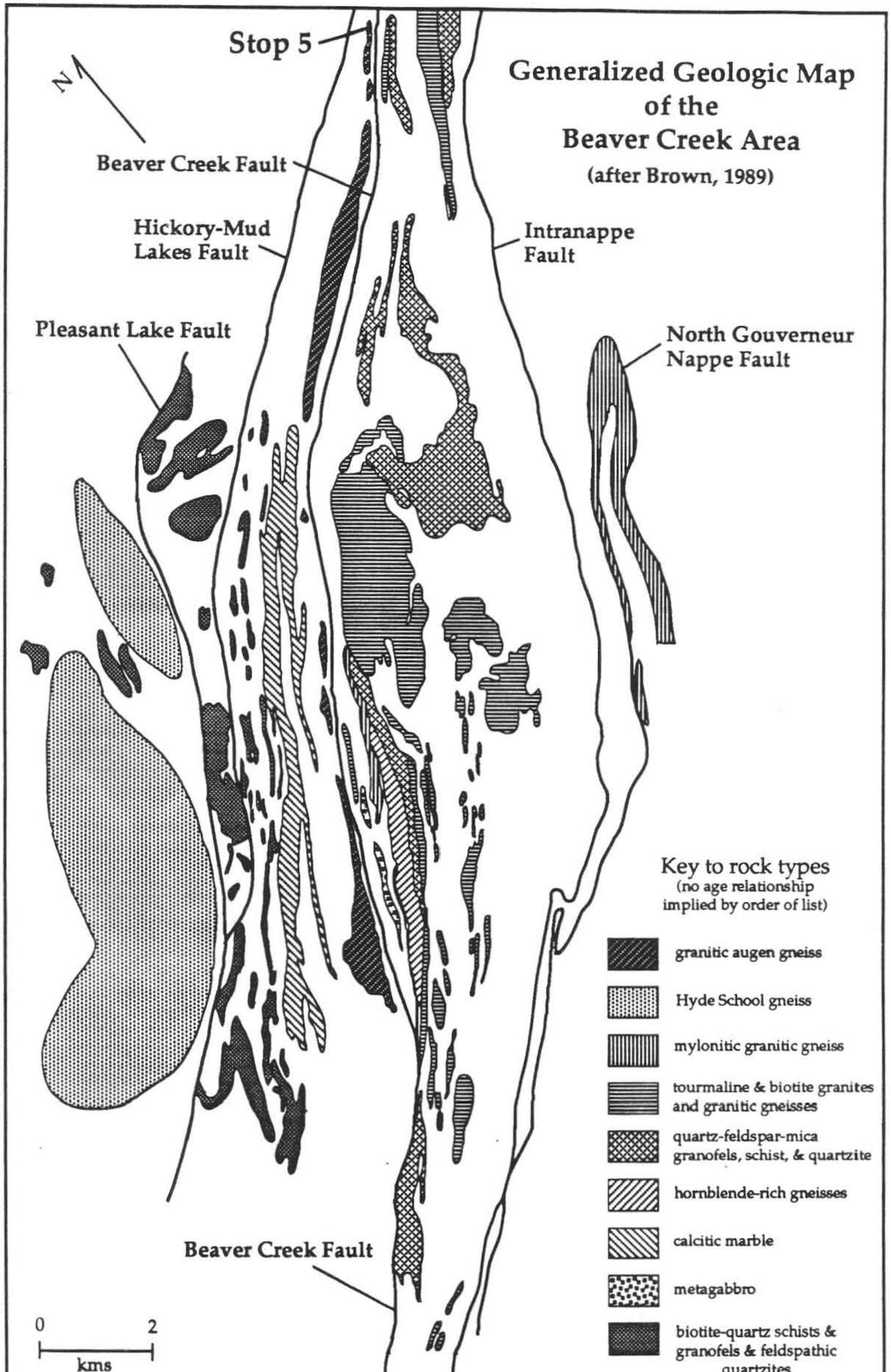
We have argued that the S_m and S_{d1} shear fabrics developed progressively in an environment of regional ductile shear (Tewksbury *et al.*, 1991; Tewksbury and Kirby, 1992). We have proposed early development of the main regional foliation in a deep, subhorizontal regional shear zone in an an environment of plate collision. As shearing continued, irregularities within the shear zone (formed by intrusion of granitic magmas that eventually became the Hyde School Gneiss??) amplified into sheath folds. The early shear fabric (S_m) and related lineations were reoriented as sheath folds grew, and a younger shear fabric (S_{d1}) developed parallel to the axial surfaces of the developing sheaths and with the same sense of shear as S_m. This discordant fabric formed parallel to S_m along the limbs of the sheaths but cut across S_m at the hinge regions.

Shear Fabrics in Other Rocks in the Lowlands

If the Hyde School Gneiss bodies were shaped by regional ductile shear, it stands to reason that one might also expect to find evidence for ductile shear in rocks other than the Hyde School Gneiss. We examined granitic gneisses and metasedimentary rocks immediately east and west of the Beaver Creek Fault Zone and immediately beneath the North Gouverneur Nappe Fault, including the following map units shown in figure 5: granitic augen gneiss, mylonitic granitic gneiss, tourmaline and biotite granites and granitic gneisses, calcitic marble, and quartz-feldspar-mica granofels, schist and quartzite. We chose to extend our investigation into the Beaver Creek region for several reasons. First, Brown (1989) reported both well-lineated granitic gneisses and mylonitic granitic gneisses in several portions of the region. Second, outcrop patterns of metasedimentary units lying in the area immediately west of the Beaver Creek Fault are attenuated in comparison to outcrop patterns of identical units farther northwest (figure 5). We suspected that shear along a major zone parallel to the Beaver Creek Fault Zone might have been responsible for the "shredded" character of the units. Third, east of the Beaver Creek Fault Zone, outcrop patterns of granitic gneisses and metasediments form a large sigmoid, with a smaller sigmoid containing mylonitic granitic gneisses "riding piggyback" to the west (figure 5). These sigmoids bear a striking similarity in morphology to the sigma grains seen in thin sections of ductily sheared rocks. We suspected that asymmetry of attenuated units in the "tails" might reflect major dextral shear in the region.

We made four important discoveries in the Beaver Creek region. First, the "main regional foliation" in many portions of the granitic gneisses is a well-lineated, strongly annealed fabric showing considerable evidence for ductile shear. This suggests that an early event of shear across the entire Beaver Creek region was responsible for creating the "main regional foliation", a conclusion consistent with similar findings in the Payne Lake and Dodds Creek bodies.

Second, while the general importance of early regional shear appears to be clear, the correlations between shear directions in the Payne Lake/Dodds Creek region and those in the Beaver Creek region are not. Stretching lineations and structural geometries in the Payne Lake/Dodds Creek region suggest dominantly vertical, east-side-up shear. Shear directions in the Beaver Creek region are more variable but are clearly not dominantly vertical. In fact, shallowly- to moderately-plunging shear directions are the rule, even where the main foliation dips steeply. Highly variable stretching lineation


Generalized Geologic Map
of the
Beaver Creek Area
(after Brown, 1989)
Stop 5
N
Beaver Creek Fault
Hickory-Mud
Lakes Fault
Pleasant Lake Fault
Intranappe
Fault
North Gouverneur
Nappe Fault
Beaver Creek Fault
0
2
kms
Key to rock types
(no age relationship
implied by order of list)
granitic augen gneiss
Hyde School gneiss
mylonitic granitic gneiss
tourmaline & biotite granites and granitic gneisses
quartz-feldspar-mica granofels, schist, & quartzite
hornblende-rich gneisses
calcitic marble
metagabbro
biotite-quartz schists & granofels & feldspathic quartzites

orientations in the Beaver Creek region suggest re-orientation of both shear fabrics and stretching lineations by subsequent deformation. Such a conclusion is supported by our observation that, while foliation and lineation orientation varies with position within the "sigmoid" east of the Beaver Creek Fault, no correlation appears to exist between fabric development and position in the sigmoid. This suggests that whatever produced the sigmoid occurred after development of the main regional foliation and was not accompanied by a pervasive fabric-forming event. We suggest that early, intense, pervasive regional shear may have been followed in the Beaver Creek region by more "compartmentalized" shear, with localized, complex folding between zones of high shear strain, producing multiple shear fabrics in some rocks, but not in others. The asymmetry of the sigmoid and the pattern of re-orientation of lineations suggests that the region may have been dominated by late dextral shear across wide, steeply-dipping, northeast-striking zones.

Third, Brown (1989) used the geometries of folds to suggest that the North Gouverneur Nappe was emplaced by southeast-directed transport. Our study of kinematic indicators in granitic gneisses beneath the North Gouverneur Nappe Fault and megacrystic hornblende gneisses above the Fault indicates northwest-directed transport.

Fourth, metasediments in the Beaver Creek Region show remarkably poor preservation of ductile shear fabrics. Strongly-lineated and sheared granitic gneisses lie within centimeters of featureless metasediments. We believe that, while substantial annealing in the granitic gneisses eliminated much of the fine-grained, dynamically-recrystallized material, quartz ribbons and porphyroclasts survived. Many of the metasedimentary lithologies did not have appropriate mineralogies to develop quartz ribbons and asymmetric porphyroclasts, and annealing completely wiped out whatever ductile shear features were produced in the rocks. Rare lineated quartzites, quartz-clot marbles, and megacrystic igneous interlayers are the only lithologies we found that preserve shear fabrics. This has implications for studying regional shear in the Lowlands and suggests that the absence of shear fabrics in metasedimentary rocks does not necessarily imply the absence of ductile shear.

Figure 5 (previous page). Generalized geologic map of the Beaver Creek Region (after Brown, 1989).

Evidence for Protracted Shear and Changes in Shear Sense

All of the granitic gneisses we have examined have shown evidence of multiple shear fabrics suggesting protracted ductile shear. Most of the regions we examine also showed complexities in both shear sense and character of deformation. In both the Dodds Creek and Beaver Creek areas, late ductile shear had a sense of slip opposite to that of earlier ductile shear, and, in the Dodds Creek body, late ductile shear occurred along discreet zones rather than pervasively throughout the body. In addition, all of the regions show a progression from ductile to brittle shear conditions. Limited evidence from cataclasites in the Dodds Creek body adjacent to the Pleasant Lake Fault Zone suggest that late brittle faulting may have involved a large component of strike slip. Late reversals in shear sense, compartmentalized shear, and the transition from ductile to brittle conditions may all be part of a complex system of extensional unroofing of the Lowlands region, a process that has also been suggested by a number of people for evolution of the Carthage Colton Mylonite Zone.

The most important implication of complex shear fabrics in the Lowlands is that ductile shear was an important regional aspect of Lowlands deformation and is not something to be viewed as limited in importance either spatially or temporally. Models for evolution of both major and minor structures must be consistent with this observation, and we believe that a sheath fold origin for the bodies of Hyde School Gneiss, as well as other quasi-circular structures in the Lowlands, is the most reasonable one.

MAJOR QUESTIONS FOR THE LOWLANDS AS A WHOLE

The sheath fold model raises a number of interesting questions for the Lowlands as a whole. First, quartz ribbon lineations and the orientations of long axes of possible sheath folds vary considerably across the Lowlands. Both are vertical at Dodds Creek and Payne Lake. Lineations plunge shallowly westsouthwest in the Canton body and shallowly northwest in the North Pyrites body. Stretching lineations plunge north, northwest, and northeast in granitic gneisses of the Beaver Creek region. Why the variation? We also do not yet know the pattern of shear sense across the entire region.

There is also the question of timing of shear. U/Pb zircon geochronology indicates that the Hyde School Gneiss has an average age of 1225-1230 Ma

(McLelland *et al.*, 1992). Granitic gneisses from the Beaver Creek region are part of a younger series of granitic rocks intruded between 1150 and 1170 Ma. The oldest fabric in each is a shear fabric. Is it the *same* shear fabric? If it is, shearing clearly must postdate intrusion of the younger granites, long after formation of the Hyde School Gneiss protolith. On the other hand, both Tewksbury (1992) and McLelland (in press) have postulated that the Hyde School Gneiss was intruded into a major regional shear zone, providing a nice irregularity to be amplified into large sheath folds and localizing major shear by melt-enhanced deformation. If this were true, then the two main shear fabrics (S_m and S_{d1}) in the Hyde School Gneiss must have formed before the Beaver Creek gneisses were intruded. This means, of course, that the oldest shear fabric in the Beaver Creek gneisses formed by a later shearing event in the Lowlands, complicating still further the picture of regional shear.

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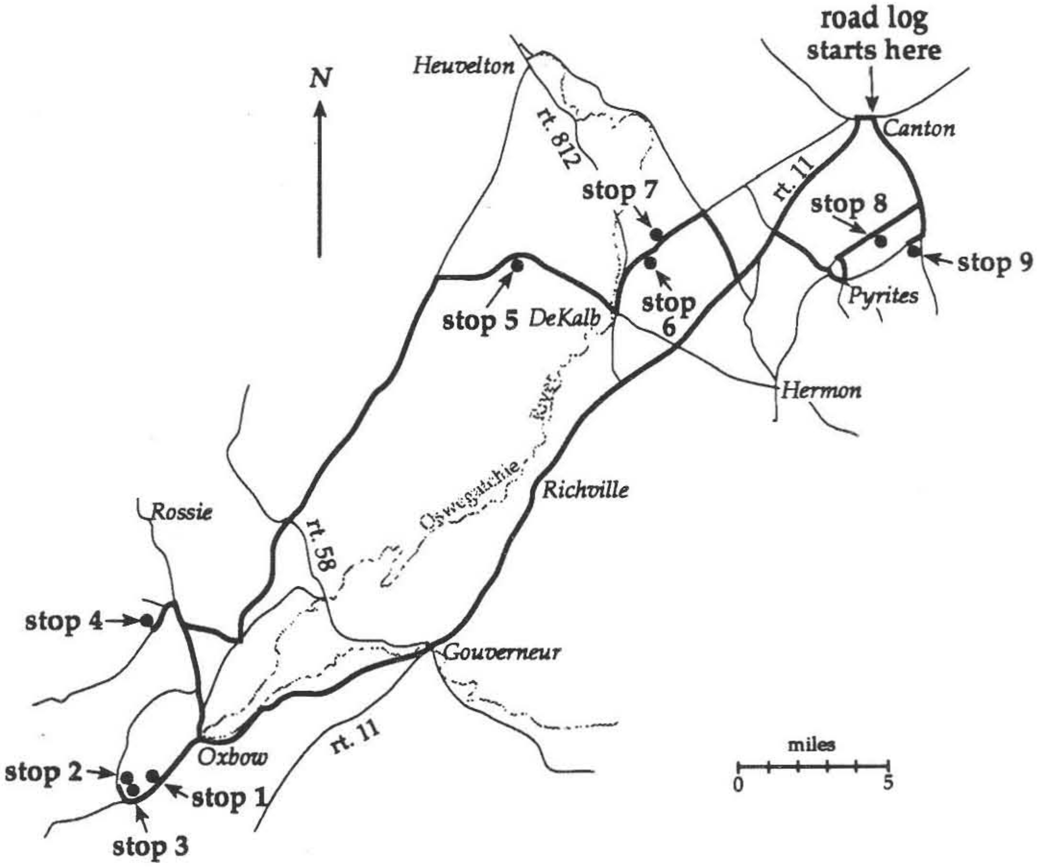

Figure 6. Route map for the field trip, with heavy lines indicating the route of the trip. The trip begins and ends in Canton.

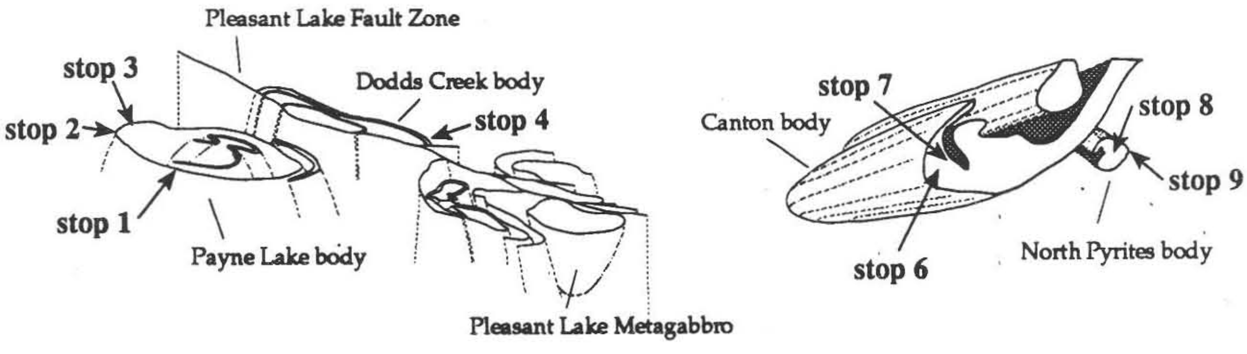

Figure 7. Locations of field trip stops relative to the geometries of the pertinent structures.

ROAD LOG FOR RETHINKING GRENVILLE-AGE DEFORMATION

A map view of the following road log appears in figure 6. The locations of stops relative to the geometries of pertinent structures in the Lowlands appear in figures 1 and 7.

cumulative mileage	miles from last point	route description
0.0	0.0	Begin at traffic light in downtown Canton at the intersection of Park Street and Main Street. Proceed west on route 11 (Main Street).
0.3	0.3	Turn left at traffic light, following route 11.
1.0	0.7	Hyde School Gneiss of Canton body on the left.
4.3	3.3	The famed Snake Roadcut. The marbles exhibit complex fold patterns, some of which may be sheath folds.
8.4	4.1	Dekalb Junction. Continue straight ahead on route 11.
24.2	15.8	Downtown Gouverneur at intersection of routes 58 and 11. Continue straight ahead on route 11.
24.5	0.3	Turn right onto Johnstown Street.
30.6	6.1	Hamlet of Wegatchie.
33.0	2.4	Hamlet of Oxbow.
33.2	0.2	Turn left onto Jefferson County route 22.
34.7	1.5	Payne Lake Fishing Access Site, where there are good views of cliffs of Hyde School Gneiss on the eastern margin of the Payne Lake body. Fields east of the Payne Lake body are largely underlain by marble.
35.6	0.9	**STOP 1.** Park at the gate on the west side of the road, and walk west along the State access at the edge of the field to the outcrops of Hyde School Gneiss (HSG) visible from the road.

STOP 1: Main foliation (S_m) in the Hyde School Gneiss along the eastern limb of the Payne Lake body (figure 7). The contact is not exposed but lies between non-resistent metasediments and resistant HSG at the west edge of the field. The best exposures lie along a very small stream bed in the first set of outcrops immediately south of the fishing access path. The stop description begins in the stream bed at the first set of HSG outcrops.

The HSG at this stop is a lineated and well-foliated pink leucogneiss with interlayered amphibolite. The main foliation (S_m) is subvertical and strikes parallel to the margin of the body, and the lineation (L_m) plunges steeply in the main foliation plane. Because the outcrops are dominantly quasi-horizontal glacially polished surfaces, the lineation is difficult to see. One of the only vertical foliation surfaces exposed at this stop lies in the stream bed, where it is clear that the lineation is a very nice quartz ribbon lineation.

Thin sections show that the main foliation is a well-developed shear fabric displaying sigma grains with core and mantle texture, quartz ribbons ramping from shear plane to shear plane, and dynamically-recrystallized grains. The fabric is well-recovered and shows polygranular quartz ribbons, no grain shape preferred orientation of fine quartz grains, and generally straight grain boundaries. Sense of shear is east side up.

The HSG contains abundant coarse-grained granite and pegmatite, and it is clear at outcrops such as this one that granitic liquids were present throughout the shearing history of the HSG. Some pegmatites are thoroughly sheared, streaked out parallel to S_m as bumpy chains of coarse K-feldspar porphyroclasts in the main foliation plane. Other coarse granitic phases are discordant and show no fabric at all. A particularly nice example of an intermediate state occurs on the flat part of the outcrop above the lineated surface along the stream bed. Here, coarse-grained granite is clearly discordant to the main foliation but is itself weakly foliated.

While much of the HSG in the Payne Lake body is devoid of amphibolite, localities such as this one at the margin of the body typically contain abundant interlayers of amphibolite. Amphibolite interlayers lie parallel to the main foliation, and have been sheared out and streaked out parallel to the foliation. It cannot be overemphasized that the current relationship between amphibolite and leucogneiss is structural.

Amphibolite layers are discontinuous along strike and are commonly segmented. Some of the layer terminations are clearly isoclinal fold hinges. A fine example of such a fold occurs in the stream bed immediately east of the large patch of junipers separating the stream bed from the higher outcrops extending to the south. Here, a biotitic amphibolite layer has been folded into an isoclinal fold with the main foliation lying parallel to the axial plane.

Outcrops stretching to the south show other important features of the interlayered amphibolites. As in many of the other bodies of the HSG, amphibolite layers are divided into blocky segments separated by medium- to coarse-grained granite. The granite separating the blocks is typically unfoliated or less well-foliated than the adjacent leucogneiss. A number of amphibolite interlayers in this large exposure show fold terminations, and many of the fold sets are not simple trains of isoclinal folds. While none are unequivocally sheath folds, many have outcrop patterns that could reflect sheath folding.

35.6	0.0	Continue southwest on Jefferson County route 22.
36.6	1.0	Turn right onto New Connecticut Road. Ask permission to visit stops 2 and 3 at the Raymon Farm on the corner of New Connecticut Road and County Route 22. Resistant rocks of the Hyde School Gneiss form a prominent knobby, wooded rise across the fields to the north of New Connecticut Road. The southern contact of the Payne Lake body with surrounding metasediments lies approximately at the break in slope below the wooded rise.
36.8	0.2	Newly excavated roadcut of serpentinized marbles.
37.2	0.4	**STOP 2.** This property is farmed by the Raymon family at the corner of New Connecticut Road and County Route 22; ask permission before visiting this locality. Park along the road, and walk north across the field. Follow the farm track that rises from the field diagonally up across the contact at the break in slope marking the transition to more resistant HSG. Proceed to the top of the rise immediately west of the farm track. The stop description begins in the HSG outcrops at the top of the rise.

STOP 2: Discordant foliation (S_{d1}) in the Hyde School Gneiss along the southern end of the Payne Lake body (figure 7).

Figure 7 shows clearly that this locality lies at the southern end of the Payne Lake body, where the main foliation (S_m) in the HSG, as well as the compositional layering and the contact between HSG and surrounding units, swings around the end of the body. At this locality, S_m and the contact dip very steeply south and strike approximately N35W.

The most prominent feature in the dark pink leucogneiss at the top of the rise is a well-developed, steeply-plunging quartz ribbon lineation that is well-exposed on quasi-vertical surfaces. Close examination of the outcrop shows a striking thing. The lineation is associated with a weak, sub-vertical foliation striking approximately N55E, parallel to the *length* of the Payne Lake body, *not* parallel to the margin of the body. The main foliation (S_m) is difficult to locate in this exposure, where the lineated fabric is well-developed and where there is no compositional layering in the HSG. Interlayered amphibolite *is* exposed to the northwest below the crest of the rise, and the orientation of the foliated amphibolite makes it very clear that the prominent lineated fabric is discordant to both the main foliation (S_m) and the compositional layering. We have named the well-lineated, discordant fabric S_{d1} (the subscript "1" derives from the fact that we find a second discordant fabric of slightly different character but similar orientation in the Dodds Creek body, as described in the log for stop 4).

Thin sections show that S_{d1}, like S_m, is a well-developed shear fabric displaying sigma grains with core and mantle structure, quartz ribbons ramping from shear plane to shear plane, and dynamically-recrystallized grains. The fabric is well-recovered and shows polygranular quartz ribbons, no grain shape preferred orientation of fine quartz grains, and generally straight grain boundaries. Sense of shear is east side up. S_{d1} is a distributed shear fabric but is not equally well-developed at all localities around the southern margin of the Payne Lake body. In some areas of outcrop, the most prominent fabric in the rock is the well-lineated S_{d1} fabric; in other areas, the most prominent fabric is S_m.

In summary, the Payne Lake body exhibits two distinct shear fabrics, one parallel to the margins of the body (S_m) and one parallel to the long dimension of the body (S_{d1}). Along the eastern and western limbs of the body, the two fabrics are parallel to one another and are consequently particularly well-developed, especially along the eastern margin (e.g., at the previous stop). We have interpreted S_m to have been the earlier fabric developed in a sub-horizontal regional shear zone. As sheath folds grew, S_m was folded into very large folds with highly recurved hinges. Figure 1 shows the geometry of the Payne Lake sheath. S_{d1} developed with a shear sense consistent with earlier shear directions as a younger shear fabric parallel to the axial plane of the growing sheath. Along the limbs of the sheath, S_{d1} lies parallel to S_m (figure 4b); in the hinge region at the north and south ends of the body, S_{d1} is discordant to S_m.

37.2	0.0	Continue west on New Connecticut Road.
37.3	0.1	Turn around in driveway, and go back east on New Connecticut Road.
37.5	0.2	**STOP 3.** This property is owned by the Raymon family at the corner of New Connecticut Road and County Route 22; ask permission before visiting this locality. Park along the road, and walk north across the field. The contact between the Payne Lake body and adjacent units lies at the prominent break in slope at the edge of the wooded area. This stop description covers exposures of units adjacent to the Payne Lake body in outcrops on the low shoulder north of the field but south of the Payne Lake body itself. The stop description begins in the cluster of outcrops at the top of the rise.

STOP 3: Sheath folds and late dike with mylonitic fabric in units adjacent to the Hyde School Gneiss (figure 7).

A variety of metasediments, including feldspathic quartzites, marbles, and garnet-sillimanite gneiss, are interlayered with leucogneiss in these outcrops. The contact here between HSG and surrounding metasediments gives every appearance of being gradational.

Compositional layering and the main foliation (S_m) in these units dip steeply and strike approximately N80W, parallel to the margin of the Payne Lake body. In the flat outcrops at the top of the rise, garnet-sillimanite gneiss exhibits both eye-shaped folds several centimeters across and refolded isoclinal folds. We interpret the eye-shaped folds as sheath folds. Sheath fold long axes and isoclinal fold hinges plunge steeply in the foliation plane, parallel to quartz ribbon stretching lineations in adjacent HSG. The presence of mesoscopic sheath folds provides additional supporting evidence for interpretation of the Payne Lake body as a large sheath fold.

Refolds in isoclinal intrafolial folds and tight to open folds in S_m also occur in this outcrop. A moderately well-developed foliation is locally developed parallel to the axial planes of these folds. The folds plunge steeply, and the axial plane foliation is subvertical and strikes approximately N50E, subparallel to S_{d1} and to the long dimension of the Payne Lake body.

Walk east to the last major set of outcrops before the ground drops away to the farm in the distance to the southeast. A sub-vertical dike approximately 15cm thick and striking N60E cuts interlayered metasediment and leucogneiss. The dike displays a prominent mylonitic fabric that is best developed within the dike itself but that does occur in the country rock on either side of the dike. The mylonitic fabric is subvertical and strikes approximately N50E, subparallel to the long dimension of the Payne Lake body and at a high angle to S_m in the outcrop.

Thin sections of the mylonitic fabric of the dike show sigma grains with core and mantle structure, quartz ribbons, and dynamically-recrystallized grains. The fabric in this dike is the least-recovered of the fabrics at Payne Lake. Serrated grain boundaries are common, and some grain shape preferred orientation of fine quartz is present. Some later cataclasis is also evident. Orientations of outcrop surfaces makes it impossible to locate the stretching direction of quartz ribbons in the field. Multiple sections, however, show that quartz elongation is best developed in vertically-oriented sections, suggesting a steep stretching lineation. Shear sense is east side up, consistent with shear sense on both S_m and S_{d1}. Based upon shear sense and orientation, the argument could be made that this fabric, and the axial planar fabric in the sheath fold outcrop, are likely S_{d1} fabrics. The fact that the mylonitic dike fabric is relatively poorly-recovered suggests that it represents a late phase in development of S_{d1}.

The features at this stop have two implications. First, shear in the Payne Lake body apparently had a long history, developing from a major shear zone with regional shear fabrics (S_m) to sheath folds with axial planar shear fabrics (S_{d1}) to late, discreet zones of shear with poorly-recovered mylonitic fabrics (late phase S_{d1} or S_{d2}). Second, intrusion of granitic liquids persisted until the latest phases of shear, and some were clearly important in localizing melt-enhanced shear. This raises the question of the importance of melt-enhanced deformation in the overall development of the Payne Lake sheath fold.

37.5	0.0	Continue east on New Connecticut Road.
37.6	0.1	Newly excavated roadcut of strongly lineated metasediments.
37.9	0.3	Turn left onto Jefferson County route 22.
41.2	3.3	T-intersection. Turn left onto Jefferson County route 25.

42.1	0.9	Turn left toward Rossie on St. Lawrence County route 3. Good views of the northern end of the Payne Lake body to the southwest.
46.0	3.9	Bridge over the Indian River. The river channel follows the Pleasant Lake Fault Zone (PLFZ), a major late brittle structure (figure 1). Slip along the PLFZ has removed half of the Dodds Creek body, leaving the body with a steep-sided, half-dome shape. Cataclasis along the PLFZ significantly reduced the resistance of gneisses along the fault. Swamp and floodplain cover the entire width of the fault zone, and rocks within the fault zone itself are not exposed anywhere along the Indian River. Mylonites in the Lowlands, on the other hand, are *not* typically more weakly resistant than unmylonitized lithologies.
46.2	0.2	Turn left onto Lead Mine Road.
46.7	0.5	Turn left onto River Road.
46.9	0.2	Good view of the north end of the Dodds Creek body "at 2:00". The contact between the HSG of the Dodds Creek body and the weakly resistant metasedimentary lithologies underlying adjacent fields lies at the base of the knobby exposures of pink leucogneiss.
47.3	0.4	**STOP 4.** This property is owned by Mr. Edmon Phalen in the farm next door; please ask permission to visit this locality. Park along the road, and walk to the outcrops on the knoll west of the road.

STOP 4: Main fabric (S_m) and discordant foliation (S_{d2}) in the Hyde School Gneiss along the northern end of the Dodds Creek body (figure 7).

The HSG at this stop near the margin of the Dodds Creek body is a pink, medium- to fine-grained leucogneiss with abundant interlayers of amphibolite. Flat outcrops on the north side of the knoll near the barnyard have abundant interlayers of quartzite several centimeters thick. While both the Dodds Creek and Payne Lake bodies have been previously mapped as consisting entirely of leucogneiss and interlayered amphibolite, we have established that both bodies, in fact, contain significant metasedimentary interlayers. The Dodds Creek body contains both thin interlayers of metasediment, as at this stop, and two wide swaths of interlayered marble, quartzite, and garnet-sillimanite gneiss (the concentric half-rings in figure 1). The Payne Lake body also contains a prominent metasedimentary interlayer with a peculiar "ear" shape (also shown in figure 1).

The HSG at this stop is well-foliated parallel to compositional layering, and thin sections from outcrops at the top of the knoll show that this main foliation (S_m) is a shear fabric, as it is at Payne Lake. Fabrics display sigma grains with core and mantle structure, quartz ribbons ramping from shear plane to shear plane, and dynamically-recrystallized grains. The fabric is well-recovered and shows polygranular quartz ribbons, no grain shape preferred orientation of fine quartz grains, and generally straight grain boundaries.

S_m at this stop is sub-vertical and strikes N95E, parallel to the margins of the Dodds Creek body. Stretching lineations plunge moderately steeply in the S_m foliation plane. At the southern margin of the Dodds Creek body, quartz ribbon lineations have a somewhat more variable orientation, consistent with reorientation during development of a large sheath fold in S_m. Along the margins of the body, quartz ribbon lineations plunge steeply. Sense of shear along the margins is east side up, as it is at Payne Lake.

A number of prominent discreet ductile shear zones several centimeters thick occur in the HSG at this stop and elsewhere in the Dodds Creek body. These zones are sub-vertical and strike N40E, approximately parallel to the long dimension of the Dodds Creek body. At the north and south margins of the Dodds Creek body, the mylonitic fabric of these zones is clearly distinct from the main shear fabric (S_m), because it is discordant to S_m. As at Payne Lake, the two fabrics can be difficult to distinguish from one another where they are quasi-parallel along the long margins of the Dodds Creek body.

There are four significant differences between the main foliation (S_m) and the younger discordant mylonitic fabric. First, S_m is a distributed shear fabric, moderately well-developed to well-developed over zones 10's to 100's of meters wide. The discordant fabric, one the other hand, is distinctly domainal, occurring in ductile shear zones several centimeters thick. Second, the discordant fabric is clearly a mylonitic fabric in outcrop; the taffy-like appearance stems from abundant quartz ribbons and dynamically-recrystallized grains. Despite the fact that S_m is not as clearly mylonitic in outcrop, thin sections show unequivocal shear fabrics. In fact, when one is hunting for foliations in lichen-covered outcrops without amphibolites in the Dodds Creek body, it is very easy to latch onto the discordant fabric as the main fabric, because it is so easy to see. One must be careful to look for *both* fabrics. Third, the discordant fabric is only poorly-recovered. Serrated grain boundaries are common, and pronounced undulose extinction and grain shape preferred orientation of fine quartz are present. Fourth, and most interesting, the sense of shear in this discordant fabric is *opposite* from shear both on the main foliation (S_m) in the Payne Lake and Dodds Creek bodies and on the discordant fabric (S_{d1}) in the Payne Lake body. Every sample we examined from the discreet ductile shear zones shown west side up and slightly oblique sense of shear. We named this fabric S_{d2}, presuming from its un-recovered character and opposite shear sense that it is younger than S_{d1} at Payne Lake. Figure 4c shows the geometry of S_{d2} in the Dodds Creek body.

As at Payne Lake, melt-enhanced deformation seems to have been important in late shearing at Dodds Creek. Several of the ductile shear zones at this stop are developed in what appear to be granitic dikes injected across S_m roughly parallel to S_{d2}.

We have found only limited evidence at Dodds Creek for an older pervasive discordant fabric (S_{d1}) similar to the one discovered in the Payne Lake body. At this stop, HSG exposed on the sloping outcrop surface below the phone pole shows a very weak pervasive fabric discordant to S_m and roughly parallel to the long dimension of the Dodds Creek body. Exposures along the Indian River immediately adjacent to a spur of the Pleasant Lake Fault Zone at the northern end of the Dodds Creek body show fabrics of similar orientation related to folds much like those at stop 3 in the Payne Lake body. Thin sections show that these are weakly-developed shear fabrics.

The HSG at this stop and at many places in the Dodds Creek body is laced with fracture swarms oriented approximately parallel to the Pleasant Lake Fault Zone. Thin sections show nice cataclastic textures. It is important to note that, while intensity of cataclasis does increase with proximity to the Pleasant Lake Fault Zone, intensity of ductile shear fabric development does *not*.

Evidence from the Dodds Creek body reveals an even more protracted history of ductile shear than that suggested by work in the Payne Lake body. Fabrics in the Dodds Creek body are consistent with development of S_m in a regional subhorizontal shear zone. Sheath fold development followed, with production of weak axial planar fabric (S_{d1}). Shear sense reversed, and discreet ductile shear zones developed but did not have time to recover before they were affected by cataclasis. Shear sense reversal and progression to brittle deformation is consistent with late extensional unroofing in the Lowlands.

47.3	0.0	Continue south on River Road.
47.4	0.1	Turn around in the partly paved road on the left. Proceed back north on River Road.
47.6	0.2	Good views of the Pleasant Lake Fault Zone to the right of the road, with rocks east of the PLFZ underlying the wooded rise beyond the river.
48.0	0.4	T-intersection. Turn right onto Lead Mine Road.
48.5	0.5	T-intersection. Turn right onto St. Lawrence County route 3.
48.7	0.2	Bridge over the Indian River (again).
49.3	0.6	Turn left onto Scotch Settlement Road.
51.2	1.9	T-intersection. Turn left onto Old State Road (St. Lawrence County route 10).
55.9	4.7	Intersection with route 58. Continue straight on Old State Road (St. Lawrence County route 10).
65.7	14.5	Turn right onto Mayhew Road.
68.3	2.6	Farm for permission for STOP 5.
68.9	0.6	**STOP 5.** Park on the right hand side of the road. This property is owned by the Amish farm on the north side of the road and back about 0.6 miles; please ask permission to visit this locality. Walk into the field to the south of the road to the crest of rounded knobs of granitic gneiss about 1600′ from the road, bypassing the wooded knoll approximately 1050′ from the road.

STOP 5: Shear fabrics in biotitic augen gneiss and surrounding metasedimentary lithologies immediately west of the Beaver Creek Fault Zone (figures 1 and 5).

Figure 5 shows a condensed and generalized geologic map of the Beaver Creek region based on an excellent and detailed map by Brown (1989). Several aspects are pertinent to this stop. First, granitic gneisses of two different ages appear in the region. The *ca.* 1225-1230 Ma Hyde School Gneiss is exposed in two bodies, the larger Hyde School body and the smaller Hickory Lake body. The other granitic gneisses in the region are younger than the HSG, having been intruded *ca.* 1150-1170 Ma. Second, while controversy rages over whether the HSG was originally plutonic or volcanic, no one disputes the plutonic origin of the younger granitic gneisses in the Beaver Creek region. Third, the region is divided into a number of elongate zones by northeast-striking faults, including the Hickory-Mud Lakes Fault, the Beaver Creek Fault and the Pleasant Lake Fault (the northern extension of the fault that cuts off the Dodds Creek body, as described at stop 4). Fourth, outcrop patterns of both metasedimentary and metaplutonic units lying immediately west of the Beaver Creek Fault Zone (BCFZ) are attenuated and elongated NE-SW, while similar units have larger, more irregular outcrop patterns farther northwest, suggesting the possibility of major ductile shear in the zone west of the BCFZ. Fifth, both the granitic gneisses and the metasedimentary units immediately west of the Beaver Creek Fault Zone are well foliated and display the "main regional foliation". This stop will give us a chance to study the character of the main regional foliation in several lithologies other than the Hyde School Gneiss.

At this stop, we will examine the most northerly of a series of seven elongate granitic gneiss bodies lying immediately west of the Beaver Creek Fault Zone (figure 5). This granitic gneiss is not only different in age from the HSG but also different in character, being a pink to grayish-pink, coarse-grained biotite-quartz-oligoclase-microcline gneiss with distinctive microcline augen.

These gneisses are spectacularly lineated – quartz ribbons lie like pieces of linguini on the steeply-dipping foliation surfaces exposed on the knoll. The lineations are best viewed on the sub-vertical faces of the east side of the knoll. Lineations plunge approximately 25, 40E.

The foliation carrying the lineation is the main regional foliation and shows characteristics that indicate formation by ductile shear. In thin section, the foliation shows clear fabric asymmetry in sigma grains and ramping quartz ribbons. Relatively coarse grain size in porphyroclast tails suggests a period of annealing following development of the main shear fabric. By contrast, the early shear fabric in the HSG of the Payne Lake and Dodds Creek bodies is recovered but not well-annealed. Shear sense in the Beaver Creek granitic gneisses shows consistent sinistral strike slip with a small component of east side down motion along subvertical planes striking approximately N40E. The fact that the gneiss bodies are elongate parallel to the main shear fabric in the rocks suggests that the individual bodies may have been intruded into a developing shear zone. Several of the bodies farther southwest (figure 5) that show only poorly-developed shear fabrics may have been intruded either relatively late or into zones that did not experience as much subsequent shear strain.

As in the Payne Lake and Dodds Creek bodies of the Hyde School Gneiss, the granitic gneisses west of the Beaver Creek Fault Zone show evidence in thin section of additional shear fabrics younger than the shear fabric of the main regional foliation. Some samples show a younger well-recovered but unannealed shear fabric. This fabric displays classic core and mantle structure around sigma porphyroclasts, polygranular quartz ribbons, and grain size reduction. This younger fabric appears to lie parallel to the main foliation. Shear sense consistently shows a component of sinistral strike slip.

Some thin sections show a third shear fabric, this one an unrecovered fabric. This fabric is characterized by quartz grain shape preferred orientation, biotite fish, and rare feldspar sigma grains. Sense of shear is opposite that shown by earlier phases and has a component of dextral, rather than sinistral, strike slip.

Interestingly enough, the dextral shear sense observed on this youngest shear fabric is, in fact, the sense of shear that one might infer from the geometry of the sigmoid east of the Beaver Creek Fault Zone (figure 5), and it is also the same as the sense of shear that we have suggested for the Pleasant Lake Fault Zone where it truncates the Dodds Creek body farther southwest. Late formation of the sigmoid is consistent with the results of our reconnaissance work in the sigmoid, which suggest that formation of the sigmoid postdates development of the main foliation and lineation in that region.

Multiple shear fabrics suggest a protracted history of ductile shear, much as we observed in the Payne Lake/Dodds Creek region. Changes in shear direction also suggest a complicated kinematic picture.

Walk 240′ southwest from the crest of the knoll. While the contact itself between granitic gneiss and surrounding metasediments is not exposed, outcrops of metasediment lie very close to granitic gneiss on the east side of the knoll. As you walk southwest, examine the character of the foliation in the marbles. The foliation in the marbles and calc-silicate gneisses is defined primarily by compositional layering and lies parallel to the prominent lineated foliation in the augen gneiss. One of the striking aspects of these rocks is that spectacularly lineated granitic gneisses with well-developed shear fabrics lie within centimeters of metasedimentary lithologies that show absolutely no microfabric evidence of ductile shear.

The only fabric asymmetry that we noted in the metasediments occurs in the outcrop of marble 240' southwest of the knoll where we first examined the granitic gneiss. This rock is a marble with quartz lozenges several millimeters to centimeters in size floating in a sea of coarse, equant calcite grains. While the calcite aggregate preserves absolutely no record of the shear that produced the spectacular lineations in the adjacent granitic gneisses, many of the quartz lozenges have a distinctly asymmetric shape. The lozenges are sigmoidal in shape and ramp from one foliation plane to another. Shear sense is left lateral with a small component of down-to-the-east motion. This sense of slip is consistent with that observed for the earliest microfabrics in the granitic gneiss.

Well-lineated rocks with good kinematic indicators appear to occur primarily in the granitic gneisses of the region. The metasediments we examined here and elsewhere in the Beaver Creek region preserve essentially no kinematic indicators, even in locations within centimeters of highly-sheared granitic gneiss. Several factors may be involved. If melt-enhanced deformation was important during shear, the granitic material may have concentrated the shear, leaving adjacent metasediments with comparatively low shear strains. Second, quartzofeldspathic aggregates are ideal for preserving the kinds of kinematic indicators associated with ductile shear. Marble-bearing metasediments may, in fact, have accumulated as much shear strain as adjacent granitic gneisses, but shear fabrics may have annealed completely in the metasediments. Regardless of the cause, the effect is an important one to note and suggests that trying to sort out the history of regional shear in the Lowlands by examining the metasedimentary lithologies may well be an exercise in frustration.

In summary, the "main regional foliation" is clearly a shear fabric in both the well-foliated and lineated granitic gneisses we observed in the Beaver Creek Region and in the Hyde School Gneiss of the Payne Lake and Dodds Creek bodies, despite the fact that the Hyde School Gneiss is considerably older. Granitic gneisses of both ages also show conspicuous evidence for protracted ductile shear with complex shear histories. All of the evidence points toward the importance of ductile shear across the Lowlands over a protracted period of time in development of both major and minor structures.

Walk approximately 200' toward Beaver Creek to small, rounded outcrops of

gray to white marble. You are standing right at the edge of the Beaver Creek Fault Zone, which is very much like the Pleasant Lake Fault Zone that we saw at stop 5 – no outcrop and lots of creek and swamp. Close examination of the marbles in these outcrops reveals little in the way of fault zone features, despite proximity to the Beaver Creek Fault Zone. The lack of major cataclastic features in outcrops immediately adjacent to the Beaver Creek Fault Zone is also reminiscent of the Pleasant Lake Fault Zone.

68.9	0.0	Continue southeast on Mayhew Road.
69.1	0.2	Cross Beaver Creek and the Beaver Creek Fault Zone. The lack of outcrop in the weakly-resistant cataclastic rocks of this late fault zone is similar to that along the Pleasant Lake Fault Zone.
72.5	3.4	Cross the Oswegatchie River.
72.6	0.1	Hamlet of DeKalb. Turn left onto route 812.
73.7	1.1	Hamlet of Coopers Falls.
73.9	0.2	Turn right onto Old Canton Road.
74.0	0.1	Crossing the "Race Track". This low area of fields and swamps sweeping in a broad arc marks the weakly resistant metasedimentary rocks bordering the more resistant Hyde School Gneiss of the Canton body. The road climbs out of the Race Track onto HSG, which crops out in abundant rounded knobs.
74.7	0.7	**STOP 6.** Low pink outcrop on the east side of the road.

STOP 6: Lineated Hyde School Gneiss of the Canton body (figure 7).

If time permits, we will stop briefly to examine well-lineated Hyde School Gneiss of the Canton body at this locality.

HSG in the Canton body is lithologically similar to that in the Payne Lake and Dodds Creek bodies, but the orientations of lineations are conspicuously different. Lineations in the Payne Lake and Dodds Creek bodies plunge very steeply, consistent with the vertical orientations of the sheaths. Lineations in the Canton body, on the other hand, plunge moderately to shallowly throughout much of the body. Lineations at this locality plunge approximately 25, S60W, an orientation consistent with the overall shallow plunge of the complex Canton body finger.

Some publications have portrayed the Canton body as a southsouthwest plunging structure, e.g., Foose's "Loch Ness Monster" model that interprets the Canton body and the Gouverneur body as two eroded "humps" connected in the subsurface (figure 2). Orientations of both foliations and lineations in the Canton body, however, suggest that a shallowly westsouthwest-plunging overturned structure is a better interpretation (figure 1). Preliminary data from our on-going work on the Canton body show shear fabrics similar to those in the Payne Lake and Dodds Creek bodies, and we would argue that a shallowly westsouthwest-plunging sheath fold is a better interpretation for the large-scale structure of the Canton body. Preliminary thin section analysis on samples collected during summer 1993 suggests that shear sense in much of the Canton body is top side down to the west.

Work in progress shows fabrics in this part of the Canton body to be more like those in the Beaver Creek region than those in either the Payne Lake body or Dodds Creek body. Fabric asymmetry is present in ramping quartz ribbons and relict sigma grains with coarse asymmetric tails, but fabrics are strongly annealed, as they are in granitic gneisses of the Beaver Creek region. Fine, dynamically recrystallized grains, so common in the Payne Lake and Dodds Creek bodies, are rare in the Canton body.

74.7	0.0	Continue straight ahead on Old Canton Road.
75.7	1.0	**STOP 7.** Outcrops in the field to the west of the road. Ask permission at the Theron Stacey farm (west side of the road, first farm to the north across the creek). Park on the side of the road, and walk through the gate. Walk a short distance into the field along the farm track, and then proceed approximately 135' south to the first knoll.

STOP 7: Hyde School Gneiss of the Canton body and garnet-sillimanite gneiss in adjacent metasediments (figure 7).

This locality lies along the southern edge of a narrow, curved inlier of metasedimentary rock that gives the Canton body its famous "double fish hook" appearance. The inlier dips shallowly southwest beneath the westsouthwest to west-plunging tip of the Canton body (figures 1 and 7).

At this stop, we will examine the garnet-sillimanite gneiss that commonly (but not ubiquitously) occurs at or near the margins of the Hyde School Gneiss bodies. Similar garnet-sillimanite gneiss borders inliers of metasedimentary rock within both the Payne Lake and Dodds Creek bodies.

The small outcrop approximately 135′ south of the farm track displays well-lineated porphyroblastic garnet-sillimanite gneiss. Lineations plunge approximately 25, S75W, parallel to quartz ribbon lineations in adjacent mylonitic HSG.

Walk 40′ farther south to a small knoll consisting of HSG south of the contact. This outcrop nicely displays a characteristic feature of well-lineated and sheared HSG. Many outcrops of HSG are better lineated than they are foliated, and the appearance of the HSG varies considerably with the orientation of the outcrop surface. Surfaces oriented parallel or nearly parallel to the stretching lineation typically appear streaky and "well foliated". Surfaces oriented perpendicular or nearly perpendicular to the stretching lineation typically appear "poorly foliated" or even massive. This results, of course, from the fact that much of the "foliated" appearance arises from the lineation. Some exposures of HSG are essentially L-tectonites (lineation but no foliation, like a bundle of pencils); in such exposures, the foliation is impossible to locate on surfaces oriented perpendicular to the lineation. Understanding the variation in appearance with orientation can be very useful in outcrops where glacial erosion has sculpted rounded knobs and where there are no foliation surfaces to examine for stretching lineation orientation. Studying the outcrop for the "streakiest" surfaces can give a clue to the approximate stretching direction.

This locality exhibits other interesting features that we will not have time to examine during this trip. Roughly 800′ northwest along strike from the knoll where we examined the HSG, dip slopes on the main foliation (S_m) have prominent southwest-plunging corrugations 0.5-1.0m in wavelength. Examination of west-facing outcrop surfaces reveals that these corrugations are erosional features marking the intersection of slightly less resistent HSG with slight more resistent shear zones 1-2cm thick. The shear zones dip steeply west at a high angle to S_m. The shear zones have a prominent mylonitic fabric and a component of down-to-the-west slip. As in the Payne Lake, Dodds Creek, and Beaver Creek gneisses, younger shear fabrics such as these attest to prolonged and complex ductile shear in the Lowlands.

75.7	0.0	Continue northeast on the Old Canton Road.
75.8	0.1	Crossing the Little Race Track. The stream valley marks

		the location of a weakly resistant, curved "finger" of metasedimentary rock that gives the Canton body its well-known double fish hook map pattern (figure 1).
77.3	1.5	Turn right onto Forest House Road.
79.8	2.5	Turn left onto route 11.
81.8	2.0	Turn right onto the Eddy-Pyrites Road.
84.2	2.4	Turn right onto route 47. The following convoluted route description results from the fact that the route 21 bridge over the Grass River was closed when this road log was written. When the bridge re-opens, turn *left* onto route 47, instead of right. Cross the Grass River, and take the first left. That will bring you to mileage 86.0 below. If, on the other hand, you simply must see bustling downtown Pyrites, carry on and follow the road log as written.
84.6	0.4	Turn left onto North Woods Road.
85.0	0.4	Turn left.
85.1	0.1	Bridge across the Grass River.
85.2	0.1	Village of Pyrites. Turn left.
86.0	0.8	Continue straight ahead.
86.9	0.9	First outcrops of Hyde School Gneiss in the North Pyrites body.
87.4	0.5	**STOP 8.** Low outcrops on the right side of the road.

STOP 8: Spectacularly lineated Hyde School Gneiss of the North Pyrites body (figure 7).

This outcrop displays fabulously lineated Hyde School Gneiss near the northern margin of the North Pyrites body. Quartz ribbons in this outcrop are many centimeters long and plunge shallowly northwest (approximately 10, N60W). The foliation carrying the lineation dips shallowly north (N89W, 18NE).

The power pole at the east end of the outcrop is rip-rapped with locally-derived lineated HSG containing mafic tongues highly elongate parallel to the stretching lineation. The origin of these tongues is not clear at this outcrop. In a spectacular series of outcrops in the yard and gardens of the house across the road and to the west, however, features clearly show that the mafic material occurred originally in layers and has since been isoclinally folded and disrupted by shear. The tongues may very well have sheath-shaped geometries, although this is difficult to prove in these glacially-polished outcrops. We will not visit this locality during the field trip, because the gardens could easily be trampled. The owners, the French's, have been kind enough to give individuals permission to view the outcrops and could be approached for permission if a group is very small.

Unlike the annealed fabrics in rocks of the Canton body, fabrics in many samples from the North Pyrites body are recovered but not annealed. In addition, fabrics in some samples from the North Pyrites body are spectacularly *un-recovered* and display highly strained porphyrclasts and serrated grain boundaries, suggesting, once again, a complex and protracted shear history in the Lowlands.

87.4	0.0	Continue east.
89.4	2.0	Turn right onto St. Lawrence County route 25.
90.4	1.0	Turn right onto Pink School House Road.
91.0	0.6	**STOP 9.** Outcrops to the south of the road between the road and the woods.

STOP 9: Discordant shear fabric in the Hyde School Gneiss along the eastern margin of the North Pyrites body (figure 7).

This exposure of lineated and foliated Hyde School Gneiss lies along the eastern margin of the North Pyrites body. Unlike the HSG at stop 8, the HSG at this locality displays well-developed compositional layering oriented approximately N20W, 70SW. The prominent lineation in the outcrop does not, however, lie in the plane of the compositional layering. Rather, the lineation is carried in a foliation that dips shallowly northwest, discordant to the compositional layering. This discordant foliation has the same kinds of distinctive asymmetric structures characteristic of shear fabrics that can be seen in the HSG of the Payne Lake, Dodds Creek, and Canton bodies, and we have termed this discordant foliation S_d. The lineation at this locality plunges approximately 20, N50W.

The orientations of S_d and the associated lineation L_d vary somewhat across the North Pyrites body, but L_d generally plunges shallowly NW. Shear sense determined from preliminary examination of thin sections appears to be top side down to the northwest.

Based on both the shape of the North Pyrites body and on the presence of prominent shear fabrics, we tentatively suggest that the North Pyrites body is a sheath fold plunging shallowly northwest (figure 1). We would interpret the lineated foliation, S_d, in much the same way as we have interpreted the discordant foliation in the Payne Lake body, as an axial plane foliation that developed parallel to the sheath fold axial surface as the sheath fold grew in a major shear zone. Where the earlier fabric, S_m, lies at a high angle to the axial plane of the sheath, as it does at this locality, S_d is clearly discordant to S_m and to compositional layering.

91.0	0.0	Turn around, and go back east on Pink School House Road.
91.6	0.6	Turn left onto St. Lawrence County route 25.
93.8	2.2	Join St. Lawrence County route 27.
95.0	1.2	Bridge across the Little River.
95.3	0.3	St. Lawrence University campus.
96.0	0.7	Intersection of Park and Main Streets, downtown Canton.

TRIP B1

BEDROCK EROSIONAL FORMS PRODUCED BY GLACIAL PROCESSES, NO. 2 MINE, GOUVERNEUR TALC CO., GOUVERNEUR, NEW YORK.

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INTRODUCTION

This trip takes place at the No. 2 (Arnold) open pit mine of the Gouverneur Talc Co. NYSGA last visited this location in 1971 on a trip led by J.S. Street (St. Lawrence University). He described a number of highly polished grooves, flutes, crescentic fractures, and striations present on bare bedrock surfaces (Street, 1971). Stops on this field trip will focus on the geologic setting of the region, history of pit operations, and a re-examination of the intriguing erosional forms present on both horizontal and vertical exposed bedrock surfaces. Our discussion will center on the nature and classification of these erosional forms, field evidence suggesting their possible origin, relationship to regional ice flow patterns, and their possible implications regarding subglacial processes and conditions.

History of Talc-Tremolite Mining and Regional Structural Framework

Talc-Tremolite mining has been continuously active in this area since the 1880's. The Arnold open pit is located in the immediate vicinity of some of this early mining. The abandoned Arnold mine at the north end of the present #2 pit and the Wight mine at the south end represent old underground operations that are a part of this early history. Changes in demand, property lines, company acquisitions, and the lack of accurate geological data have combined to account for the fact that the current ore body is still available to modern mining.

Current markets for this industrial filler material are predominantly in the ceramic and paint industries. The tremolite is used mostly in ceramics and the talc ores in paints. Their use as filler in flooring, caulking compounds, rubber, and many other products account for the remainder of the market.

The ore zone is identified by the no. "13" of a total of 16 stratigraphic layers in the Upper Marble Formation of the Precambrian Grenville series. It is this Upper Marble that exhibits the zones of mineralization and alteration that are commercially important. These well layered beds are the southeastern flank of an overturned anticlinal fold, which created a post erosional surface with a regional N-43-E strike and N-W dip of about 45-50 degrees, resulting in the older beds overlying the younger. These rocks have been highly metamorphosed and folded through 4 or 5 phases. A.E.J. Engle proposed that the tremolite was formed during metamorphism by dedolomitization of dolomite, a process that has almost completely obliterated the host rock. Underground relationships, such as the usual lack of direct contact of tremolite and calcitic marble, do not entirely support this theory. Another theory proposed that the talc-tremolite-anthophyllite schist is associated with evaporites in the Upper Marble Formation; the protolith of this unusually Mg-rich rock was probably a magnesite-bearing, siliceous evaporite. The evaporates of anhydrite and gypsum increase as a constituent of this unit down dip. Unit 12 is a relatively pure marble, usually coarse-grained, ranging in color from dead white to light gray. Unit 14 was described as quartzose calcitic marble by Engle, but subsequent underground mapping shows as many as fourteen possible subunits, many of which may be repetitions by folding. A relatively recent discovery within these marble units that support relationships of orientation has been the identification of "stromatolites" in unit 4. The fact that they were found upside down in the underground zinc mines confirms the direction of the youngest to oldest layers.

Geomorphic Setting and Glacial History

The open pit mine of the Gouverneur Talc Co. is located in the region known as the Frontenac Axis. It bridges the south and north sides of the flat-lying sedimentary rocks of the western St. Lawrence Lowland and consists of low relief, northeast-southwest trending ridge and valley topography resulting from differential erosion of Precambrian crystalline bedrock. The No. 2 mine occupies one of a series of linear valleys which extend from the Axis to the edge of Lake Ontario. Most of the linear valleys are sub-parallel to ice flow directions, but are more closely oriented down-dip in the direction of regional slope towards the Ontario Basin. The rock-walled linear valleys (as first named by Wilson (1904)) are 5-10 km in length, 0.5-1.0 km wide, and are best developed in the cuestaform remnants of the Black River Group to the southwest of the Axis.

Features relating to the most extensive glaciation recorded in the area can be found to the south on the upland of the Tug Hill Plateau . Streamline forms trending southeast record regional flow patterns probably formed during maximum glaciation of the region. A change in orientation of the streamline features on the northwestern edge of the Tug Hill suggests a subsequent shift in flow conditions and may correspond to late glacial movement of an ice lobe, identified here as the St. Lawrence- Lake Ontario Lobe, which was funneled into the Ontario Basin with its border "wrapped" around the Tug Hill sometime after the Port Huron advance (> 13 ka B.P.) (Street, 1966; Muller, 1978; Pair and Muller, 1990).

Thinning and wastage of the post-Port Huron ice mass constrained ice lobes to the Black River Valley and the Lake Ontario and St. Lawrence Lowlands (Muller and others, 1986). Secondary sets of striae with clear crosscutting age relationships, and ice marginal borders which parallel contours along the northern promontory of the Tug Hill and sides of the Black River Valley, attest to the sensitivity of the ice mass and its margins to local relief during ice marginal recession. The positions and morphology of former ice borders in the study area are functions of bedrock relief. The ice margin initially descended off the slope of the Tug Hill, and later, onto the low relief of the western St. Lawrence and eastern Lake Ontario Lowlands. Following northward encroachment by Glacial Lake Iroquois, ice-border morphology during recession was additionally influenced by deep water at the ice margin. Ice border features in the Lowland include subaqueous ice marginal fans and morainal banks, while in the upland, subaerial outwash plains and moraine-esker-outwash complexes typify recessional margins (Pair and Rodrigues, 1993). Detailed understanding of the deglacial setting in the Frontenac Axis is important and establishes that the erosional features described below were produced beneath an ice lobe fronting on a deep proglacial lake. They are therefore probably subglacial, rather than subaerial, in origin.

BEDROCK EROSIONAL FORMS

Description and Classification

The bedrock erosional forms which can be studied at the No. 2 mine are only those which have escaped the stripping and quarrying operations associated with ore extraction. The remaining forms can be found on most of the unweathered bedrock surfaces. Erosional forms are present on both horizontal pavements and on the preserved vertical headwall above the operating pit. For the purposes of discussion, the erosional forms and other associated features have been grouped into the following categories:

- I. Chattermarks and Crescentic Fractures
- II. Plucked Surfaces
- III. Striations
- IV. Grooves (>.5 m wide and deep)
- V. Roches Moutonnees
- VI. Crag-and-Tail forms
- VII. Precipitates of calcium carbonate and cemented crusts
- VIII. S-Forms (categories from Kor et al.(1991)):
    - Transverse: Muschelbruch, Sichelwanne, Comma forms, Transverse Troughs
    - Longitudinal: Spindle flutes, Cavettos, Furrows.
    - Nondirectional: Undulating Surfaces, Potholes

Possible Origins

Many of the bedrock erosional forms present appear to be a direct result of the action of glacial ice. Forms most often attributed to crushing and fracture of bedrock such as the chattermarks and crescentic fractures were reported by Street (1971) and can still be observed on several of the remaining horizontal surfaces. Pressure melting and regelation, together with basal slip can readily explain the plucked surfaces, precipatates, and cemented crusts on both horizontal and vertical bedrock surfaces. Striations produced by glacial abrasion can be observed on virtually all of the unweathered surfaces. These include the insides of many overhanging ledges. In addition, Fred Totten reports that when the overburden was first stripped from these features in the 1960's, there were obvious 'tools' still associated with some of the grooves. Larger more complex features including the roches moutonnees and crag-and-tail forms present at the site reflect a combination of abrasion and plucking especially associated with several of the resistant intraclasts present within the marble units.

Other forms on the vertical bedrock face suggest processes other than those associated with the direct action of glacial ice. These are certainly the most controversial as to their origin. Similar forms were described by Sharpe and Shaw (1989) and a formal classification has been proposed by Kor et al. (1991). These S-Forms (sculpted forms) are a suite of erosion marks attributed by Kor et al. (1991) to the action of subglacial meltwater. "Broad sheetfloods of turbulent subglacial meltwater " (Shaw, 1989, p. 853) that were released catastrophically were invoked in explaining S-Forms, the linear valleys, and both depositional and erosional forms throughout the Lake Ontario Lowland and Frontenac Axis by Shaw and Gilbert (1990). This interpretation has been questioned by Muller and Pair (1992).

DISCUSSION

The bedrock erosional forms at the No. 2 Mine may provide information about subglacial processes and conditions. Pair and Muller (1990) suggested that the linear valleys in the region were initially the products of the structure and differential erosion of the rocks of the Frontenac Axis. These may have existed prior to the most recent glaciation but have been modified either by: a) glacial abrasion and plucking by ice; b) erosion by subglacial meltwater trapped between the relatively impermeable bed and the glacier, or c) some combination of both processes. The presence of such valleys would also have served as a conduit for available meltwater in a basal drainage system.

Field observations of the S-Forms present on the wall of the linear valley at the No. 2 Mine are suggestive of a medium other than glacial ice. Both the transverse and longitudinal forms display furrows with very sharp rims present on their up-ice (or up-current) sides, attached lateral furrows on either side which broaden and become more shallow with distance, and have sharp edges on highly curvilinear, often asymmetric forms. The morphology of such forms is remarkably similar to the scour which occurs around a bluff body or obstacle under conditions of unidirectional water flow. The distinct upstream and lateral furrows around the concretions strongly suggest separated flow and turbulent conditions. Such obstacle marks have been well described by Allen (1982) as forming in both fluvial and eolian environments. Further, consideration of the properties of the erosive agent responsible for the scour-forms also suggests an erosive agent of low viscosity. The sharp edges on highly curvilinear, often asymmetric forms, as pointed out by Allen (1982), as well as early workers like Chamberlin (1885), suggest that such behavior is unlikely for ice (Reynolds Numbers for ice have been estimated to be 1×10^{-13} (Sharpe and Shaw, 1989)) and is more readily ascribed to the action of a fluid. Corrasion by small volumes of sediment-charged subglacial meltwater satisfies the requirement for an erosive agent of low viscosity that could have produced the erosional forms described.

However, the above assertion must be carefully balanced by critical considerations of the volume and thickness of the flows necessary to have produced the forms described. A water depth greater than the relief of the S-forms is all that is required to produce the incised features on the vertical face. Further, most of the S-Forms have been subsequently striated. This suggests that the degree of separation of the glacier sole from the bed, or decoupling, was very limited, and only a moderate water depth (cm's only) was needed to produce the forms observed. We suggest that any available meltwater would have been channelized around bedrock highs and that turbulent conditions and therefore most of the meltwater erosion would have been restricted to the sides and bottom of the valley. Pair and Muller (1990) suggested that meltwater flows may have produced small-scale meltwater erosion forms with relatively small volumes of water present as channelized meltwater focused into bedrock lows and that the rock-walled, linear valleys functioned as tunnel valleys for subglacial drainage. Such an interpretation may provide a viable explanation for the presence at the No. 2 Mine of bedrock erosional forms attributable to both abrasion by ice and to water.

ACKNOWLEDGEMENTS

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ROAD LOG

The road log begins at the intersection of Park Street and Route 11 in the center of Canton.

Persons using this log in the future should be aware that the field trip stops within the No. 2 Mine are located on private property that is owned by the Gouverneur Talc Co. Permission must be obtained from the company to access this property.

Cumulative mileage	Miles from last point	Route description
Start		Junction of Main Street (Route 11) and Park Street in the center of Canton. Follow Route 11 signs out of Canton towards Gouverneur.
23.8	23.8	In Gouverneur turn left on Route 58 towards Fine.
28.8	6.0	Turn right onto Route 812 and proceed south.
30.6	.8	Turn left into the No. 2 Mine and bear left along the pit road **Watch out for very large ore trucks on the same road.**
31.1	.5	Park and walk to the bedrock knob overlooking the pit.

STOP 1:. North end of pit

At this stop mine geologists will discuss the tectonic framework of the Northwest Adirondack Lowlands, provide a overview of the mineralogy and stratigraphy at the pit, and summarize the history of mining operations.

Return to vehicles and retrace route along pit road.

31.6	.5	Turn left at the far end of the pit and descend onto the remaining bedrock floor of the valley.Park and assemble near the bedrock knob.

STOP 2: South end of pit.

NOTE: Please stay off the bedrock faces directly above the open pit. Beware of slippery footing on the vertical face. Many of the forms are best viewed from the road !

At this stop we will examine the bedrock erosional forms present on both horizontal and vertical surfaces at this end of the No. 2 Mine. Discussion will focus on the regional glacial setting, proposed classification of the forms, the possible origin of these intriguing forms.

Possible questions for discussion:

1. Do differences in the bedrock lithology control the location and distribution of the various bedrock erosional forms ?
2. Can embedded 'tools' frozen in the ice abrade with smooth, sharp surfaces and produce all of the erosional forms present ?
3. What do these forms say about the plastic nature of glacial ice at several scales ?
4. What is the significance of these erosional forms occurring on the sides of the linear, rock-walled valley that includes the No. 2 Mine ?

Return to vehicles, go back out and turn right on 812.

32.4	.8	Turn left onto Route 58 and proceed back to Gouverneur.
38.4	6.0	Turn left on Route 11 towards Watertown.
47.0	8.6	Turn right onto Fox Ranch Road.
47.6	.6	Turn right onto Co. Road 24 towards Oxbow.
50.6	3.0	Enter Oxbow, turn left onto Pulpit Rock Road.
51.0	.4	Park on the right and walk to Pulpit Rock.

STOP 3: Pulpit Rock.

Discussion at this stop will address the possible origin of the well known Pulpit Rock.

Return to vehicles and continue southwest on Pulpit Rock Rd.

54.2	3.2	Turn right onto Hull Road.
55.0	.8	Turn right onto Vroom Creek Road. We are in the bottom of another of the linear, rock-walled valleys common to this region.
58.2	3.2	Turn right onto Co. Road 24 and return towards Route 11.
59.0	.8	Bear right to stay on Co. Road 24.
61.8	2.8	Turn left onto Fox Ranch Road.
62.5	.6	Turn left onto Route 11 and return to Canton.

END OF LOG

TRIP B2

THE POTSDAM-GRENVILLE CONTACT REVISITED (II)

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INTRODUCTION

This field trip continues the review, initiated in Trip A2 (Bursnall and Elberty, this volume), of outliers and enclaves of younger arenaceous rocks in contact with Grenvillian gneissic basement. The trip will concentrate on relationships between this basement and cover rocks, or enclaves, in the Dekalb and Richville areas whose affinity with the Upper Cambrian (?) Potsdam Sandstone is equivocal .

Repeats of Stop 7 and 8 of Trip A2 (described below) provide a basis for the trip, which includes an investigation of a newly exposed section in Dekalb (Stop 3). This exhibits somewhat similar relationships to those seen at the Rock Island roadcut (A2, Stop 6) which have been proposed as originating from burial of a karst topography - specifically the infill of solution pockets in marble on the Precambrian erosion surface (Carl and Van Diver, 1971). Outcrops of quartzarenite, "metaquarzite", and conglomerate to the south of Dekalb which were seen by Bloomer (1965, 1967) as being inseparable from the surrounding Grenville will also be visited. These, or very similar lithologies are contained as clasts in some of the sandstones (*e.g.*, Stop 8, A2 and Stop 3, this trip).

"The Potsdam-Grenville contact in the St. Lawrence valley represents a hiatus of some 500 Ma. Mineral assemblages within the Grenville gneisses indicate that perhaps 25 km of material was eroded prior to the deposition of the transgressive Cambro-Ordovician sequence, over an undulating Precambrian erosion surface. Many of the isolated, predominantly sandstone, bodies within the Grenville gneiss terrane of the Adirondack Lowlands can be confidently interpreted as outliers of Potsdam Sandstone. .... Such enclaves vary in composition from matrix supported breccias to equigranular orthoquartzites and their depositional environment has been variously interpreted as: pre-Potsdam solution pocket infills (related to a karst topography on a Grenville marble surface); fault related debris slides; fault breccias; and fault scarp talus accumulations. Some are seemingly crudely interlayered with the gneisses and are locally foliated" (Bursnall and Elberty, this volume).

The age of these enigmatic rocks have variously been proposed as:

1) outliers of Potsdam Sandstone

2) pre-Potsadam, post-Grenville remnants

3) Grenville in age, in that they suffered at least the later stages of Grenvillian deformation

ROAD LOG

Cumulative mileage	Miles from last point	Route description
Start		Junction of Route 11 and Park Street in the center of Canton.
0.3	0.3	Head west, cross the Grasse River, and turn left (south) at the traffic lights.
8.2	8.2	Proceed through Dekalb Junction.
11.7	11.7	Cross over a railroad bridge and park well onto the shoulder at low but extensive outcrop of marble on north side of road opposite a contact with iron-stained sandstone. This is Stop 1.

STOP 1: East Dekalb (Stop 8, Figure 1)

For convenience, the following description is taken from Trip A2 of this volume.

"Roadcuts due south of Dekalb further illustrate the complexities within these post-Grenville rocks. A long series of low exposures on the northwest side of the road contain well-layered marble (reclined folds at the north end) to the north which in the central section give way to a complex steep contact with rusty weathered arenaceous rocks. A number of small sandstone breccia wedges penetrate downwards into the marble in the vicinity, again supporting the solution-pocket infill model for the larger scale relationships seen elsewhere (*e.g.* Rock Island Road). Evidence for sulfide mineralization is present at some contacts, in common with the Rock Island Road locality (Elberty and Romey, 1990).

Bedding in the sandstone is irregular and breccia/conglomerates are common, particularly lower in the section a hundred yards to the south and at the base of the southernmost outcrop

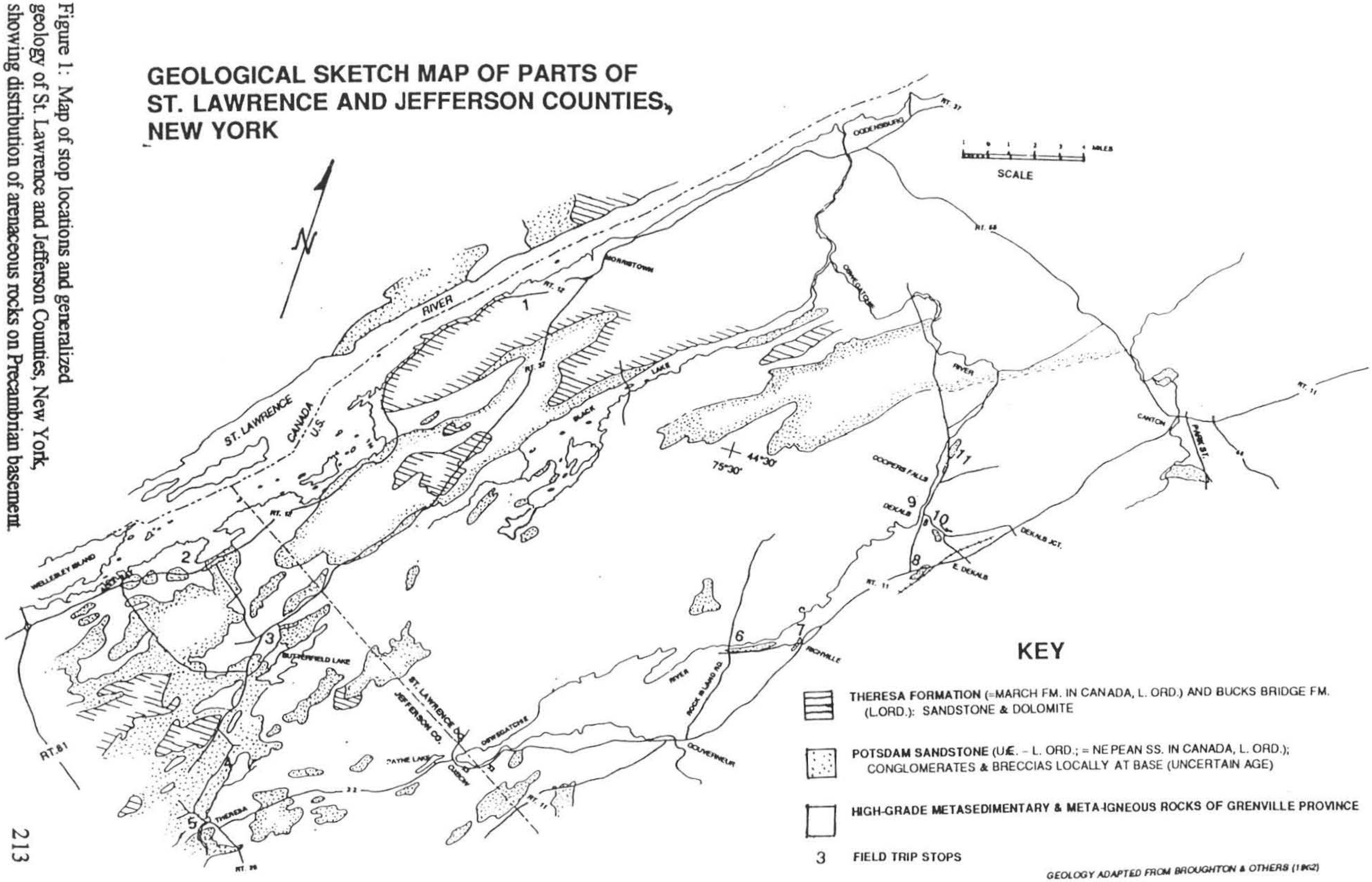

Figure 1: Map of stop locations and generalized geology of St. Lawrence and Jefferson Counties, New York, showing distribution of arenaceous rocks on Precambrian basement.

on the northwest side of the road. A poorly defined cylindrical structure exists in the latter section and may be seen on the top of a low outcrop of compact, rusty weathered, quartzarenite (compare with Stop 3).

Similar conglomerate and breccia occur in a large outcrop on the southeast side of Route 11. Quartz, quartzarenite, and metaquartzite are common clast compositions and are similar to those at Stop B3 (Stop 9 in Fig. 1). In both the western and southeastern outcrops shaly zones could have been generated by shear and at the northern end of the latter bedding dips are steep (> 60°; Karboski, 1977), yet again supporting the notion of significant post-depositional deformation" (Bursnall and Elberty, this volume).

Continue south on Route 11.

12.4	0.7	Junction with Route 812. Continue south on Route 11.
17.3	4.9	Road rises and curves to the left. Large outcrop on northwest side of road. Park well onto shoulder. Stop 2.

PLEASE WATCH OUT FOR FAST MOVING TRAFFIC

STOP 2. The Richville breccias (Stop 7, Figure 1)

Please note that description for this stop is taken from Trip A2, this volume

"Karboski et al. (1983) described this outcrop this as containing a "flow breccia" at its base, followed by a densely consolidated breccia with an overlying highly deformed metaquartzite, overlain by an "orthoquartzite", which contains pebble-sized quartz clasts - possibly derived from the underlying metaquartzite. The breccias contain quartz clasts set in a hematite stained, medium grained arenite. Large (0.5 m) phacoidal blocks of the breccia are enveloped by thin shaly borders, the whole giving the impression of a shear zone.

This outcrop certainly inhibits any notion that the unconformity between Precambrian basement and overlying cover is a simple one!

Points to concentrate on are:

1) Possible shear fabrics in the lower part of the outcrop, in part defining the borders of coherent blocks of breccia
2) Compositional variation of clasts in the breccias
3) The relationship between and the textural character of the each of the recognized rock types

If the upper part of this outcrop is indeed comprised of rocks which are part of the Grenville basement as supposed by Karboski (1976) and Karboski et al. (1983) then the current disposition of lithologies seems not to be satisfied by a model that involves karst infill alone (Van Diver, 1976) as seems possible at Rock Island Road. It is possible, however, that the wall collapse of a large solution depression may have allowed a slab of basement to slide into the argillite filled basin. The presence of shear fabrics within the breccia may be accommodated by this model provided that these rocks were only partially lithified at the time.

Cross over to the vehicle and investigate the southern end of the outcrop on the southeast side of the road. Here, a narrow zone of sandstone breccia dips steeply through marble. One of the contacts is sheared indicating high angle faulting. Is there evidence for displacement sense?

[Note: if time permits an outcrop of marble on the northwest side of Route 11, a few hundred yards to the south should be visited. It contains narrow subvertical veinlets of arenite at the northeast end, which provide good evidence for solution cavity infill]" (Bursnall and Elberty, this volume.

Return to vehicle and continue up the hill to Welch Road.

17.4	0.1	Turn around at Welch Road and head back towards Dekalb Junction
25.2	7.8	Turn left and follow Route 812 northwards
27.2	2.1	Stop 3. Large outcrop on the west side of the road at the entrance to Dekalb (gentle left-hand bend)
		Note: the best parking is approximately 100 yards to the south along a disused section of Route 812, opposite an Amish farm.

STOP 3. (Stop 9 in Figure 1)

This outcrop possesses a newly exposed surface resulting from road-widening and contains a number of irregular sandstone bodies within well-layered Grenvillian marble.

A number of sub-vertical arenaceous zones, varying in width from less than one to greater than ten feet in width occur throughout the outcrop. In the largest of these fragmentation and "stoping" of the marble wall rock is evident but relatively few areas exhibit "permissive-veining" relationships - where separated marble fragments could be pieced together in their pre-fragmentation configuration. In places, it is possible that injection of sand could have occurred from below. In addition to marble, abundant angular quartzarenite fragments similar to the foliated quartzite of Stop 4 are present.

Continue north on Route 812, towards the center of Dekalb. (Refreshments may be purchased at the village store)

27.55	0.25	Turn right (south) just after the store onto County Route 17 and head south towards Route 11 and Dekalb Junction.
28.55	1.0	Where this road curves sharply to the left, continue and park on the right. Stop 4 is in the woods on the right (west).

STOP 4. Deformed quartzarenites of possible Grenville age (Stop 10 on Figure 1).

Outcrops of conglomerate and foliated quartzarenite occur 75 yards into the woods (stay close to an old fence on the left until these outcrops are visible). These rocks were thought by Bloomer (1965, 1967) to belong to an infolded sequence within the Precambrian gneisses and to have been affected by probably the latest stages of Grenvillian deformation. Foliated and folded metaquartzite (best seen on the top of the ridge) is petrographically very similar to clasts at the previous stop and at Stop 1.

Fold style here, however, and the nature of the conglomerate (at the base of this outcrop), suggests that these rocks may not have experienced the full Grenvillian deformation sequence but could represent a post-Grenville, pre-Potsdam, depositional period.

Turn around and head north on Route 17 to Dekalb.

29.55	1.0	Turn right (north) onto Route 812 and continue along a section containing a number of outcrops containing quartzarenite of questionable age. A number of stops should be made in this section (see also Trip A3, Stop , this volume)

STOP 5, et seq. (Stop 11 on Figure 1)

The section referred to above extends from the northern part of Dekalb village to the Oswegatchie bridge near Kendrew Corners (at 3.25 miles from the junction of Routes 812 and 17). It includes a quarry on the east side of the road (at 0.8 mi.) and conglomerates exposed on the Oswegatchie River at Coopers Falls (at 1.3 mi.). To reach the falls ask permission at the house opposite the intersection of the Old Canton Road and follow the track that passes the house to the south. For additional description of a part of this section see Trip A3, Stop 2 in this volume.

Return to Canton along the Old Canton Road or through Rensselaer Falls, following County Route 14 (turn right at Kendrew Corners just north of the Oswegatchie bridge) and Route 68 south - about 12 miles.

END OF TRIP

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TRIP B3

SEDIMENTOLOGY AND DIAGENESIS OF THE POTSDAM SANDSTONE AND THERESA FORMATION, SOUTHWESTERN ST. LAWRENCE VALLEY

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INTRODUCTION

The lower Paleozoic strata in the southwestern St. Lawrence Valley consist of the Late Cambrian-Early Ordovician Potsdam Sandstone, Theresa Formation and Odgensburg Dolostone. This sequence totals approximately 120 meters in thickness, and is the age equivalent of the considerably thicker passive margin carbonate sequence of eastern New York and Western Vermont. In the area of this field trip, the lower Paleozoic sequence is well-exposed along the St. Lawrence River from the vicinity of Ogdensburg to Alexandria Bay, New York. In this area glacial sedimentary cover is thin or absent, and outcrop exposures are common in streams tributary to the St. Lawrence, and as roadcuts. The contact between the basal Potsdam Sandstone and the Late Proterozoic metamorphic and igneous rocks of the Grenville province is well-exposed in the area. The so-called Frontenac Axis is a topographic high in the Grenville rocks, and trends NNW-SSE. Regionally, Paleozoic strata dip gently to the ENE to the east of the Frontenac Axis and to the WSW to the west. The Medial Ordovician Black River Group overlies the Theresa Formation to the southwest of the area of this field trip. To the north and east Early Medial Ordovician Chazy Group strata overlie the Ogdensburg Dolostone and its equivalents.

The Potsdam Sandstone

The Potsdam Sandstone is one of the most widely-exposed units in the circum-Adirondack region of New York State, and is well-known as a building stone, particularly in the area of the type section near Potsdam, New York. In the St. Lawrence Valley, the Potsdam rests unconformably upon Late Proterozoic gneisses, granites, quartzites and other metasedimentary rocks. The Proterozoic rock beneath the unconformity is often rather altered, and in areas where the Proterozoic rock is marble, there are often very complicated relationships between the basement marble and sandstone. These altered zones and complex basement-cover contacts have been interpreted as soil horizons, and pre-Potsdam karst zones. Some workers have proposed that the sandstones in marble terrains may represent infolded Proterozoic sandstones that were not metamorphosed. The best explanation for the alteration of sub-Potsdam basement and the complex sandstone-marble relationships is that the basement rock was extensively altered chemically and mineralogically by hot sedimentary brines that circulated through the porous Potsdam and "attacked" the basement rock. Lateral and vertical transport of sand during brine flow into encavernated marble accounts for the complex features observed. High fluid pressures, perhaps driven by hydrocarbon cracking and methane production, likely characterized this period of brine flow. The production of organic acids during hydrocarbon maturation may have provided a source of aggressive reactants which facilitated dissolution of marble and alteration of silicate minerals in other basement rocks. K-Ar dates on illite associated with hydrothermal minerals in the basal Potsdam suggest that significant activity occurred in Late Devonian-Early Carboniferous time, perhaps when the region was buried deeply, and tectonic disturbance to the east and southeast triggered cratonward flow of fluids.

In the area of this field trip, the Potsdam is divided into a lower portion consisting of medium-coarse quartz arenites that exhibit types of cross-stratification and other primary structures suggesting shallow marine shoreface and subtidal tide-dominated depositional conditions. Non-marine aeolian facies, and braided stream facies are present in the lower Potsdam, but will not be seen on this trip. Conglomeratic facies are common in the Potsdam, but are particularly abundant in the basal portion in the vicinity of quartzite basement exposures. The lower Potsdam is virtually unfossiliferous in this area. Simple vertical living burrows and rare surface crawling traces are present in some of the thin-bedded sandstones. The enigmatic crawling trace *Climachtichnites* is present in slabs used for construction of a wall at the visitor's center of Wellsley Island State Park, near our last stop on the trip. The recent work of Yochelson and Fedonkin suggest that *Climachtichnites* may be limited to the Late Cambrian (Yochelson, personal communication 1990). Other fossils have not been found, and thus the age assignment of the lower Potsdam remains a question.

The lower Potsdam consists almost entirely of well-rounded detrital quartz grains cemented by secondary quartz, illite and rare kaolinite. The coloration of the basal Potsdam is due to finely disseminated iron and titanium oxides, which were formed during diagenetic breakdown of detrital ilmenite and magnetite. Detrital feldspars in the lower Potsdam are generally highly altered to masses of illite and/or kaolinite. The percentage of kaolinite increases abruptly near the contact with the upper unit of the Potsdam, and SEM study indicates that this kaolinite predates secondary quartz cements and illite. Such kaolinite may have formed during weathering and subaerial exposure that accumpanied the depositional hiatus between the lower and upper Potsdam.

Upper Potsdam Sandstone

The upper portion of the Potsdam Sandstone consists of medium-fine quartz sandstones with 3-6% detrital feldspar. Carbonate cements and recrystallized carbonate mud are present, often outlining burrows. The phosphatic inarticulate brachiopod *Lingulepis* is present, but other body fossils have not been found in the area of this study. The upper Potsdam commonly exhibits alteration of meter-scale burrowed and unburrowed facies. The burrowed sections display good examples of the u-shaped burrow *Diplocraterion;* the unburrowed sections expose ripple marks, mudcracks and small-scale bipolar cross-strata with reactivation surfaces. These facies are interpreted as sandy tidal flat environments, with the burrowed sections representing slightly more emergent or well-protected environments, and the cross-bedded and laminated sands slightly more active environments lower on the tidal flat system. The rhythmic repetition of burrowed and unburrowed meter scale units may record high-order sea-level changes. Similar patterns are seen in the middle portion of the Theresa Formation.

The age of the upper portion of the Potsdam has been assigned to the early Ordovician on the basis of a Tremodocian conodont fauna (Greggs and Bond, 1971). As noted above, the contact between the lower and upper units of the Potsdam Sandstone represents a hiatus in deposition of unknown duration. The hiatus is characterized by the development of intraformational sandstone breccias, kaolinite-silica cemented sandstone concretions, and abundant early kaolinite cement. This surface may represent a significant period of terrestrial weathering.

Theresa Formation

The transition from the uppermost Potsdam Sandstone to the lower Theresa Formation is rather abrupt, and at least in some sections marks a deepening from tidal flat facies to near wave-base offshore facies. The basal 5-8 meters of the Theresa Formation consists of thin to medium bedded calcareous and dolomitic siltstones and fine sandstones. Typical beds are 2-10 cm. in thickness, and contain a basal portion that is plane-laminated or ripple cross-laminated, and an upper portion that is bioturbated. These couplets are interpreted as tempestites - storm deposits - with the upper, bioturbated portion representing the recolonization of the substrate following a depositional event. A variety of horizontal grazing trails are present, as well as vertical escape burrows in the laminated basal portion of many beds.

Lingulepis debris is common in the lower portion of the Theresa Formation, and rare discoidal gastropods are found. Conodonts are abundant in some beds. Detrital feldspar may form up to 25% of the siliciclastic material in the lower Theresa. The original sediment is inferred to have been a mixture of siliciclastic silt and sand, with significant amounts of lime mud and other carbonate material. Most carbonate is now in the form of secondary dolomite and coarsely crystalline (neomorphic?) calcite.

The transition from the thin-bedded lower Theresa to the middle Theresa Formation involves overall increase in grain size, and increase in bedding thickness. The middle Theresa is packaged into alternating burrowed and unburrowed units that are grossly similar to the upper portion of the Potsdam Sandstone. The unburrowed units consist of trough and planar-tabular cross-stratified medium sandstones with common bipolar crossbed sets. The burrowed units are slightly finer grained, bioturbated sands with abundant dolomite and calcite. Carbonate may comprise up to 40% of the rock, with dolomite apparently replacing earlier lime mud and carbonate grains, and calcite present as neomorphic spar and late cement. The uppermost portion of the burrowed units is often capped by thin-laminated dolostone with calcite spar-filled voids. These sequences are interpreted as originating from the progradation of tidal flats, with the unburrowed portions representing the more wave and current reworked low tidal flat, and the burrowed, carbonate-rich facies the upper tidal flat, capped by algally-laminated muds. Spar-filled voids in the finer-grained facies may represent early voids formed by shrinkage or dissolution of sediments, or voids formed by later dissolution of evaporites. Significant later diagenetic alteration is indicated by pervasive dolomitization and calcite recrystallization. Pyrite is a common authigenic mineral, and commonly is confined to the darker, carbonate-rich facies. The pyrite may be an early authigenic phase formed during sulfate reduction resulting from shallow burial. Alternatively, the pyrite may have been introduced during later diagenesis that accompanied the movement of regional brines, and thus represents hydrothermal mineralization. The occurrence of galena, chalcopyrite and sphalerite in the upper Theresa and Odgensburg Dolostone suggests that Mississippi Valley Type mineralization occurred on a small scale in the region.

Upper Theresa Formation

The upper portion of the Theresa Formation differs from the middle portion in the presence of more carbonate-rich beds (sandy dolostones with up to 80% carbonate) and more evidence of subaerial exposure (mudcracks, intraclast breccias). The overall meter-scale pattern of burrowed and unburrowed facies is less regular in the upper Theresa, but evidence of tide-dominated deposition, in the form of bipolar sets of cross-strata, is abundant.

Carbonate-rich beds in the upper Theresa commonly contain abundant voids partially filled by "saddle" dolomite, calcite, quartz, barite, fluorite, pyrite and chalcopyrite. This mineralization is likely related to the regionally pervasive MVT-like hydrothermal mineralization widely observed in the St. Lawrence Lowlands. The timing of this mineralization is likely synchronous with the illite cements present in the lower Potsdam, because illite has been observed intergrown with barite and pyrite in the lower Potsdam. If this is so, then the mineralization occurred in late Devonian-early Carboniferous (355-360 mya), perhaps when regional brines were being transmitted through the sequence. Fluid inclusion studies of quartz crystals from nearby localities in the Ogdensburg Dolostone indicate minimum temperatures in excess of 180 degrees C. and salinities in excess of 25% NaCl equivalent.

REFERENCE

Greggs, R.G. and Bond, I.S. (1971) Conodonts from the March and Oxford formation in the Brockville Area, Ontario; Canadian Journal of Earth Sciences, vol. 8. #11; p. 1455-1471

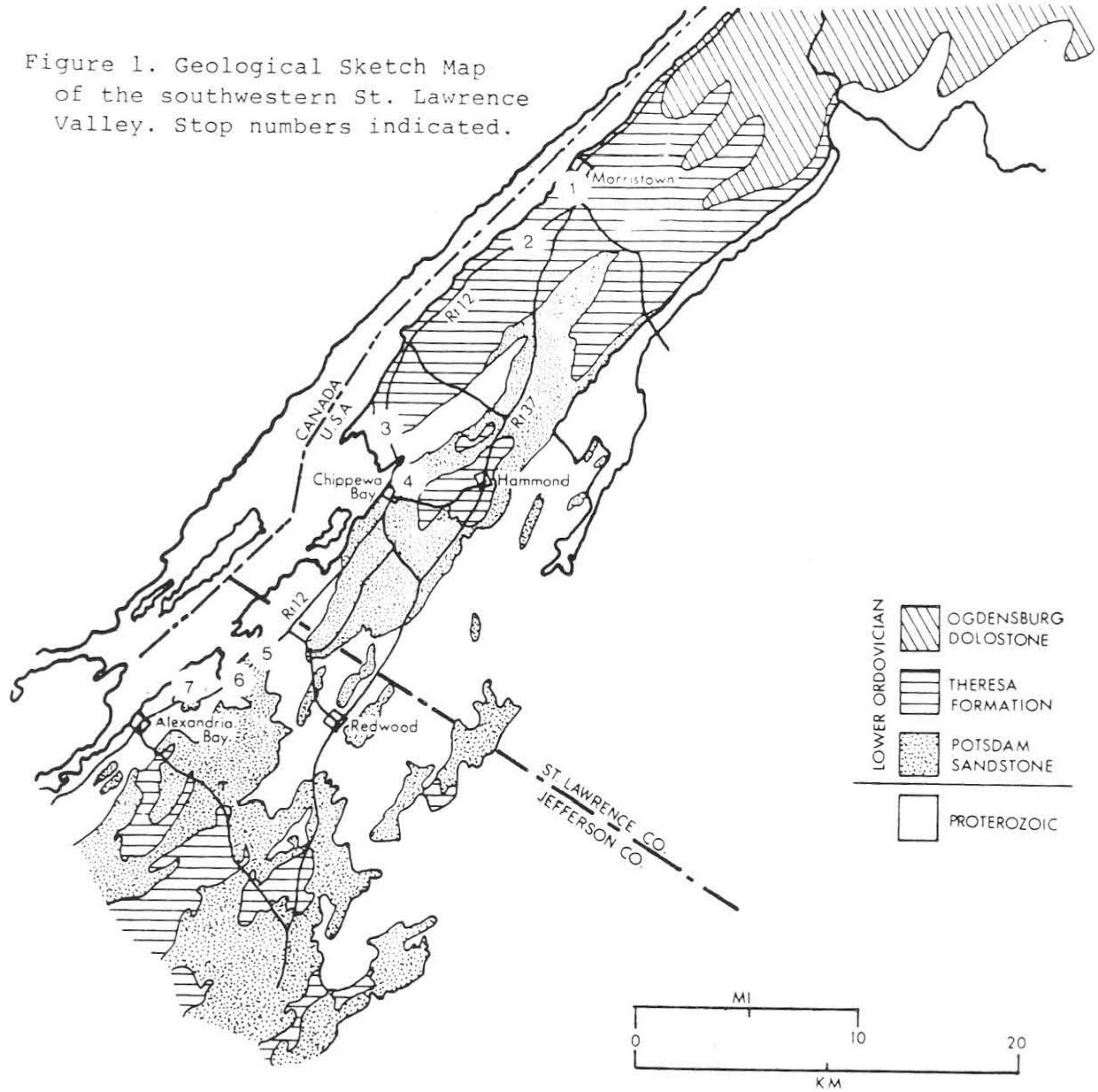

Figure 1. Geological Sketch Map of the southwestern St. Lawrence Valley. Stop numbers indicated.

ROAD LOG

This field trip will take us through the Theresa Formation and Potsdam Sandstone, and in general we will travel down section, with the first stop in the upper Theresa Formation, and the last stops in the basal Potsdam Sandstone. This route, although not the most desirable (it usually is easiest on the brain to move forward through time, rather than backwards) allows us to end the trip with easy access to Interstate Route 81, at the Thousand Islands International Bridge.

The road log begins at the intersection of Park Street and Route 11 in the Village of Canton, New York.

The road log begins at the intersection of Park Street and Route 11 in the Village of Canton, New York.

Cumulative mileage	Miles from last point	Route description
Start		Junction of Main Street (Route 11) and Park Street in the center of Canton. Follow Route 11 signs out of Canton towards Ogdensburg.
0.3	0.3	Proceed west (straight) onto Route 68
16.9	16.6	Proceed west (left) on Route 37
23.6	6.7	St. Lawrence State Park
28.1	4.5	Stop 1

STOP 1:

Roadcuts on both sides of Route 37 expose tidal flat facies of the upper portion of the Theresa Formation. The basal beds here consist of sandy dolostone with irregular voids containing authigenic dolomite, calcite, quartz, barite and sulfides. Other lithofacies present here include laminated silty dolostones, burrowed dolomitic sandstones and cross-laminated sandstones exhibiting well-developed bipolar cross-sets.

Continue southwest on Route 37.

28.8	0.7	Proceed west (bearing right) onto Route 12
30.6	1.8	Entrance road to Jacques Cartier State Park
31.9	1.3	Stop 2

STOP 2:

The roadcuts at this stop expose the middle portion of the Theresa Formation. The rhythmic interbedding of yellow-white cross-laminated sandstones and darker, burrowed dolomitic sandstones is typical of this portion of the Theresa Formation. The burrowed units are often capped by a few centimeters of laminated dolostone. The cross-laminated sandstones are interpreted as low tidal flat deposits, the burrowed dolomitic sandstones as upper intertidal, capped by algal-stromatolitic carbonate mud. These sequences are tidal flat cycles, and their origin may reflect both high-order sealevel change and tidal flat progradation that produced slight but sudden deepening followed by shallowing due to progradation.

Note the apparent soft-sediment deformation represented by the intraformational folds near the northeast end of the outcrops. Note also the more general low-amplitude folding and the high-angle faults.

Continue southwest on Route 12

35.5	3.6	Stop 3

STOP 3:

The contact between the uppermost Potsdam Sandstone and basal Theresa Formation is exposed on the south side of Route 12 near the southwest end of the outcrop. The contact is marked by the change in color (Potsdam = white-grey-yellow; Theresa = grey-brown), the abrupt increase in carbonate content in the basal Theresa Formation, and the change in bedding thickness from the thick to massive beds of the Potsdam to the thin beds of the basal Theresa. The lower Theresa in this area consist of 2-10 cm. beds of plane-laminated or ripple cross-laminated fine sandstones/siltstones that are capped by calcareous and dolomitic fine sandstones/siltstones. Escape burrows that traverse the laminated beds record attempts of organisms to return to the sediment surface following a sudden influx of sediment. The environment of the lower Theresa is interpreted as a subtidal, near wave base shelf lagoon setting characterized by sporadic sediment influx and physical disturbance of the bottom (storms). These events produced laminated and cross-laminated sheets of sand and silt, the tops of which were subsequently colonized by organisms, producing the bioturbated tops on each bed.

The upper Potsdam Sandstone at this stop consists of burrowed calcareous sandstones and laminated and cross-laminated sandstones that generally resemble the middle portion of the Theresa Formation seen at the last stop. The trace fossil *Diplocraterion* is common in the burrowed units, and a alternating low tidal flat - high tidal flat environment is envision for these facies. Fragments of the brachiopod *Lingulepis* are common in the upper Potsdam and lower Theresa at this stop. The lower Theresa here also contains scattered pebbles and cobbles of quartzite that were clearly derived from a Proterozoic basement knob exposed approximately 300 meters south of this outcrop.

Continue southwest on Route 12

38.8	3.3	Stop 3a

STOP 3a:

This stop replicates the sequence observed at our last stop, and we will only stay a short time to examine the features of the upper Potsdam Sandstone, which are somewhat better exposed here. Very well-preserved examples of *Diplocraterion* are present in the upper Potsdam on the southeast side of the outcrop.

Note the outcrop of Proterozoic basement immediately to the southwest along Route 12.

Continue southwest on Route 12.

40.4 1.6 Stop 4

STOP 4:

The contact between the lower and upper portions of the Potsdam Sandstone is exposed in the roadcut on the southwest side of Route 12. The lower section of the outcrop consists of medium-bedded, cross-laminated, medium-grained sandstones typical of the lower Potsdam. The upper, massive calcareous sandstone bed is highly burrowed, and is the base of the upper Potsdam. This facies of the lower Potsdam is interpreted as a shallow, wave and tidal current dominated shelf.

The contact between the lower and upper Potsdam represents a depositional hiatus of some duration. Networks of vertically-oriented cracks, brecciated sandstone and kaolinite-cemented concretions are common along this horizon, which is widely traceable throughout the St. Lawrence Lowlands, and clearly represents a sealevel lowstand. As noted in the text, the lower Potsdam may be Dresbachian, based upon the occurrence of *Climactichnites*, whereas the upper Potsdam is Tremadocian, suggesting that a portion of the upper Cambrian is absent due to non-deposition, or erosion, prior to deposition of the upper Potsdam.

Continue southwest on Route 12.

47.6 7.2 Stop 5

STOP 5:

The unconformity between the basal Potsdam Sandstone and underlying Proterozoic gneisses is exposed in the roadcut on the southeast side of Route 12. The time interval represented by this contact is some 600 million years. The basal sandstones here exhibit large-scale low-angle planar-tabular cross-bedding, and are devoid of fossils. The depositional setting for this facies is problematic, although shallow marine tidal inlet, shoreface, and perhaps even aeolian dune environments are possible facies models.

Considerable variation in color pattern is evident in the lower Potsdam here, with the basal meter or so consisting of white to light grey sandstone, and the remainder of the outcrop showing pink, red, orange and salmon colors often seen in the Potsdam used as building stone.

The colors here are due to finely crystalline hematite, goethite and anatase (=leucoxene) which form pigments around the detrital grains and within secondary quartz and illite cements. The iron and titanium for these pigments were derived from the breakdown of detrital magnetite and ilmenite grains in the original sediment. Paleomagnetic studies at this outcrop suggest that the iron-bearing minerals were precipitated in late Paleozoic time. The white sandstones at the base of the Potsdam contain no hematite or anatase, although limonite-goethite halos of relatively recent origin are developed around some magnetite grains.

Note the alteration of the underlying gneisses immediately beneath the Potsdam. The alteration assemblage here consists of illite, Fe-chlorite (or berthierine), siderite and quartz.

Continue southwest on Route 12.

48.8 1.2 Stop 6

STOP 6:

The roadcut on the southeast side of Route 12 exposes typical lower Potsdam Sandstone. Plane-bedded medium- and fine-grained sandstones underlie and overlie a one meter thick bed of cross-bedded medium sandstone. The dominant cross-bed dip direction is SSW. Above the thick cross-bedded unit small-scale cross-strata dip to NNE. As with many exposures of the lower Potsdam, the environment of deposition is difficult to assign. Although the flat-bedded sandstones lack trace or body fossils, a shallow marine environment is suggested by the continuity of individual beds, the lack of upward fining or coarsening trends, and the lack of channel form geometry. However, definitive primary structures are absent. The thick cross-bedded unit was produced by a bedform of perhaps two or three meters in amplitude. Stoss-side erosion beveled the upper portion of the structure as it migrated, leaving behind a scoured lag deposit of granules at the updip termination of the cross-strata. A structure of this size and geometry could be produced by tidal currents or by wind during a period of emergence.

Continue southwest on Route 12.

50.2 1.4 Stop 7

STOP 7:

The unconformity between the basal Potsdam Sandstone and Proterozoic gneisses is again exposed in these large roadcuts on both sides of Route 12. The dominant facies here is typical of the lower portion of the Potsdam. Flat-stratified medium- and fine-grained sandstones are interrupted by 0.2 to 1.0 meter thick sets of cross-strata. The lack of trace and body fossils again makes assignment of a depositional environment somewhat difficult.

Note the altered gneiss immediately beneath the contact. Illite, Fe-chlorite, and siderite form the alteration assemblage. The sandstones immediately above the contact contain abundant authigenic illite, which has been K-Ar dated at 355-360 million years.

Continue southwest on Route 12

56.0	5.8	Proceed north on Interstate Route 81
64.3	8.3	Optional Stop 8

STOP 8 (optional stop, time permitting):

These long roadcuts on Interstate Route 81 expose a sequence of conglomerates and pebble-cobble sandstones in the Potsdam Sandstone. The quartzite cobbles in these exposures were derived from a basement quartzite ridge immediately to the south. At least some of the conglomerate beds here were emplaced by mass flow mechanisms, perhaps as sandy debris flows. Wave reworking of the tops of mass flow beds indicates that the sands and gravel were deposited in a marine setting. Near the top of the exposure the contact between the lower and upper Potsdam is present, and burrowed calcareous sandstones succeed the conglomeratic facies.

END OF TRIP.

(To return to Interstate Route 81 South, exit at the DeWolf Point exit approximately three miles north of the last stop. Turn left off the exit ramp to head south on I81)

TRIP B-4: SOME CLASSIC MINERAL COLLECTING SITES IN ST. LAWRENCE COUNTY

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Introduction

The St. Lawrence County Rock and Mineral Club is a local organization who meets once a month and organizes collecting field trips as well as trips to rock and mineral shows. We have members from as far away as New Hampshire, Ohio and Canada. All meetings and trips are open to the public. It is our pleasure to welcome you to Canton and to lead you on this field trip of classic mineral localities in St. Lawrence County.

If you would like more information about our club for attending some of our meetings or trips, you may write to our club president Gary Stacy, 148 Rowley St., Gouverneur, NY 13642, vice-president Michael Whitton, 24 Miner St., Apt. 1, Canton, NY 13617, or Schuyler Alverson at the above address. Michael will be along on this trip as one of the principal guides. He will be happy to take a business card or take down your name and address to include you in the club's next newsletter.

We will be departing Canton, from St. Lawrence University at 8:30 AM on Sunday, September 26. Our first stop will be at Benson Mines in Star Lake. At one time, this was the largest open pit magnetite mine in the world. Our second stop will be in West Pierrepont at a tremolite locality. Our last stop will be the world famous Powers' Tourmaline Locality in Pierrepont. This site has been known for over 100 years since George Kunz wrote about it in 1892. We will learn more about each stop later. At the end of this road log will be a list of suggested readings for these stops if you would like to find more information about them.

We will be eating en route to our second stop to save time. We will spend a good deal of the morning travelling to Star Lake and back. You may eat at any time you wish, but we want to be sure to have everyone back to Canton on time so that they will not miss their rides or their scheduled time of departure. The map on the next to last page of this trip log shows our route and the stop locations.

ROAD LOG FOR SOME CLASSIC MINERAL COLLECTING SITES IN ST. LAWRENCE COUNTY

Cumulative Mileage	Miles From Last Point	Route Description
0	0	Begin the trip at SLU's J-Lot. Turn left onto Park Street.
3.3	3.3	Take the left at the Y.
6.1	2.8	Take the right at the Y and climb up Waterman Hill toward Degrasse. There will be a radio tower in a field on the right on top of the hill. Notice the vista over Canton behind you.
11.2	5.1	This intersection is known as the Turnpike Crossing. If you'll notice the tavern on the corner you will understand why. Proceed straight across this intersection, but don't forget it. On the return from Benson Mines we will be turning here for the second stop.
21.5	10.3	Turn left on County Route 77 toward Fine.
29.3	7.8	Turn right onto County Route 27 to Route 3 where we turn left toward Star Lake. Staying on Route 3, go through the village and 1.5 miles more.
38.2	8.9	Turn left just before the large blue buildings and across the road from the St. Lawrence County Solid Waste Disposal Authority's Waste Transfer Site. This is the mine entrance. We will proceed through the gate, pass the mill buildings and toward the flooded, open pit mine. We then head for the mine dumps to do some mineral collecting.

STOP 1. J & L STEEL CORPORATION, BENSON MINES, STAR LAKE, NY

History of the Mines

The ores were first discovered in 1812 when a military road was being built in the region. It became known as the Chaumont Ore Bed. In 1883 the Magnetic Iron Ore Co. was formed. Byron D. Benson, a large land owner in southern St. Lawrence County sells the company the mineral rights on 2,201 acres of the Brodie Tract in the town of Pitcairn. In 1886, a railroad is started from Carthage to Jayville in the town of Pitcairn and additional mineral rights on 40,000 acres in the southern part of the county are obtained by the Magnetic Iron Ore Co.

In 1887 and 1888, the company purchased minerals rights for magnetite iron ore on Vrooman Ridge in the Town of Fine and all the mineral rights (9185 acres) in the southeast corner of Chaumont Township (later part of Clifton Township) including the Chaumont Ore Bed. In 1888-89, work was done on the Jayville mines in the town of Pitcairn. From 1889-93 the Magnetic Iron Ore Co. shipped 150,000 tons of high grade concentrates (magnetite iron ore) to Pennsylvania.

In 1906, Benson Mines Company formed and leased Benson Mines (Chaumont Ore Bed). From 1907-1919, sporadic mining occurred at Benson Mines and in 1922 the Benson Mines Co. gave up its lease and sold its plant to the Magnetic Iron Ore Co., owner of the mineral rights.

Later in 1922, the Benson Iron Co. Inc. was formed, but did not do much mining. In 1941, Jone & Laughlin Ore Co. leased Benson Mines and built a $7,000,000 mill and upgrade of the mine. In 1946, the Benson Iron Co. Inc. and Magnetic Iron Ore Co. (both belonged to the Benson family) consolidated to form the Benson Iron Ore Corporation, and by 1950, it was the largest open pit magnetite iron ore mine in the world.

In 1978 Benson Mines shut down. The St. Lawrence County Development Corporation tried to find a buyer or lessee without success. Lumbering is the only thing going on in the mine area now that is bringing in any money, but just barely enough to pay the taxes. Of the 3,200 acres now owned by the company, only 1,200 are cut in a 15 year cycle to that allows for replenishment. There is some sale of waste rock, but the income from this is negligible. The mine used to employ 1,000 people and pump 2,000,000 gallons of water a day.

Geology

The Benson Mines ore body is not of simple origin. Sillimanite gneiss, metagabbro and pegmatitic units are located throughout the pits. Small, localized fault zones rich in secondary mineralization provide the areas of greatest interest to the crystal collector.

Minerals

The primary ores mined here are magnetite and martite. Samples of each abound. The sillimanite crystals from here are unparalleled in their size. The sunstone, relatively common in the pegmatitic zones, may be fashioned into cabochon and used to make attractive jewelry. Probably the most desired species from this mine is the dark green fluorite cubes, found about 30 years ago in a fracture zone and associated with various other secondary minerals. These, needless to say, are only rarely found. The following minerals can usually be collected, many in crystal form: aragonite, azurite, bornite, calcite, chalcopyrite, chlorite, garnet, hematite, hornblende, magnetite, malachite, martite, microcline, molybdenite, muscovite, pyrite, quartz, sillimanite, sunstone, tourmaline, and others.

The tools that will be most helpful here are a crack hammer, chisel and small pick. Be sure to bring something (sample bags, boxes) to store your specimens in on the trip.

Notes

Collecting is by clubs or groups only. Permission may be obtained and arrangements made by writing in advance. The person to talk to, his address and phone numbers are listed here. He will send you a release form for everyone to sign and return before going to the mine. Please respect the common courtesies expected of you.

David H. Ackerman
Benson Mines, Inc.
100 Bay Street
Glens Falls, New York 12801
(518) 523-9757 or (201) 267-3306

Cumulative Mileage	Miles From Last Point	Route Description
38.2	0	Back down to the gate we came in. There will be some extra mileage from driving around the mine and the dumps that is discounted here. From the gate take the same route back to the Turnpike Crossing.
65.2	27	At the Turnpike Crossing, turn right onto County Route 24.
66.7	1.5	Turn right, just after the sluice pipe bridge in West Pierrepont, onto the Selleck Road.
67.7	1	After driving one mile from County Route 24, look for a narrow, dirt road on the right. All of us may not be able to drive into the site and turn around to come back out. We may have to car pool to get everyone in or some will have to hike and leave their vehicles on the road. It is only 0.2 miles in, but uphill. Once we get to the top of the hill, we park and start digging. The site lies on either side of the road. You may start collecting anywhere along the top or front side of the hill.

STOP 2. TREMOLITE LOCALITY, WEST PIERREPONT, NY

History of the Site

This site has also been known for many years. It was discovered by prospectors looking for iron ore. Ours was one of the first rock and mineral clubs to visit the site in 1965. Originally, the amphiboles were called actinolite. However, Dr. George Robinson identified them as tremolites by electron probe.

Geology

Tremolite-actinolite crystals are commonly found in the calcium silicate rocks associated with the county's Grenville Marble deposits. The rock here appears to be a tremolite-quartz schist, with varying amounts of calcite, diopside and pyrite. A large body of leucogranitic rock nearby, if representing an igneous intrusion, may have provided the necessary physiochemical conditions to recrystallize the adjacent sediments.

Minerals

Doubly terminated crystals of tremolite-actinolite occur in pockets and fissures throughout the outcrop. Other nearby outcrops have furnished dravite, diopside, pyrite and quartz crystals.

Notes

The best crystals are in the cracks and crevasses in the ledges. Hammers, chisels, crowbar and small digging utensils will prove useful.

Cumulative Mileage	Miles From Last Point	Route Description
67.9	0.2	Starting from the top of the hill where we parked, we will start heading out to the Selleck Road.
68.1	0.2	Turn left onto Selleck Road.
69.1	1.0	Turn right onto County Route 24.
73.4	4.3	Intersection with Route 68 in the hamlet of Pierrepont. Go straight.
74.4	1.0	There will be a bridge, the Irish Settlement Road on the right and a gravel road on the left after the bridge. Take the gravel road.
75.9	1.5	The gravel road has been nearly taken over by brush since the bridge over Grannis Brook was taken out. Just before the bridge site there is a small clearing on the right. Pull in here and park.

STOP 3. POWERS' FARM, PIERREPONT, NY

The collecting site is just a short hike along the brook. Stay on the farm lane until you come to a path. Turn left onto the path and twenty-five yards or so ahead is the collecting area.

History of the Site

As stated above, George Kunz wrote about this locality in 1892. It was known to collectors for some years before that. It was originally the Ryland Clary Farm. An aggressive mineral collector by the name of Charles D. Nims would drive his horse and wagon to the site from Canton where he worked for the railroad for 10 year. He would remove baskets full of crystals.

Around 1910, Bower Powers, Sr. bought the land. In 1921, William Agar wrote about the occurrence. Just prior to 1962 the top of the hill had been bulldozed. In 1993, due to littering, people sneaking in, burning fires and digging on the hillside next to the stream, Bower Powers, Jr. closed the site to individuals. It is still possible to arrange in advance club or group trips as long as you abide by the rules. There are still many nice crystals to be found.

Geology

The geology of this PreCambrian Grenville site is very complex. The interesting minerals occur in veins and along contacts between the Grenville marble and other metamorphic rocks including a mica-tourmaline-quartz schist, an amphibolite, and other rocks locally rich in pyroxene or scapolite. Additional evidence of an origin through recrystallization is provided by the frequent presence of uralite, a pseudomorph of actinolite after diopside. The field relationships are buried by overburden. The mineral paragenesis is complicated and not worked out.

Minerals

The uvite tourmaline crystals found here are indeed world famous. They contain a relatively high percentage of iron. These dark, magnesian tourmalines are typified by their lack of prismatic striations and short c axes. Figure 1 below illustrates the most commonly encountered forms: trigonal prism (m), ditrigonal prism (a), rhombohedra (e) and (r), basal pedion (c), and trigonal pyramids (o). Other minerals to be found here in excellent representations include: uralite, rennselaerite, diopside, pyrite, pyrrhotite, quartz, apatite, micas, chlorite, calcite, actinolite, titanite and scapolite.

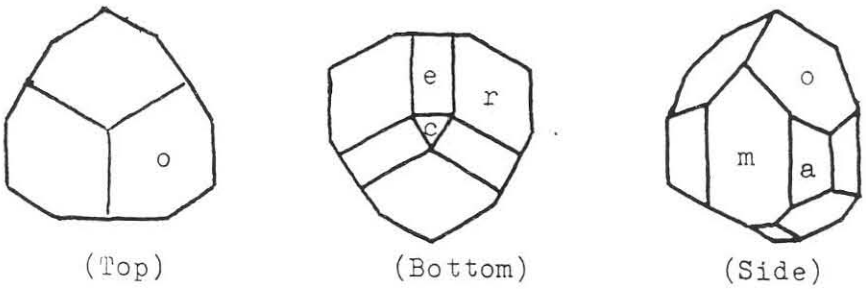
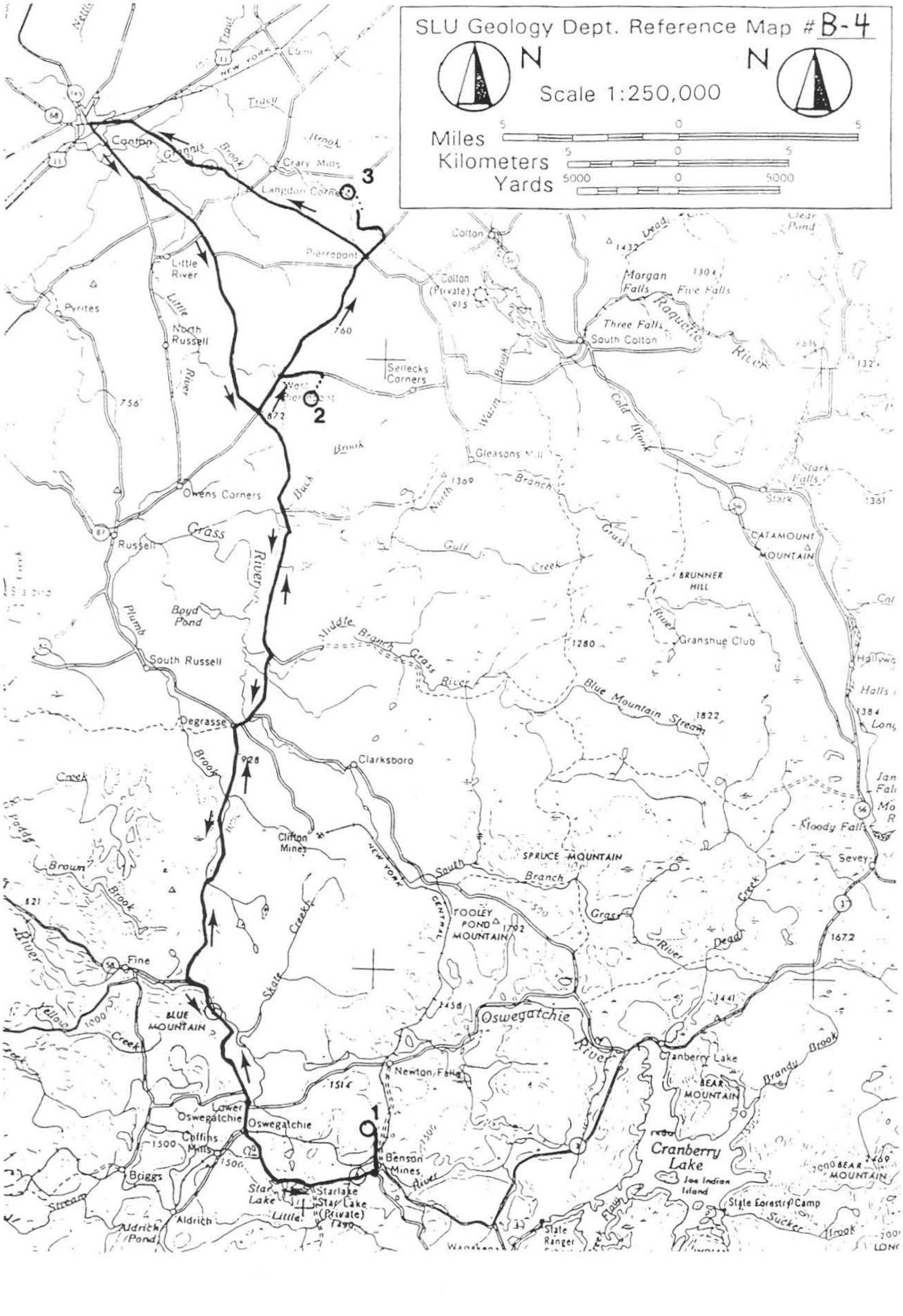


Figure 1. Hemimorphic Tourmaline Crystal

Notes

Although small single crystals may be found in the soil, the better specimens lie deeper and occur most often in pockets and seams in the solid tourmaline-quartz rock, especially where it is in contact with the calcite. Often, good specimens can be obtained by removing the calcite from the matrix rock with dilute hydrochloric acid.


SLU Geology Dept. Reference Map #B-4
N
N
Scale 1:250,000
Miles
Kilometers
Yards
Canton
Pierrepont
Colton
Little River
North Russell
Pyrites
Sellecks Corners
South Colton
West Pierrepont
Owens Corners
Russell
Grass River
South Russell
Degrasse
Clarksboro
Clifton Mines
Fine
BLUE MOUNTAIN
Oswegatchie
Lower Oswegatchie
Newton Falls
Benson Mines
Star Lake
Cranberry Lake
Oswegatchie River
Briggs
Aldrich
1
2
3

Useful tools for this location include shovels, crowbars, sledge hammers, wedges, and other heavy tools will be needed to reach the deeper crystal bearing areas, but hand tools will also be useful.

Cumulative Mileage	Miles From Last Point	Route Description
75.9	0.0	Return to the vehicles from the collecting site. Load up everything and turn around to head back out to the four corners in the hamlet of Pierrepont.
78.4	2.5	At the intersection with Route 68, turn right and head back toward Canton.
85.4	7.0	At the intersection with Route 11, turn left, drive into town.
87.5	2.1	After turning onto Route 11, you come right into Canton, turn left at the first traffic light onto Park Street and head back the starting point at St. Lawrence University.

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THE POTSDAM COLLEGE SEISMIC NETWORK

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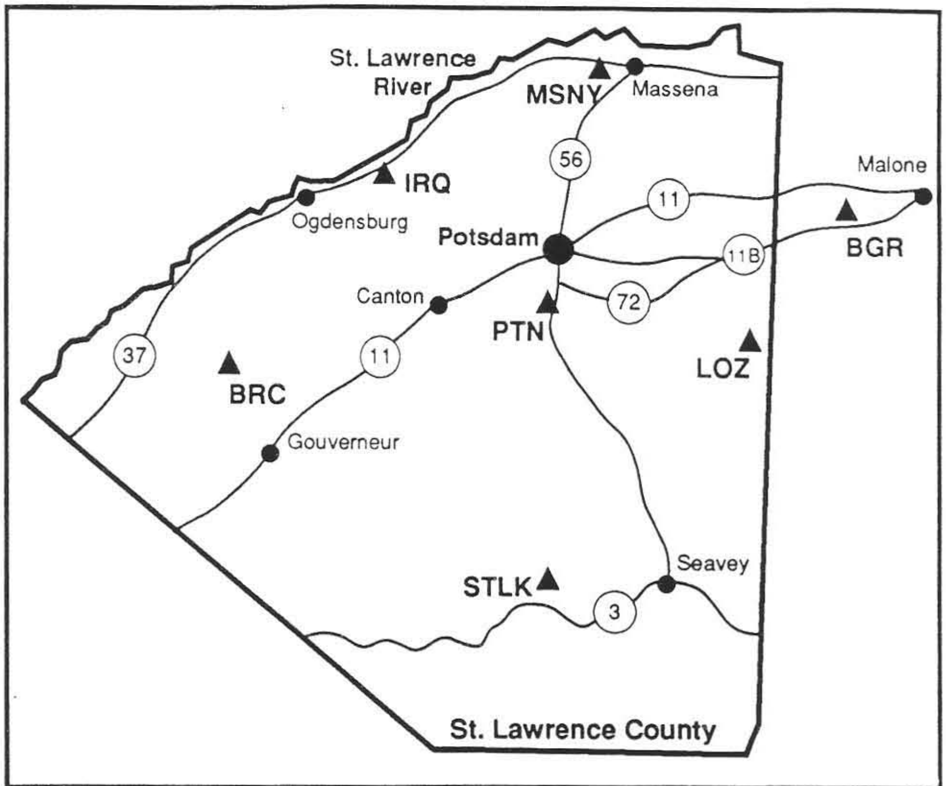
INTRODUCTION

The Potsdam Seismic Network consists of seven short period-vertical seismograph stations located in the St. Lawrence Valley in northwestern New York. All seven of the seismic field stations transmit the seismic signals to Potsdam College via FM narrow-band telemetry except the station at Star Lake (STLK) which is relayed through the Brasie Corners (BRC) station. The earthquakes recorded by the network are located by using two hypocentral location programs FASTHYPO (Hermann 1979) and HYPOINVERSE (Klein 1978) run on a Zenith personal computer. Analog data from the field stations is digitized and the digitized seismic signals fed directly into the SUN computer where the digital seismogram may be observed on the computer monitor. The program SUNPICK is used to pick the P and S phases and determine the arrival times used to locate the hypocenter.

During the past 5 years forty-eight local earthquakes have been recorded by the network. Most of the epicenters of the earthquakes are located in the St. Lawrence Valley just south of the Massena-Cornwall area. No surface evidence of any major fault is in this area nor has any noticeable ground displacement been observed however two faults of the Ottawa-Bonnechere graben, the Winchester Springs and Gloucester faults are inferred to cross the St. Lawrence River into the United States. The area also contains a possible extension of the Carthage-Colton Mylonite Zone, a major structural boundary that separates the Adirondack Highlands from the Lowlands. Most of the epicenters are located along the possible extension of the CCMZ and in the vicinity of the Gloucester and Winchester Springs faults.

HISTORY OF THE SEISMIC NETWORK

The Potsdam College Seismic Network consists of seven short period-vertical seismograph stations (Figure 1) located in the St. Lawrence Valley in northwestern New York. The first station (PTN) was installed in October 1971 as a joint venture between Potsdam College and the Lamont-Doherty Geological Observatory (LDGO) of Columbia University. LDGO installed the seismic field station sixteen kilometers south of Potsdam and telemetered the seismic signals to Potsdam College where they were recorded on a seismograph jointly purchased by the Potsdam College and Alcoa Foundations. The seismic signals were transmitted by telephone lines to LDGO at the Palisades, New York, for study



POTSDAM COLLEGE SEISMIC NETWORK

Seven short-period vertical seismograph stations in the St. Lawrence Valley

▲ BGR	Bangor
▲ BRC	Brasie Corners
▲ IRQ	Iroquois Dam
▲ LOZ	Lake Ozonia
▲ MSNY	Massena
▲ PTN	Potsdam
▲ STLK	Star Lake

Figure 1: Potsdam College Seismic Network

by the LDGO seismologists. This was the first seismograph station installed in northern New York.

During the succeeding years six more seismograph stations were installed in the area. In 1976 LDGO installed seismic field stations at the Long Sault Dam at Massena, and at Bangor, New York. The Gulf and Alcoa Foundations provided Potsdam College with grants to purchase seismographs to record from these two field stations. From 1983 to 1988 Potsdam College purchased seismic equipment to install three stations at Lake Ozonia (LOZ), Star Lake (STLK) and Brasie Corners (BRC). The funds for the installation of these stations were received from the New York State Power Authority and Alcoa Foundations. In 1988 Plattsburgh State College donated one seismograph to Potsdam College.

The Lamont-Doherty Geological Observatory of Columbia University provides much equipment and services for the operation of the network and is connected through internet to the seismic signals recorded by the network. The information from the network is also provided to the Geophysics Division of the Canadian Geological Survey so they may locate Canadian epicenters more accurately and calculate fault plane solutions.

In addition to detecting, recording and locating local earthquakes Potsdam College maintains a catalog of historical and recent seismicity in the northeastern United States and southeastern Canada. This catalog contains 3159 seismic events in the area between 39^{o} to 53^{o} north latitude and 58^{o} to 81^{o} west longitude and lists all the earthquakes that have occurred during the period 1534 through 1986. This earthquake catalog, published by Nottis (1983), serves as the core for the New York State Earthquake Hazard Reduction Program.

THE SEISMIC FIELD STATIONS

The detection of earthquakes occurs at the seven seismic field stations shown in Figure 1. All the seismometers are vertical short-period units which respond to the short-period seismic waves generated by local earthquakes. The seismometer is placed on bedrock inside a bottomless 55 gallon steel drum. The ground motion produced by the seismic waves is converted to an electrical signal and fed into a high-gain amplifier to increase the signal amplitude. This signal is frequency modulated by a voltage-controlled oscillator (VCO). The VCO frequency is now in the audible range and its output frequency changes in response to the change sensed by the seismometer. The signal is modulated to the FM radio transmitter and then transmitted to the receiving station at Potsdam College. The power output of the transmitter is 100-350 milliwatts. The frequency is usually in the UHF government experimental band.

Presently the network is converting its seismic field stations from batteries to solar energy. The solar-powered station consists of a silicon solar collecting panel that provides DC current to the regulator during the daylight hours. When sufficient current is available, a regulator recharges a battery. The battery supplies the energy that is needed to power the radio transmitter and amplifier. This results in significant financial savings.

THE SEISMIC RECORDING STATION

All seven of the seismic stations transmit the seismic signals to Potsdam College via FM narrow band telemetry except the station at Star Lake (STLK) which is relayed through the Brasie Corners station (BRC). Antennas and receivers on Raymond and Timerman Halls pick up the signals and feed them into discriminators. The discriminator removes the FM carrier wave and feeds the signal into the amplifier for amplification. The output from the amplifier is recorded on the helicorder or seismograph. The output from the discriminators also enters a SUN Computer which serves as an analog to digital converter.

The entire system is "locked on" to a satellite receiver which continuously monitors a GOES satellite which transmits time signals from the National Bureau of Standards at Washington, D.C. The clock displays the number of the day of the year, and hours, minutes and seconds in Universal Coordinated Time (UTC). Its accuracy is always within 12 milliseconds of true UTC time.

The Potsdam College Seismic Network short-period seismometers are best suited for the detection of local earthquakes. Local earthquakes are those with epicenters within 1000 kms from the seismic station. However, distant earthquakes (teleseisms) may also be recorded provided the magnitude of the earthquake is greater than 5 on the Richter Scale. Several teleseisms such as the Armenian, Loma Prieta and the more recent Landers, California earthquake have been recorded during the past few years. The teleseisms are too distant to locate the epicenter accurately by our closely spaced regional seismic network. Information about the teleseism is obtained quickly from the National Earthquake Information Center. Quick Epicenter Determinations (QED) are available to users having access to a modem and microcomputer by dialing the toll-free number 800-358-2663. The information provided by the NEIC and our seismogram of the quake are an excellent combination to understand and analyze the earthquake.

SEISMOGRAM ANALYSIS

The earthquakes recorded by the Potsdam Seismic Network are located by using the two hypocentral location programs FASTHYPO and HYPOINVERSE. The programs are run on a Zenith 159 computer with a hard disk and a SUN SPARC Workstation. These programs determine the location, depth and origin time by minimizing the difference between the observed and calculated travel times of the seismic phase arrivals for a specific crustal velocity model. For a particular earthquake the model most appropriate for the epicentral region of the quake is used. The arrival times of seismic phases are used as input to the computer programs. The times are read from the records of the network or are picked from the monitor screen of the SUN Workstation. A data sheet used to record seismogram measurements and input them into the computer is shown in Figure 2.

The magnitudes (Mc) of local earthquakes are calculated by using the signal duration (coda length) formula developed for New England (Chaplin 1980). An average magnitude

DATA SHEET (FASTHYPO on Zenith)

DATE OF EVENT ______________________

Hours and minutes of event (UTC) ______________________

Type FAST when you see C

Type ø for geographic

Type 1 for Northern NY - Adirondack Model

Type 1 to input data at the terminal

Enter a title (up to 80 characters) ______________________

STA (capitals)	P Descrip.	P-WT	TP (Secs)	S Descrip.	S WT	TS (Secs)
PTN	IC '	'	'	E '	'	'
LOZ	IC '	'	'	E '	'	'
BRC	IC '	'	'	E '	'	'
STLK	IC '	'	'	E '	'	'
BGR	IC '	'	'	E '	'	'
MSNY	IC '	'	'	E '	'	'
IRQ	IC '	'	'	E '	'	'

When finished with station data, type STOP (capitals)

Figure 2: Data Sheet showing input to computer to locate hypocenter

value based on all available station observations is reported for each earthquake. The signal duration is the time from the P-wave arrival until the coda amplitude disappears into the background noise. The formula used in magnitude determination is:

$$Mc = 2.21 \text{ Log } T - 1.70,$$ where T is the signal duration in seconds.

An example of a calculation of the magnitude of the aftershock of the Goodnow Earthquake of October 7, 1983 is shown in Figure 3.

SUN WORKSTATION

The Potsdam Seismic Network contains a SUN SPARC Workstation for analysis of the earthquakes detected by the network. The analog data from the network is digitized, and the

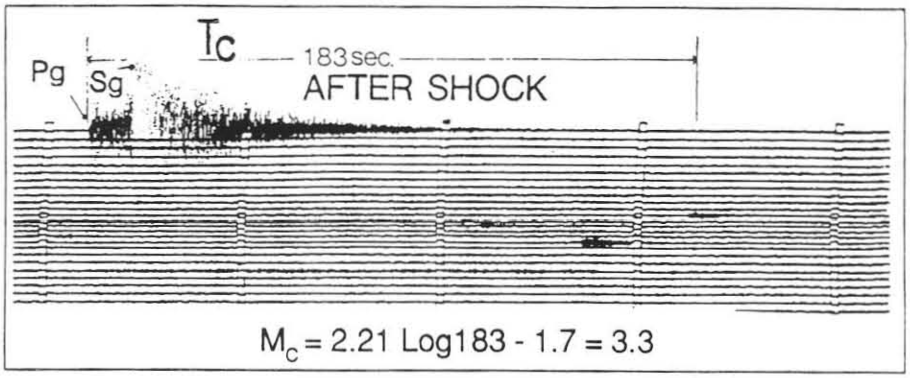


Figure 3: Seismogram showing magnitude calculation

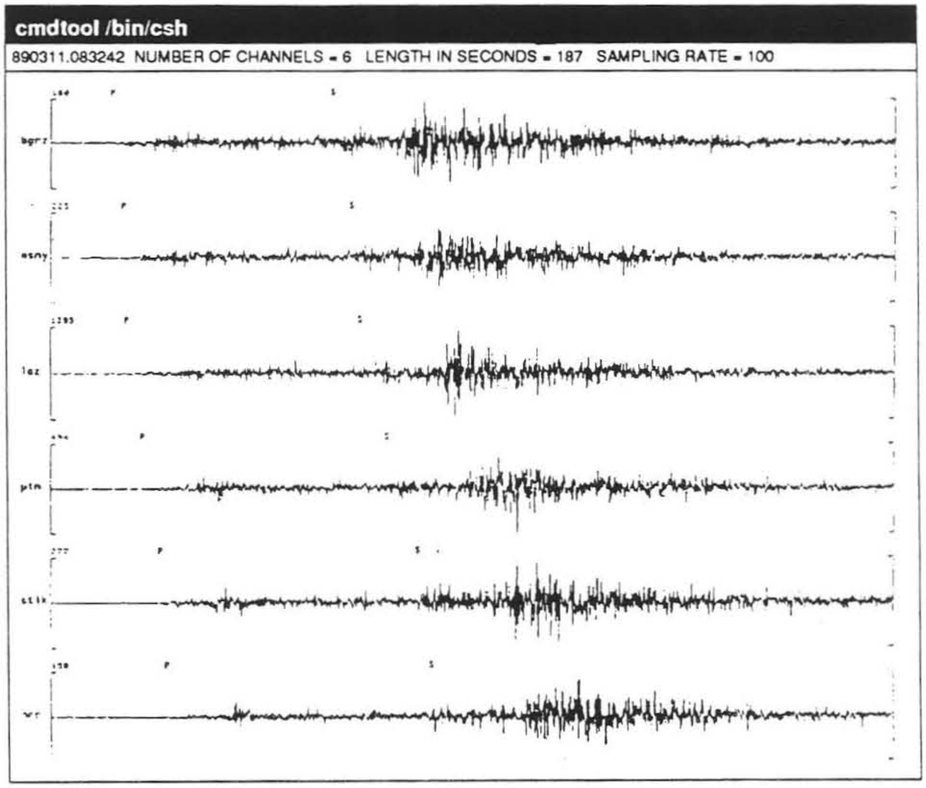


Figure 4: Digital seismogram of Charlevoix quake as seen on the SUN computer

digitized seismic signals are fed into the SUN computer for study. A typical digital seismogram of an earthquake with its epicenter located in the Charlevoix Region, Quebec, Canada, is shown in Figure 4. This earthquake occurred on March ll, 1989, at 08h:32m:42s UTC, so it is stored under the file number 890311.083242. The advantages of using the digitized seismogram are that the earthquake record may be modified so arrival times of P and S phases may be measured to l/100 of a second, and the record may be enhanced to show the

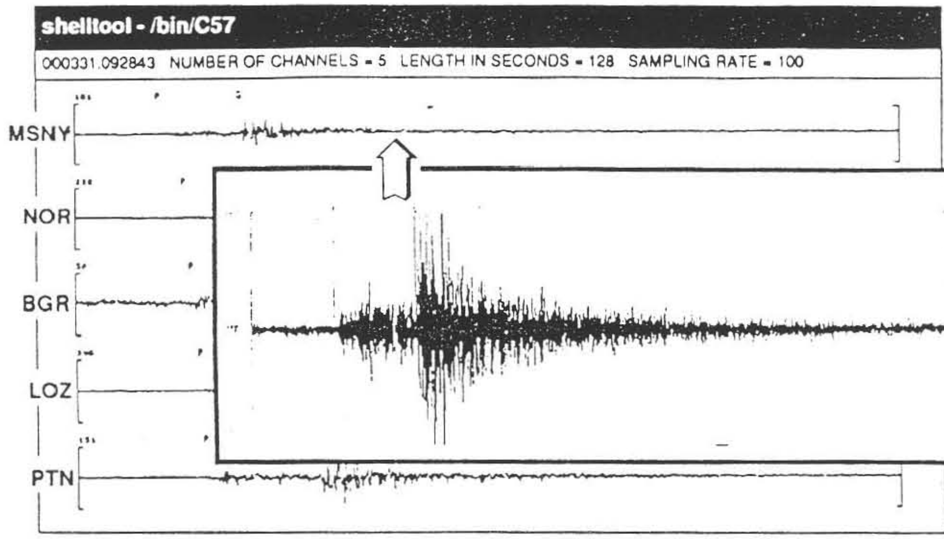


Figure 5: Digital seismogram of the earthquake recorded by the Massena (MSNY), New York stations. The enhanced output is shown in the inset

P and S phases more clearly. Figure 5 shows a local earthquake recorded by the Massena (MSNY) station before and after enhancement with the SUN computer. The program SUNPICK is used to pick the P and S phases and determine their arrival times. The arrival times of the P and S phases are used as input to HYPOINVERSE to locate the hypocenter of the earthquake. A printout of the seismic event is obtained by using a laserwriter configured with the SUN computer.

INTENSITY STUDIES

Intensity studies are conducted of all local earthquakes using the Modified Mercalli Intensity Scale. United States Geological Survey Earthquake Questionnaires are distributed to residents in the area through local newspapers. Evaluation of the responses from the residents enables the intensity at particular sites to be determined with the Modified Mercalli Intensity Scale. The intensity values are plotted and contoured to construct an isoseismal map of the earthquake.

SEISMOTECTONIC RELATIONS AND DATA

The Potsdam College Seismic Network is located in the heart of the most active seismic area in New York State. This belt of earthquakes is known as the Northern New York-Western Quebec seismic zone (Figure 6). Historic earthquake data indicate that the earthquake activity in this area has been persistent for over 400 years (Smith 1966). The largest earthquake in New York State occurred in this zone at Massena, New York, on September 5, 1944. This earthquake of Intensity VIII caused over $10,000,000 (current

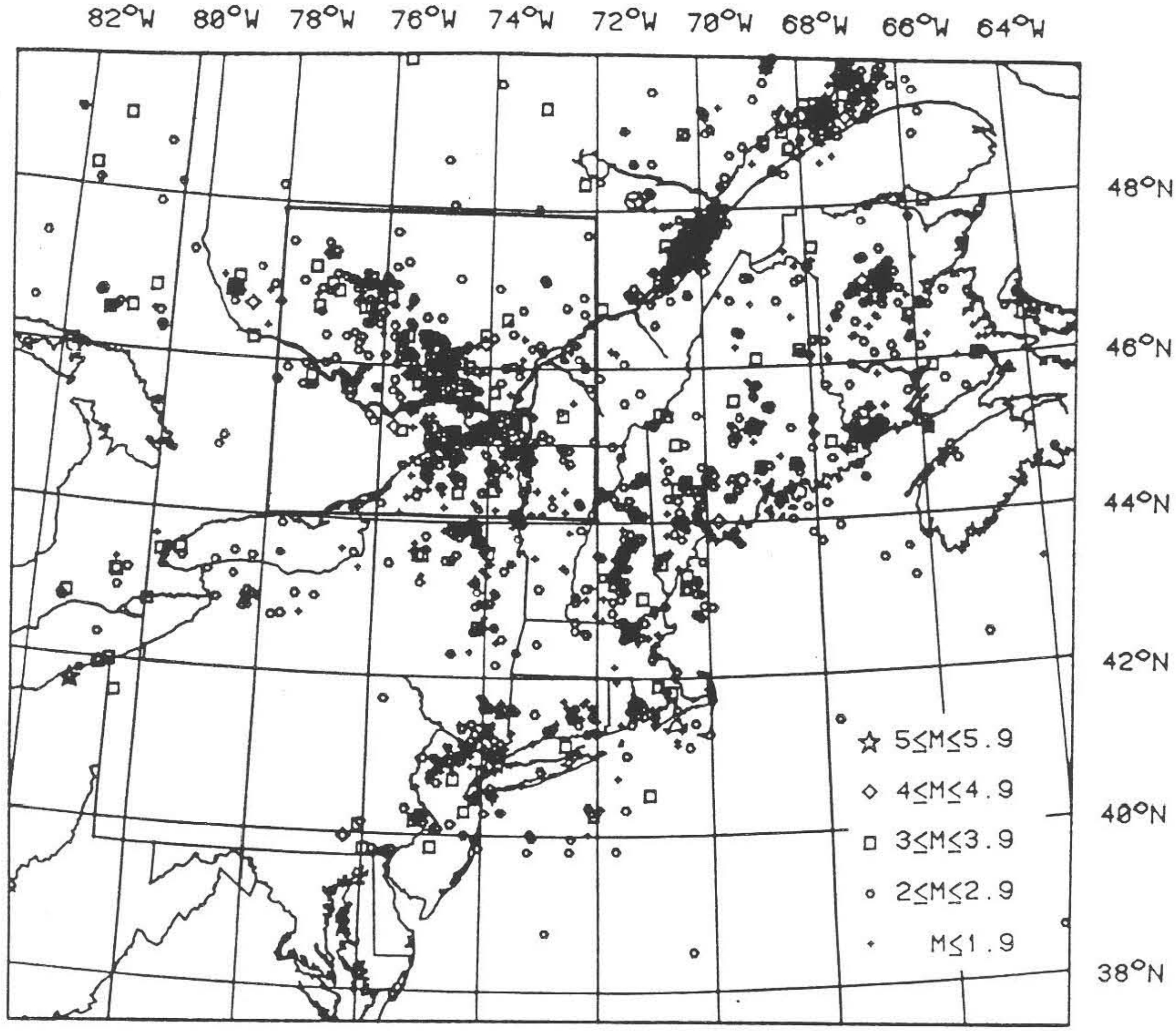


Figure 6: Earthquake epicenters in northeastern United States 1975 - 1989 (From Northeastern U.S. Seismic Report March 1992) (Northern New York - Western Quebec seismic zone is outlined)

dollars) in damage in the Massena-Cornwall area. The earthquake destroyed 90% of the chimneys in the Massena area and did extensive damage to schools and other buildings. It is likely that another damaging earthquake could occur in the area.

More recently, early in the morning of October 7, 1983, an earthquake of magnitude 5.2 and maximum Intensity VII on the Modified Mercalli Scale shook all of New York State and adjacent areas. The epicenter of this earthquake was located 20 km northeast of Blue Mountain Lake near Goodnow Mountain so it is known as the Goodnow Earthquake. Millions of dollars in damage would have resulted from this earthquake if its epicenter were located in an urban area.

Table 1. Distribution by zone of all known earthquake New York State as a function of size (1720–1980).

	Number of Events				Total
Zone	1	2	3	4	NY State
III	12	113	31	15	171
IV	12	66	17	11	106
V	2	23	13	3	41
VI	3	6	2	0	11
VII	0	2	2	0	4
VIII	1	1	0	0	2
TOTAL	30	211	65	29	335

Magnitude and Intensity

The intensity and magnitude of an earthquake are not the same. Intensity *refers to how much damage was caused, as reported by the public; it is measured by the Modified Mercalli Scale. The intensity decreases as one moves farther from the epicenter.* Magnitude *is a relative measure of the true size of the earthquake in terms of how much energy is released. It is determined by the amplitude or period of the waves and is the same at different seismic stations.*

Figure 7: Shaded area indicates number of earthquakes of Intensities III through VIII occurring in the northern New York seismic zone

The number of earthquakes in New York State in various source regions is shown in Figure 7. These data show clearly that the northern New York area is most active seismically with 211 seismic events of a total of 335 for New York State. Diment, Urban and Revetta (1972) and Sbar and Sykes (1973) originally suggested that this seismic zone was part of a larger belt of seismicity extending from Boston, Massachusetts through Ottawa into Kirkland Lake, Ontario. Most of the earthquakes recorded by the Potsdam Seismic Network during the past 5 years have their epicenters located in this zone in the Cornwall, Ontario, Massena, New York area.

Table 1 shows a list of forty eight earthquakes recorded by stations in the network during the past five years and Figure 8 shows the distribution of these epicenters in the Cornwall-Massena area. The parameters of these earthquakes were determined from the Geophysics Division of the Canadian Geological Survey, the Northeastern United States Seismic Network Reports and the Potsdam State Seismic Network. The earthquake foci in this area are located well within the Precambrian basement rocks and fault plane solutions indicate reverse faulting along NNW or NW striking fault planes (Schlesinger-Miller, 1983). No surface evidence of any major fault is in this area nor has any noticeable ground displacement been observed (Berkey, 1945).

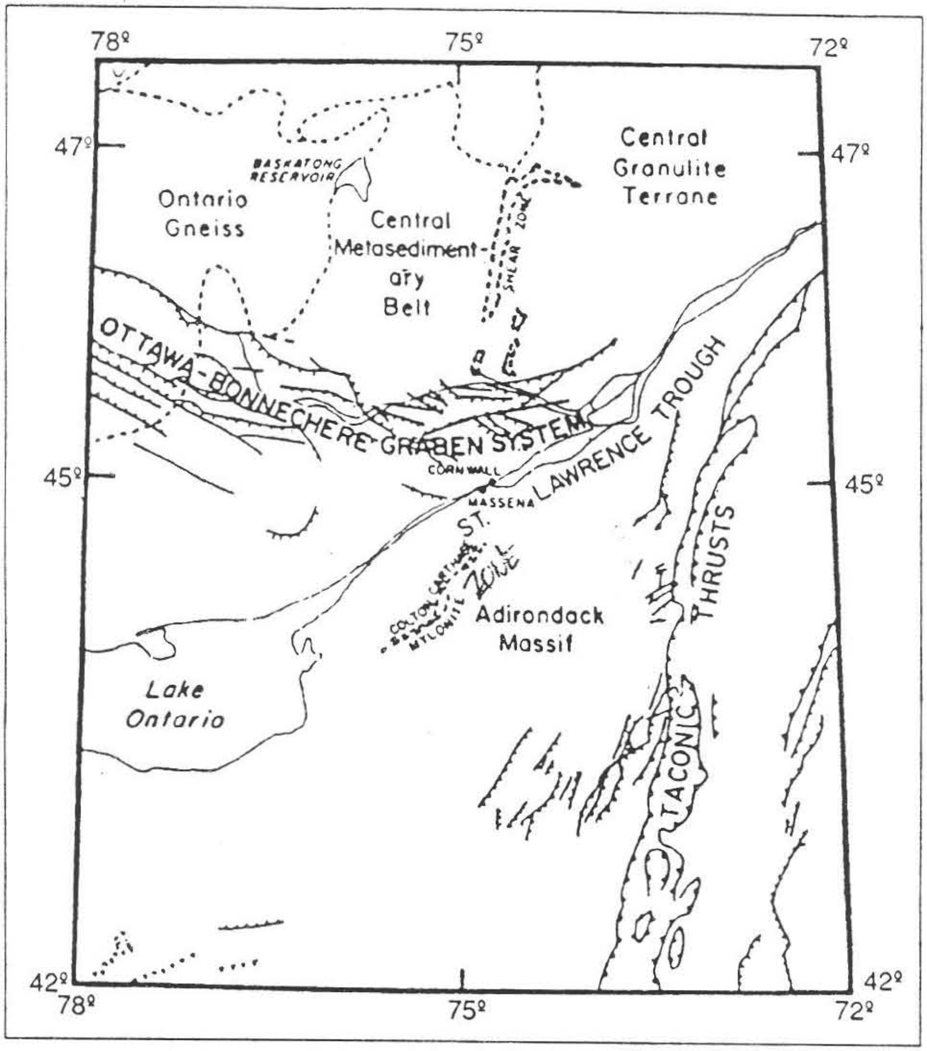


Figure 9: Generalized reference map from Tectonic Map of the United States (1944, 1962). Precambrian province boundaries are from Forsyth (1981).

The distribution of epicenters, both recent (1988-1992) and historical extends in an east-west or east-northwest direction. Comparison of the distribution of recent seismicity with historical earthquakes indicates the recent earthquakes occur in the same general area as the historical seismicity. The coincidence of recent and historical seismicity suggest some local geologic feature causing the earthquakes in the area. Two faults of the Ottawa-Bonnechere graben, the Winchester Springs and Gloucester faults are inferred to cross the St. Lawrence River into the United States in this area (Weston Geophysical 1985). Evidence of these fault extensions is based on limited outcrop and landform data correlated with similarity of VLF conductivity signatures over known faults in Canada. The earthquake epicenters in the

Another consideration is whether the earthquakes in the area are related to the Carthage-Colton Mylonite Zone. The CCMZ is a major structural boundary between the Adirondack Highlands and Northwest Adirondack Lowlands. This zone may continue northward under

Table 1 (Earthquakes recorded by Potsdam Seismic Network (1988 - 1992)

Date	Latitude North	Longitude West	Origin Time UTC	Magnitude	Comments
01/30/88	44°53.50'	75°3.23'	01:06:6.55	1.0	Louisville, N.Y.
02/12/88	44°42.44'	74°39.47'	11:41:34.14	1.7	Nicholville, N.Y.
02/13/88	45°12.10'	75°7.66'	02:11:35.93	2.1	Chrysler, Ontario, Canada
03/24/88	44°54.71'	73°25.47'	02:12:24:30		Chazy, N.Y.
04/02/88	44°32.64'	75°10.43'	10:00:22.33	1.9	Pyrites, N.Y.
04/26/88	44°57.00'	72°37.20'	19:42:10.20	2.3	U.S.A.-Quebec border
05/02/88	44°56.31'	74°48.86'	06:34:51.37	2.0	Helena, N.Y.
05/15/88	45°9.00'	75°36.00'	06:10:05.4	3.3	Vernon, Ontario, Canada
05/25/88	44°49.23'	74°58.16'	09:30:32.38	1.9	Raymondville, N.Y.
06/24/88	44°31.61'	74°37.76'	04:44:32.86	1.7	East of South Colton, N.Y.
08/09/88	45°23.05'	75°55.00'	16:5:950.59		Ottawa, Contario, Canada
08/09/88	45° 0.35'	74°59.00'	13:57:28.19	3.0	Ingleside, Ontario
08/09/88	44°59.77'	75° 0.11'	14:52:35:14	1.0	Ingleside, Ontario
08/10/88	44°32.00'	75° 3.26'	14:32:41.16		Near Potsdam, N.Y.
08/11/88	44°59.68'	74°59.78'	05:05:39.50	2.1	Ingleside, Ontario
08/28/88	43°57.17'	74°35.21'	09:10:19.63	1.7	Forked Lake, N.Y.
10/01/88	44°34.23'	75°23.71'	07:08:29.32	1.8	Rensselaer Falls, N.Y.
11/01/88	44°49.22'	74°46.21'	20:56:49.07	1.9	North Lawrence, N.Y.
11/08/88	44°52.16'	73°29.16'	07:23:9.58	1.5	Chazy, N.Y.
11/24/88	43°47.25'	74° 9.88'	16:11:28.84	2.3	East of Indian Lake, N.Y.
11/25/88	48°7.2'	71°10.8'	23:46:06	6.0	Saguenay Earthquake, Quebec, Canada
12/24/88	44°42.95'	73°26.07'	11:04:19.32	2.0	Plattsburgh, N.Y.
01/15/89	44°50.25'	73°39.97'	05:05:24.78	1.7	West Chazy, N.Y.
02/01/89	44°37.79'	74°59.82'	12:44:9.25	1.7	Hannawa Falls, N.Y.
02/03/89	44°45.48'	74°39.45'	04:21:3.67	1.9	West Bangor, N.Y.
02/08/89	44°31.63'	74°53.14'	21:05:7.43	1.7	South Colton, N.Y.
03/16/89	60°3.72'	70°3.36'	04:17:29.40	5.2	Baffin Island, Quebec, Canada
05/29/89	43°52.04'	76°10.42'	19:18:52.25	2.5	SW of Watertown, N.Y.
05/30/89	44°12.77'	75°34.87'	03:58:47.99	1.6	Antwerp, N.Y.
05/31/89	44°44.12'	74°18.68'	07:31:20.36	1.0	Mountain View, N.Y.
06/29/89	44°52.82'	74°21.42'	10:41:33.86	1.8	Malone, N.Y.
07/08/89	44°42.56'	74°31.09'	06:50:5.36	1.9	Dickenson Center, N.Y.
07/08/89	44°42.80'	74°29.72'	06:50:5.01	2.3	SW of Malone, N.Y.
11/16/89	47°12.60'	76°27.06'	09:24:06.86	3.7	Maniwaki, Quebec, Ontario
01/13/90	43°49.68'	75°30.84'	02:20	2.2	
03/03/90	45°7.68'	75°15.6'			
10/07/90	46°18.8'	75°11.7'	08:47:30.8	3.9	South of Mont-Laurier, Quebec
10/19/90	46°28.4'	75°35.5'	07:01:57.4	5.0	Mont-Laurier Region Quebec
11/15/90	47°7.68'	76°13.14'			Maniwake, Quebec, Canada
02/16/91	44°22.0'	75°1.8'	00:03:48.2	2.4	Adirondack Region
02/16/91	44°44.4'	74°8.7'	21:46:42.2	2.9	Chasm Falls, N.Y.
04/06/91	46°12'	76°48'	05:26	3.9	133 km NW Ottawa
05/16/91	44°20.64	75°05.7'	00:03:47.8	2.4	Russell, N.Y.
05/17/91	45°30'	74°24'	18:08	3.2	64 km West of Montreal, Quebec
06/16/91	47°0'	76°42'	16:46	4.2	198 km NW of Ottawa, Ontario
07/04/91	45°12'	73°48'	01:47	3.8	37 km SW of Montreal, Quebec
07/05/91	45°6'	73°58'	01:47:38.90	3.6	Ormstown, Quebec, Canada
07/08/91	45°33.48'	77°8.64'	10:08:14.9	3.4	Eganville, Ontario, Canada
08/22/91	45°52.8	75°7.2'	08:22:55.3	3.1	Cheneville, Quebec, Canada
01/05/92	44°49.86'	74°45.48'	22:36:12	3.1	Brasher Falls, N.Y.
03/31/92	45°5'	75°38'	10:45:21.01	3.0	Near North Gower, Ontario
04/17/92	44°53'	74°45'	01:38:31.59	2.9	4 km south of Helena, N.Y.
04/17/92	44°56'	74°44'	02:11:07.56	2.9	1 km north of Helena, N.Y.
04/19/92	44°54'	74°33'	09:39:30.25	3.0	Easter Sunday, Bombay, N.Y.

Table 1 (continued)

Date	Latitude North	Longitude West	Origin Time UTC	Magnitude	Comments
05/19/92	46°24'	74°54'	05:59:00	3.7	127 km northeast of Ottawa, Ontario
05/30/92	46°27'	74°57'	08:03:4.41	2.9	7 km NW of L'Annonciation, Quebec
06/01/92	45°17'	74°4.0'	12:01:4.80	2.4	2 km southwest of Les Cedres, Quebec
06/03/92	46°12'	75°0'	04:40:00	3.3	106 km northeast of Ottawa, Ontario
06/20/92	44°34'	75°2.6'	04:59:1.80	2.5	3 km southwest of Pierrepoint, N.Y. (near Carthage-Colton Mylonite Zone)
07/01/92	43°57'	74°14.2'	04:18:8.76	2.8	Epicenter located at Newcomb, N.Y. in (epicentral region of Goodnow Quake 10/7/83)
10/05/92	44°50'	74°40'	22:36:01.4	3.2	Brasher Falls, N.Y.
11/11/92	46°55.2'	75°138'	09:02:42.00	3.3	Mt. Laurier, Quebec, Canada
11/17/92	46°42'	74°54'	03:58:4.65	4.4	32 km NW of Hawkesbury, Ontario
12/16/92	44°46'	74°37'	23:24:59.38	2.6	Moira, N.Y.

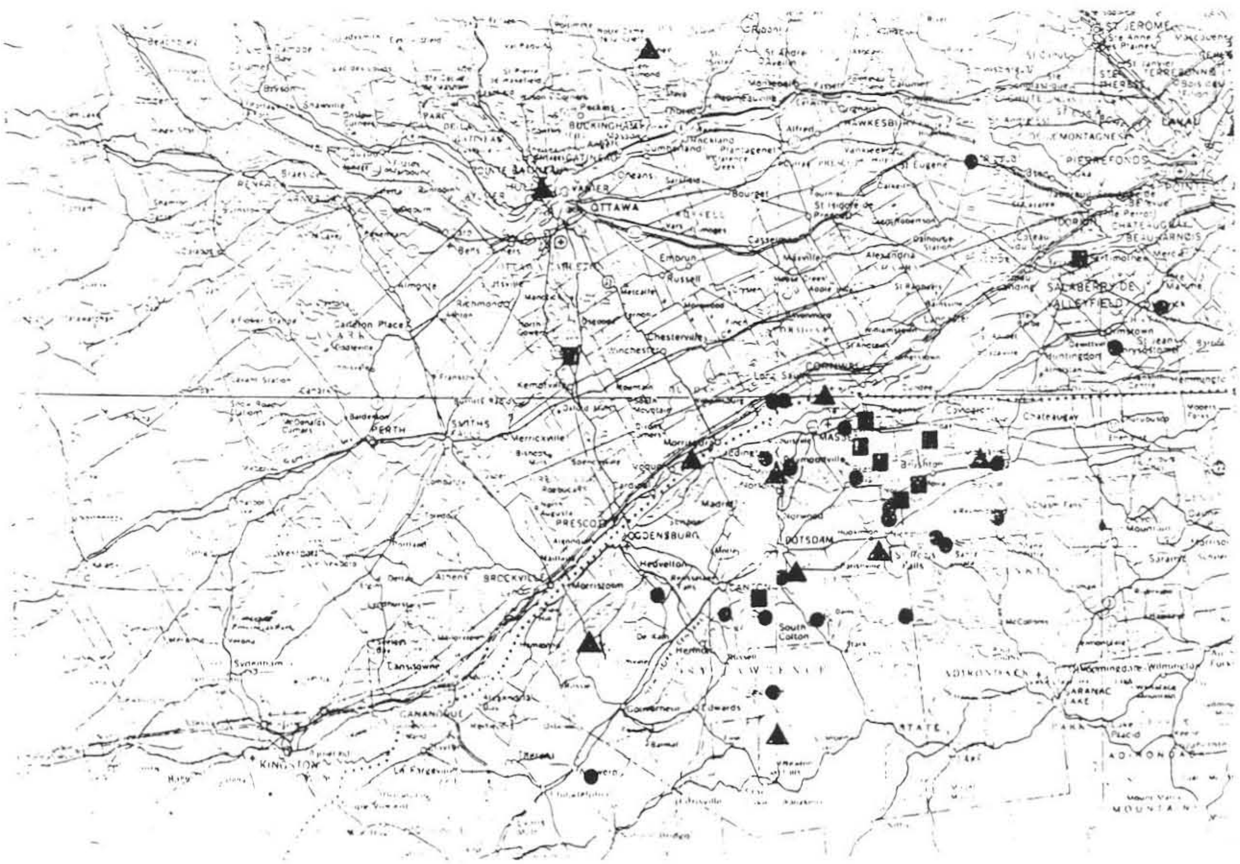


Figure 8: Earthquake epicenters recorded by Potsdam College Seismic Network (1988 - 1992)
Seismic Field Stations
Earthquake Epicenters 1992
Epicenters

the St. Lawrence Trough and connect with the north-northwest trending zone of mylonites north of the Ottawa River (Figure 9). These mylonite zones are believed to be contacts or fault zones between contrasting terrains (Buddington and Leonard 1962). If this zone is continuous the connection between them would pass directly beneath the Cornall-Massena area. Several earthquakes are aligned along the possible continuation of this zone.

SUMMARY AND CONCLUSIONS

The Potsdam College Seismic Network consist of 7 short period vertical seismographs located in the St. Lawrence Valley in northwestern New York. The network provides an important educational tool at the college and significant public service to the area. The network also provides information on the location of local earthquakes and their first motions for research in seismology. Most of the earthquakes recorded by the network occur in the northern New York-Western Quebec seismic zone. Forty eight local earthquakes have been recorded by the network during the past five years. Eighteen of these earthquakes had epicenters located in a northwest-southeast trending belt in the Massena-Cornwall area. These epicenters lie in the vicinity of two faults of the Ottawa-Bonnechere graben that are inferred to extend into the United States in the Massena, N.Y. area. A second belt of epicenters trends northeastward along the northern edge of the Frontenac axis and Adirondack Dome. Seven of the epicenters lie along the Carthage-Colton Mylonite Zone and a possible extension of it beneath the lower Paleozoic rocks in the Massena area.

WORKSHOP SCHEDULE

The workshop on the Potsdam Seismic Network will convene in the Geology Department at Potsdam College at 3:00 p.m., Friday, September 24, 1993. Participants should report to Room 120 in Timerman Hall to hear a slide talk and view a videotape about the seismic network. After the slide talk we will visit the Potsdam Seismic recording stations in the hallway of Timerman Hall to discuss the recording of the earthquakes. A field trip will be taken (transportation will be provided) to visit a seismic field station (PTN) to demonstrate how stations are installed, and operated. Finally we will return to Timerman Hall to locate a local earthquake by computer techniques. The workshop should be completed between 5:00 and 6:00 p.m. Several handouts will be given to participants including a booklet about the seismic network, a seismic report of earthquakes recorded during the past five years, a local earthquake exercise and a packet about earthquakes in New York State.

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WORKSHOP 2

CATHODOLUMINESCENCE IN SEDIMENTARY PETROLOGY

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Canton, NY 13617

INTRODUCTION

Cathodoluminescence (CL) is widely used in sedimentary petrology, especially in the study of carbonate rocks. The factors governing CL intensity and spectral emission are now becoming known and CL is gaining acceptance as an important qualitative analytical tool.

This workshop will examine the application of CL to sedimentary petrography. A lecture session will discuss and evaluate CL in carbonates and in siliciclastics. Issues such as instrumentation, cement stratigraphy, discrimination of altered vs. unaltered skeletal material, sandstone framework mineralogy, porosity occlusion and pressure solution, and diagenesis will be discussed. Projection facilities will permit real-time examination of samples during discussion. A laboratory session will permit participants to gain experience with CL equipment and effects of operating conditions.

OUTLINE

Nature of CL in crystalline materials

Variability in CL
- Intensity
- Wavelength

Quenchers

Controls on CL

Chemical
- Major elements
- Trace elements

Physical (strain)

CL in rock-forming minerals

Silicates
- Quartz
- Feldspars
- Fe-Mg minerals
- Accessory minerals

Carbonates
- Calcite

Dolomite
Aragonite
Phosphates and others
Apatite

Cements

Diagenesis

Albitization
Carbonates
Pressure solution in sandstones

Instrumentation

Luminescences chambers
SEM-type sources
Microscopes
Spectrometers

Laboratory

Practical experience in operating a Nucleide ELB-2B w/spectrophotometer
Examination of participants samples

WORKSHOP 3

QUATERNARY PALEOECOLOGY USING FOSSIL ORIBATID MITES INTRODUCTION TO ORIBATID MITES AND THEIR APPLICATIONS TO QUATERNARY PALEOCLIMATIC INTERPRETATION

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MARK ERICKSON
St. Lawrence University

ROY NORTON
SUNY College of Environmental Science and Forestry
Syracuse

ABSTRACT

This short course is designed to introduce to anyone interested in one of the newest paleoecological tools, the study of fossil soil mites. These ubiquitous organisms are numerous as fossils in bog and lake sediments. Although they have been occasionally identified from postglacial deposits, there has been little effortmade to determine paleoenvironmental conditions based on oribatid occurrences. That situation is now changing.

The course will be led by three active workers in this field; two are internationally acknowledged experts in oribatid biology, and the third, a paleobiologist, has developed techniques for study of fossil mites - particularly as paleoenvironmental indices. It will provide a rare opportunity for students interested in challenging, cutting-edge, research ideas leading to paleoclimatology and Quaternary paleoecology projects to learn what the field is about.

Paleoacarology is still in its infancy, but the subject it embraces should be of wide application to paleoclimatic investigations, particularly those in which sample sizes are limited.

TOPICS

Introduction

Oribatid biology and examples of ecological specificity

Fossil record of the Oribatida

Oribatid mites - classification within the Arthropoda and Arachnida.

Gross morphology of macropyline and brachypiline oribatids

Morphologic elements and superfamilial separation among Brachypilina

Fossil preservation and utility for paleoclimatic interpretation

Collection techniques for living and fossil material

Separation and storage techniques for fossil material

A WORKSHOP IN EXPLORATION GEOPHYSICS OF THE SHALLOW SUBSURFACE

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State University of New York
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SEISMIC REFRACTION METHOD

Applications

The seismic refraction method has many applications in shallow subsurface geologic investigations. The classic application is the determination of depth to bedrock. The method also plays an important role in groundwater investigations since it is possible to determine depth to water table. It is also an excellent tool in engineering and environmental studies since it makes possible the evaluation of dam sites, highways, bridges and landfill sites. Finally it is an excellent method of teaching refraction seismology principles that are used to investigate the crustal structure of the earth. Our seismic refraction survey will be used to determine the depth to bedrock and water table on the Potsdam College campus.

Equipment

The ES-1225 exploration seismograph will be used to conduct the seismic refraction survey. The instrument is a multichannel CRT-display, printing, signal enhancement shallow exploration seismograph. It is microprocessor-based battery operated and has 12 channels. The instrument is a light and portable field unit. A portable laptop computer with an RS-232 interface will be used in the field to store the data, print seismograms of the seimic traces and analyze the time-distance graph.

The assembly of the ES-1225 system is shown in Figure 1. The geophone cable is laid out and geophones implanted firmly in the earth and connected to the cable. The battery is connected to 12 volts D.C. outlet and the sledgehamnmer with switch is connected to start on the seismograph. A steel plate is placed firmly in the ground at an offset of 10 feet from the nearest geophone. Impacting the plate with sledgehammer when seismograph reads acquisition of data will produce a seismic record with 12 traces (Figure 2). The seismic traces may be seen on a screen and a record of them may be obtained using the print option. The data may also be entered into a portable computer so traces may be seen on a computer monitor.

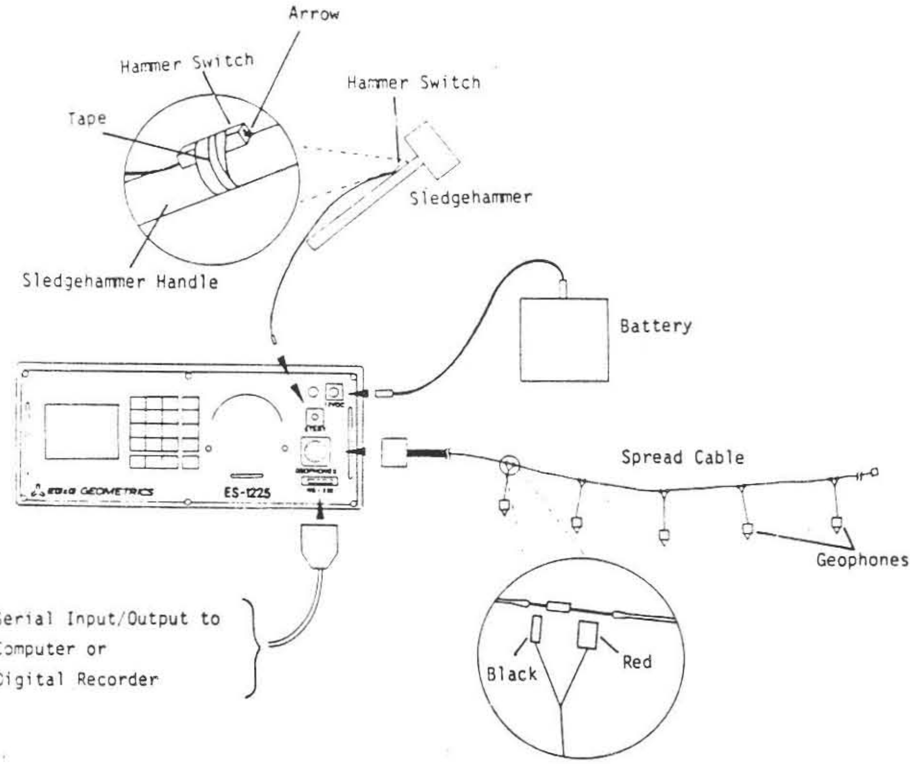

Figure 1: Assembly of the ES-1225 exploration seismograph

Seismic Refraction Method

Seismic refraction requires the generation of seismic waves into the subsurface and an instrument to detect and record the returning refracted waves. We will use an eight pound sledgehammer to generate the seismic waves and a Model ES-1225 Exploration Seismograph to record the waves. The seismograph enables us to accurately measure the travel times of the seismic waves to 12 geophones. The shotpoint and geophones are located along a line so one may plot a travel-time curve of distance versus time. The curve is used to determine velocities and calculate the depths to various layers of rock in the subsurface.

The first arrival times are measured during a refraction survey. These times represent the minimum travel time paths of the seismic waves. These minimum travel time paths are the direct waves arriving at the nearby geophones and the critically refracted waves arriving at the more distant geophones (Figure 3). The refraction method relies on the velocities of the rocks increasing with depth otherwise a critically refracted wave will not occur. Also the length of the geophone spread must be several times the depth of the layers being investigated.

Reading the Seismograms

A typical seismic record of a seismic refraction survey is shown in Figure 2. The vertical lines are time lines with each line representing 2 milliseconds. The horizontal lives are the

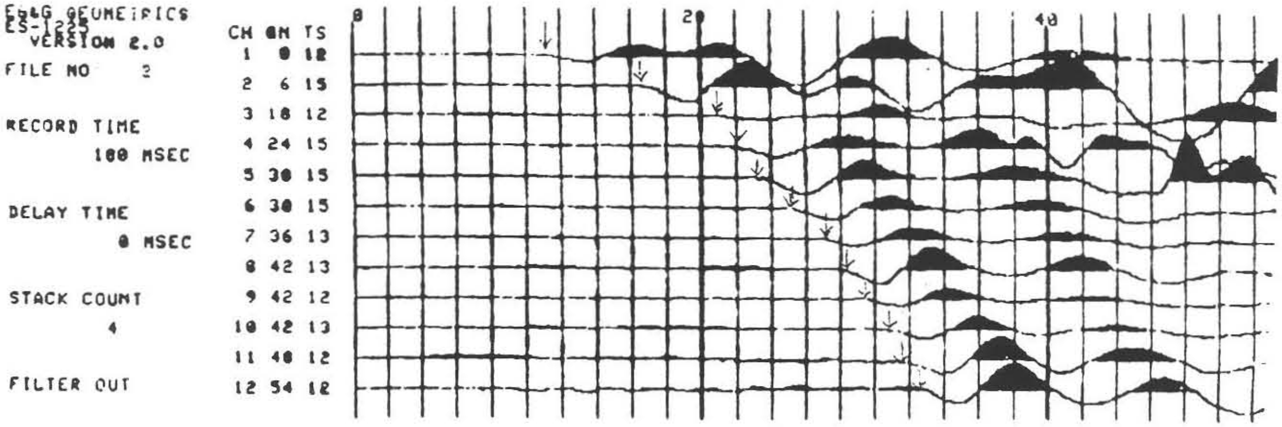

Figure 2: Seismic record showing 12 traces in variable area (VA) mode

traces made from the output of 12 geophones with the top trace representing the nearest geophone at 10 feet offset and the bottom trace representing the output of the furthest geophone at 120 feet distance. The numbers on the left are channel numbers, gain and trace size set for each channel. This is a variable area trace (VA) however wiggle-traces (WT) are also available. First arrival times are determined by picking the first break in the trace. The first break picks are indicated by the arrows in Figure 2. For example trace 1 has a travel time of about ll.8 msecs while trace 12 shows a travel time of about 33 msecs. Trace 1 is the time for the seismic wave to travel 10 feet since geophone 1 is 10 feet from the shot point while trace 12 is first arrival time at a distance of 120 feet.

Seismic records may also be drawn from the data transfered to the laptop portable computer. The advantages of this method is the traces may be enhanced to help pick the first arrivals and the time scale may be increased to make more accurate time measurements. Also many seismic surveys can be conducted in a day with all the data stored in the laptop. It is also possible to do the time-distance graph with the computer using the SEISVIEW program.

Figure 4 shows two seismic records of traces produced by the laptop computer data. Figure 4a shows the traces 3 and 4 have poor first breaks. Enhancement of these traces with the laptop computer is shown in Figure 4b where the first breaks are seen more clearly. In Figure 5 the time scale was increased so 1 mm equals 0.2 msec.

Analysis of seismic data

The travel times of the first arrivals may be determined from the seismic record. The first arrivals are plotted to construct a time-distance curve with the distance plotted as the X horizontal axis and the time in milliseconds plotted is the Y or vertical axis. Figure 6 shows

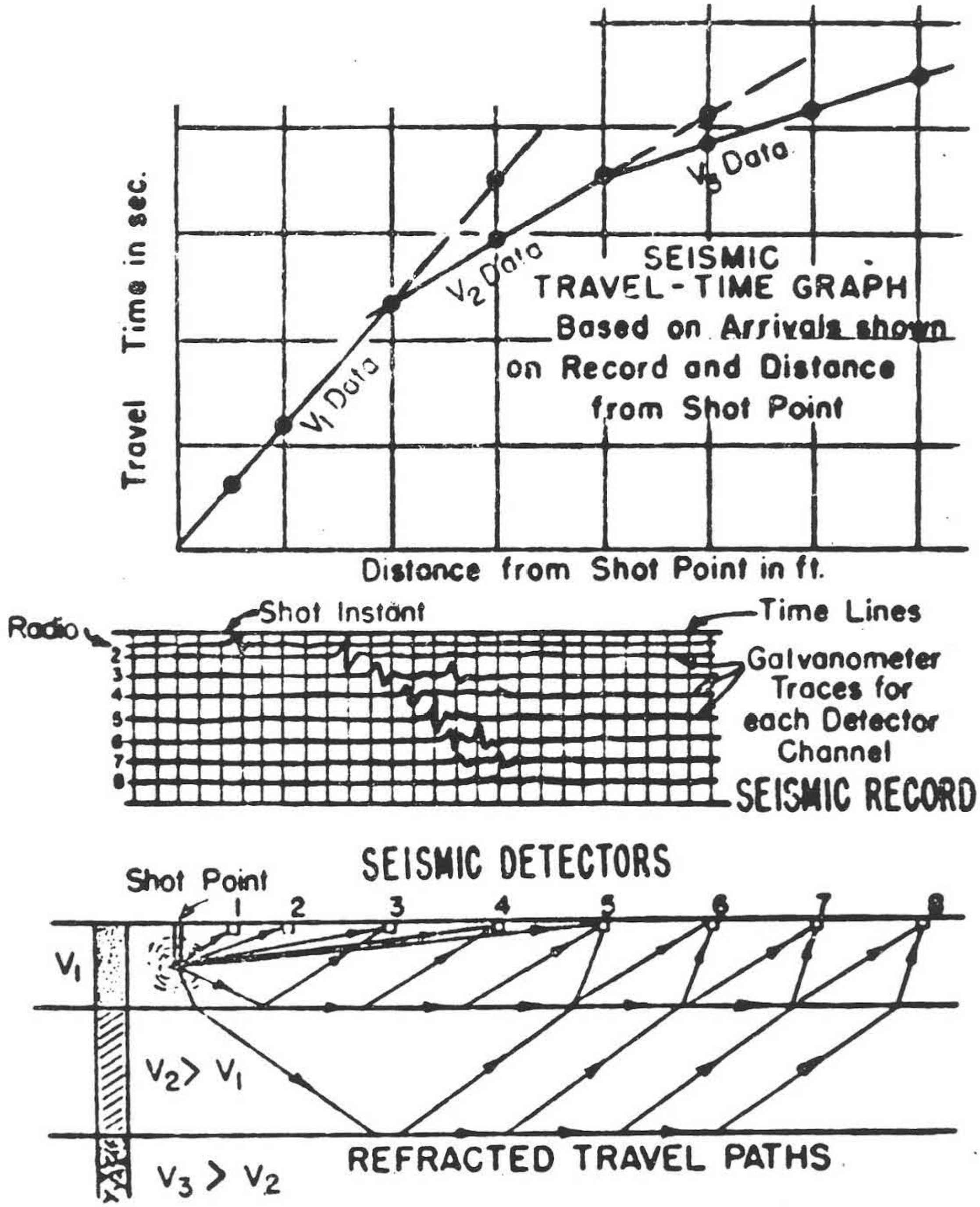

Figure 3: Travel-time graph, seismic record and wave paths of a seismic refraction survey (From Woollard 1954)

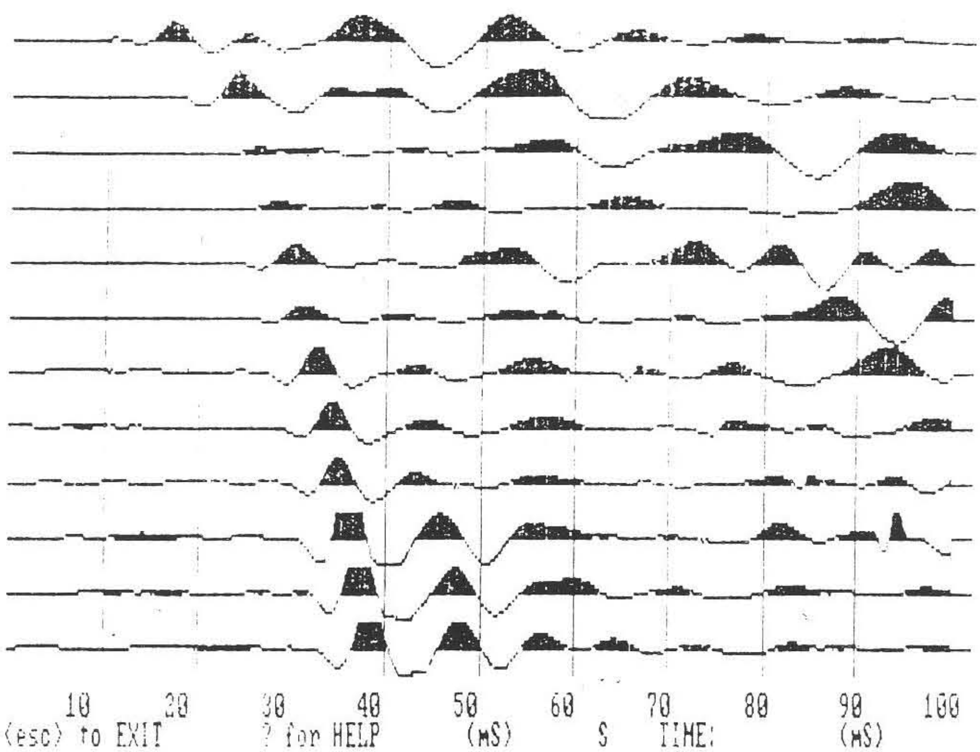

Figure 4a: Seismic traces drawn from output to laptop computer.
Seismic record before enhancement of traces 3 and 4

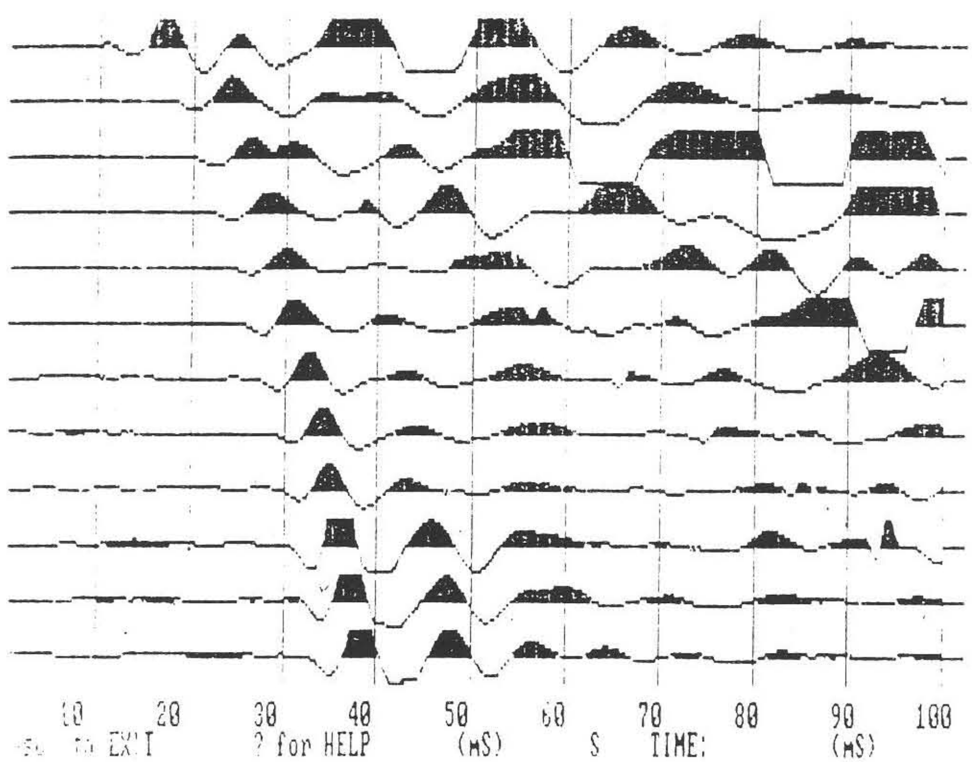

Figure 4b: Seismic traces drawn from output to laptop computer.
Seismic record after enhancement of traces 3 and 4. Note how traces 3 and 4 have more well defined first breaks.

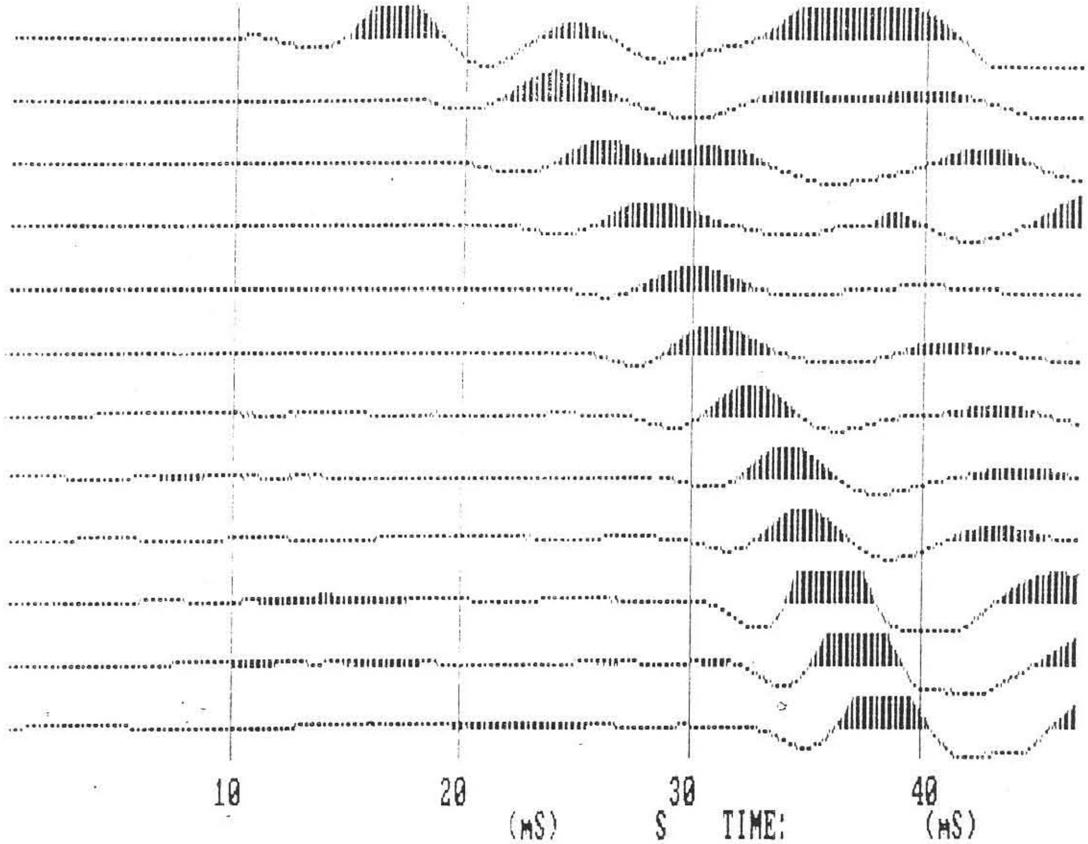

Figure 5: Seismic record with horizontal time scale increased with laptop computer to make more accurate time measurements

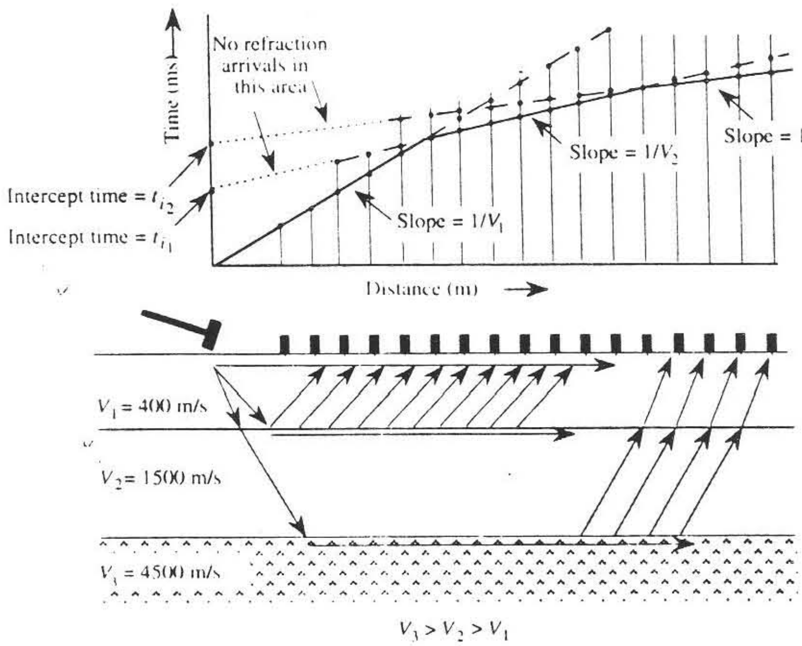

Figure 6: Diagram showing ray paths and time-distance graph for the direct and two critically refracted rays. (From Burger 1992)

an example of a time distance curve for a three layer case. The geophones closest to the shot receive seismic waves traveling through the first layer. These are P waves which travel directly to the geophones along the minimum travel time path. Each segment of the travel-time curve represents a layer of rock. The inverse of the slope of each segment is equal to the velocity of the layer. The extension of the lines to the time axis gives the intercept times and the intersections of the lines give the crossover distances. The crossover distances, intercept times and velocities are used to calculate the depths to the various layers of rock.

How to make velocity and depth determinations

The basic procedure for measuring the velocities and depths are listed below:

1. Draw lines that best fit the points plotted on the time-distance curve.

2. Pick two points on the lines and divide the distance between the points (Ft) by the time interval (msec). The velocities will be in feet/msec. Change the velocities to feet/sec by multiplying by a thousand.

3. Extend the slopes of the lines to the time axis to obtain the intercept times and project the line intersections downward to the distance axis to obtain the crossover distances.

4. Use the formulas below to obtain the thicknesses of the layers. The thicknesses may be obtained by using crossover distances or intercept times. The thickness of the first layer using crossover distance is:

$$Z_1 = \frac{X_c}{2}\sqrt{\frac{V_2 - V_1}{V_2 + V_1}}$$

Where X_c = crossover distance
Z_1 = thickness
V_1, V_2 = Velocities of first and second layers

The thickness of the first and second layers by using intercept times are given by:

$$Z_1 = \frac{T_1}{2}\frac{V_1V_2}{\sqrt{V_2^2 - V_1^2}}$$

$$Z_2 = \left[T_{12} - \frac{2Z_1\sqrt{V_3^2 - V_1^2}}{V_1V_3}\right]\frac{V_2V_3}{2\sqrt{V_3^2 - V_2^2}}$$

Interpretation of data

A good interpretation of the travel-time curve requires some knowledge of the geology of the area and seismic velocities of various rocks. The campus has bedrock overlain by glacial deposits. The bedrock is either Precambrian gneiss or Potsdam sandstone. The velocities of various rock types are given in Table 1. It is customary to draw a model of the subsurface indicating the number of layers, their velocities, lithologies, and thicknesses.

EARTH RESISTIVITY SURVEY

Applications

The electrical resistivity method has many applications in shallow subsurface geologic studies. The method may be used to determine depth to bedrock and water table. It can be used to locate sand and gravel deposits, buried stream channels and mineral deposits. It is also an effective tool for mapping salt water-fresh water interface and contaminant areas associated with landfill sites. Some other uses are in geothermal exploration and mapping archaelogical sites.

Earth Resistivity Methods

Electrical resistivity surveying measures the apparent earth resistivity from the surface. Various types of earth materials have resistivities that can be distinguished from one another. The basic types of field procedures used are vertical electrical sounding and resistivity profiling. In vertical electrical sounding (VES) we determine how resitivity varies with depth by increasing electode spacing. In resistivity profiling, a fixed electrode separation is maintained however the location of the spread is changed to determine horizontal variations in resitivity.

Vertical Electrical Sounding (VES):

Figure 7 shows the main elements of electrical resistivity surveying and Figure 8 illustrates the procedure used in vertical electrical sounding. Four electrodes are laid out along a line. The outer electrodes (C1 and C2) are current electrodes and the inner electrodes (P1 and P2) are potential electrodes. A current is supplied through the current electrodes and the voltage drop is measured between the potential electrodes. Measurements of the current flow, potential drop and electrode spacing are used to calculate the apparent resistivity of the material to a depth assumed equal to the electrode spacing. Measurements at greater depth are made by increasing the spacing between electrodes. The method for most vertical electrical sounding surveys is the Wenner configuration where the spacing between electrodes is kept equal. When the Wenner method is used the apparent resistivity (p) is computed by the formula:

$$P = 2\pi A \frac{V}{I}$$

Where	P	= apparent resistivity
	A	= electrode spacing
	V	= voltage drop
	I	= current flow

The term V/I is resistance with units of ohms. The electrode spacing will be measured in meters so our resistivity values will have units of ohm-meters. As a rough guide materials with resistivities less than 100 ohm-meters are considered low and materials greater than 1000 ohm meters are considered high. Resistivities of various earth materials are listed in Table 2.

Equipment:

Our resistivity survey will be conducted with a Keck Earth Resistivity Instrument designed for making earth resistivity measurements. Four 45 volt batteries furnish the power for the instrument. Extra batteries may be added in cases where dry earth makes electrical resistance high. When current is introduced into the earth a meter needle is reflected from the zero or null position. Rotating a dial brings the needle back to the zero position. The dial gives an ohmmeter reading in ohms which is equal to V/I in the formula for calculating resistivity. The ohmeter reading is recorded then used to calculate the resitivity. Elctrodes can now be moved to greater distances to calculate resistivities to greater depths.

Field Procedure:

Figure 9 is a data sheet used to record the resitivity values for vertical electrical sounding by the Wenner method. Readings are made at intervals shown in the left column. Note these spacings begin with 1 meter then proceed through values equally spaced on a logarithmic scale. This is done because a larger electrode spacing yields information over a much larger volume of earth, thus a given volume becomes proportionately less important. A second reason is that normally data is plotted on log-log graph paper. We record resistance values for current flow from C1 to C2 and C2 to C1 under R1, and R2 on the data sheet. The average is determined in column Rav and the calculated resistivity written in the last column under (Pa). The resistivity is calculated by using:

$$P = 2 \; p \quad AR$$

Where R is resistance (ohms)
A is electrode spacing (meters)

Table 1: Velocities of compressional waves (P) for various rocks found in the Earth's crust

A. *Classification According to Material*

Material	Velocity*	
	Ft./Sec.	M./Sec.
Weathered surface material	1,000— 2,000	305— 610
Gravel, rubble, or sand (dry)	1,500— 3,000	468— 915
Sand (wet)	2,000— 6,000	610— 1,830
Clay	3,000— 9,000	915— 2,750
Water (depending on temperature and salt content)	4,700— 5,500	1,430— 1,680
Sea water	4,800— 5,000	1,460— 1,530
Sandstone	6,000—13,000	1,830— 3,970
Shale	9,000—14,000	2,750— 4,270
Chalk	6,000—13,000	1,830— 3,970
Limestone	7,000—20,000	2,140— 6,100
Salt	14,000—17,000	4,270— 5,190
Granite	15,000—19,000	4,580— 5,800
Metamorphic rocks	10,000—23,000	3,050— 7,020
Ice	12,050	

B. *Classification According to Geologic Age*

Age	Type of Rock	Velocity	
		Ft./Sec.	M./Sec.
Quaternary	Sediments (various degrees of consolidation)	1,000— 7,500	305— 2,290
Tertiary	Consolidated Sediments	5,000—14,000	1,530— 4,270
Mesozoic	Consolidated Sediments	6,000—19,500	1,830— 5,950
Paleozoic	Consolidated Sediments	6,500—19,500	1,980— 5,950
Archeozoic	Various	12,500—23,000	3,810— 7,020

C. *Classification According to Depth* †

	0—2000 ft. (0—600 M.) Ft./Sec.	2000—3000 ft. (600—900 M.) Ft./Sec.	3000—4000 ft (900—1200 M.) Ft./Sec.
Devonian	13,300	13,400	13,500
Pennsylvanian	9,500	11,200	11,700
Permian	8,500	10,000	.....
Cretaceous	7,400	9,300	10,700
Eocene	7,100	9,000	10,100
Pleistocene-to-Oligocene	6,500	7,200	8,100

* The higher values in a given range are usually obtained at depth.

† Data from B. B. Weatherby and L. Y. Faust, *Bull. Amer. Assoc. Petrol. Geologists*, 19 (1935) 1.

[a]Reprinted from pg. 660 of Jakosky[2].

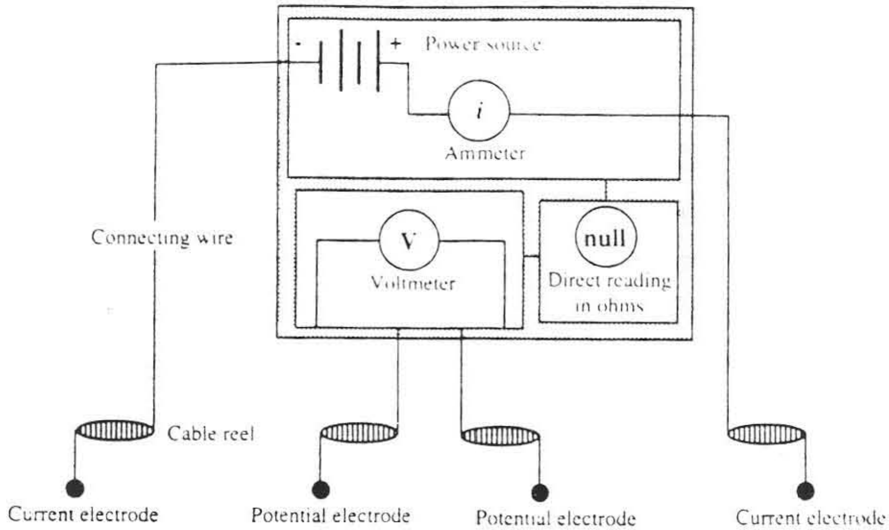

Figure 7: Main elements of electrical-resistivity surveying including electrodes, power source, ammeter, and voltmeter. (From Burger 1992)

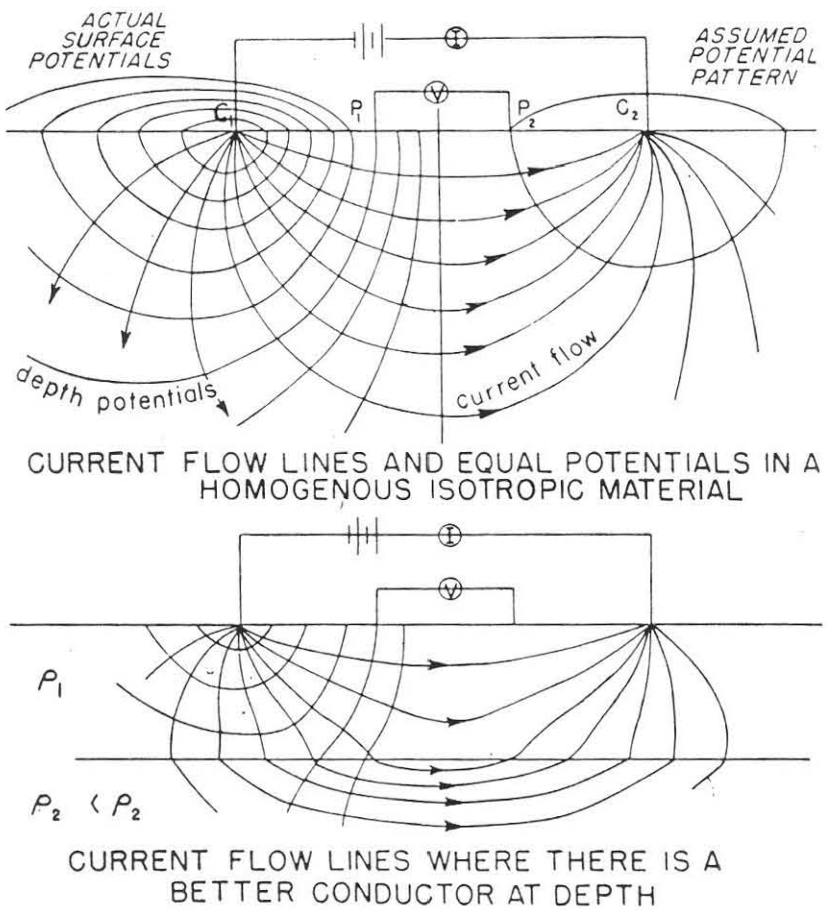

Figure 8: Diagram illustrating current flow lines and equal potentials in a resistivity survey. (From Woollard 1954)

WENNER SURVEY

Sounding Location/Number ____________________

Operator/Date ____________________

Equipment ____________________

____

Computation Formula: $\rho_a = K\ (V/I)$. $K = 2\pi a$

a(m)	TAPE P	TAPE C	K	R_1	R_2	R (AV)		ρ_a
0.47			2.95					
0.68			4.27					
1.00	0.50	1.50	6.28					
1.47	0.75	2.25	9.24					
2.15	1.07	3.22	13.5					
3.16	1.60	4.74	19.9					
4.64	2.32	6.96	29.2					
6.81	3.40	10.22	42.8					
10.0	5.00	15.00	62.8					
14.7	7.35	22.00	92.4					
21.5	10.75	32.25	135					
31.6	15.80	47.40	199					
46.4	23.20	69.60	292					
68.1	34.05	102.15	428					
100	50.0	150.0	628					
147	73.5	220.50	924					
215	107.5	322.5	1351					
316	158.0	474	1985					
464	232	696	2915					
681	340	1021	4279					
1000	500	1500	6283					

Figure 9: Data sheet for recording field measurements using the Wenner configuration

The value of 2 p A is a constant for each value of A and is known as a geometrical factor (K). The value of K is in column 4 so the resistivity is calculated simply by multiplying K times Rav.

Interpretation of Data:

While the acquisition of resistivity data is relatively simple the results are difficult to interpret. A procedure normally followed for the interpretation of the resistivity data is as follows. First the resistivity data is plotted on log-log paper with one axis being electrode spacing and the other being apparent resistivity. The preferred method of plotting on log-log paper makes the shape and size of the curve independent of units and electrode separation used. The standard procedure used to interpret resistivity sounding data consist of the following steps.

1. Assume a resistivity model based on the resistivity profile and any other information you may have such as well logs and seismic surveys. A model consist of the number of layers, resistivity of each layer and thickness of each layer.

2. Compute the apparent resistivities expected from your assumed model. This is usually done with a computer program.

3. Compare the observed resistivity field curve with the computed values based on your assumed model.

4. Modify the model until a best possible agreement is obtained between computed and field values. Keep in mind that a good fit means only the fit is good and that the model isn't necessarily the correct one. It is always possible that many different models may produce equally good fits. Additional information such as well logs are always needed to choose the most likely correct model. Also computer software is available (Burger 1992) that will modify your model until an excellent fit occurs between the field curve and computed resistivity values. Some typical resistivity curves are shown in Figure 10 with their interpretation or model shown below the curve.

It is difficult to correlate resistivities with specific rock types without geologic information because of the great range of resistivity values of rocks. No other physical property of naturally occurring rocks or soils displays such a wide range of values. Bedrock has higher resistivities than saturated sediments. Unsaturated sediments above water table have higher resistivities than saturated sediments. Table 2 from Burger (1992) show various materials and their resistivities in ohm-meters.

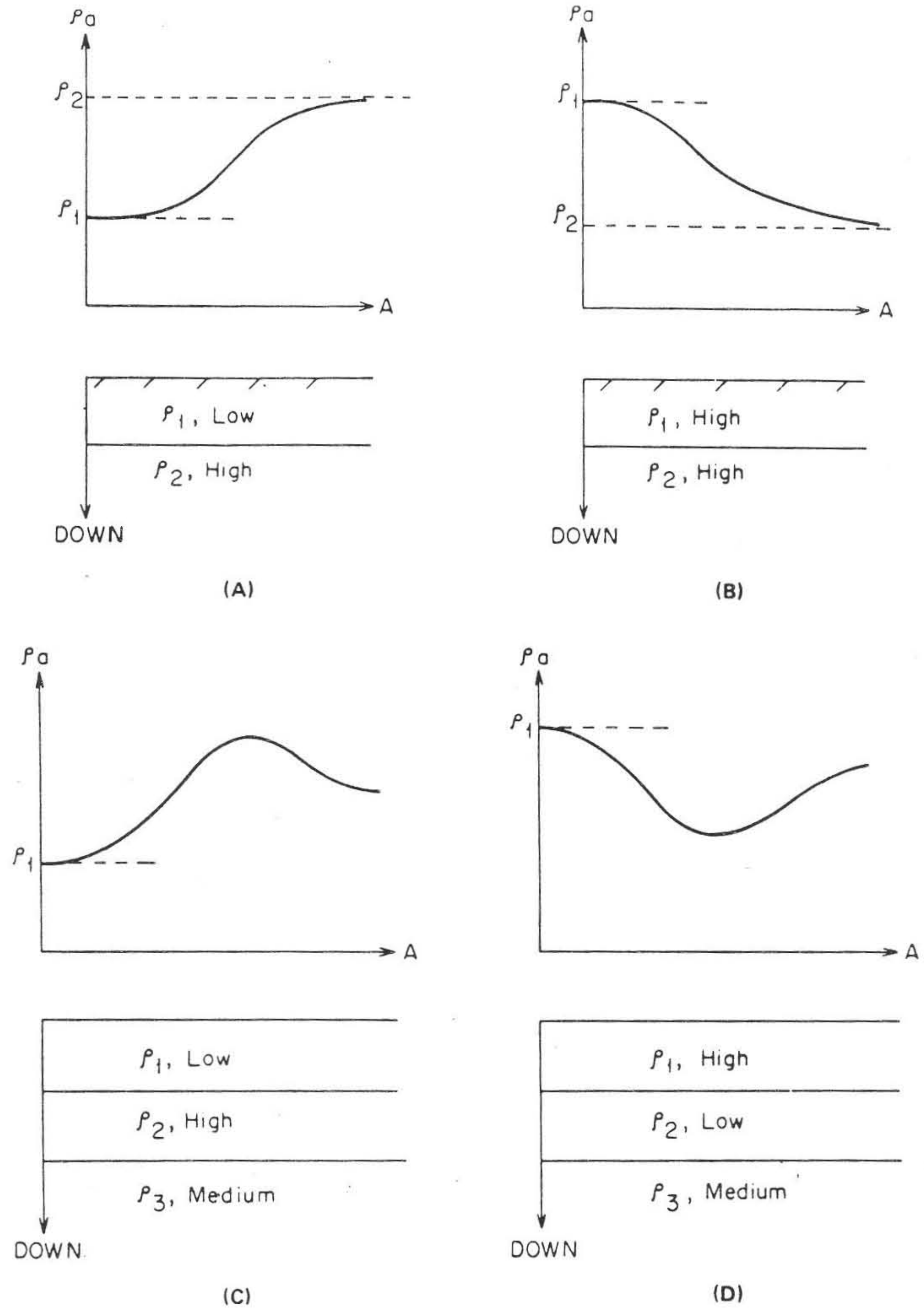

Figure 10: Resistivity sounding curves over two and three layer models.
(From Mooney 1980)

Table 2: List of resistivities of various materials (From Burger 1992)

Material	Resistivity (Ω-m)
Wet to moist clayey soil and wet clay	1s to 10s
Wet to moist silty soil and silty clay	Low 10s
Wet to moist silty and sandy soils	10s to 100s
Sand and gravel with layers of silt	Low 1000s
Coarse dry sand and gravel deposits	High 1000s
Well-fractured to slightly fractured rock with moist-soil-filled cracks	100s
Slightly fractured rock with dry, soil-filled cracks	Low 1000s
Massively bedded rock	High 1000s

MAGNETIC METHOD

Applications

A magnetometer measures changes in the earth's magnetic field strength. Any magnetic object that alters the earth's magnetic field can potentially be detected by magnetic surveying. Traditionally magnetic surveys have resulted in the construction of magnetic maps that show patterns diagnostic of a particular rock assemblage thus the method was useful in geologic mapping. The method has also been used to estimate depth to Precambrian basement by oil companies. More recent applications are the use of magnetics to detect buried steel tanks and drums containing hazardous waste materials. Archeologists have also found the method useful for locating cultural features with anomalies being due to ferrous metals, hearths, and kilms.

Magnetometers also have the option of measuring the vertical magnetic gradient which has several advantages over the use of total field measurements. Near surface sources of magnetic anomalies are accentuated over deeper regional bodies by the gradient measurements. The magnetic gradient also exhibits superior resolving power. This combined effect is important in locating lithologic contacts and shallow buried steel drums. Magnetic gradient data also aids in the interpretation of the physical characteristics of the source.

Equipment

A portable proton magnetometer with gradiometer option G856 AX will be used to conduct a magnetic survey over a small area or campus. The magnetometer reads and displays total magnetic field strengths (gammas) at the touch of a button. The readings are stored along with time, date and station number. The data is then fed into a computer and a printout may be obtained. A computer program MAGLOC and a contouring program can be used to plot magnetic contour maps based on the data.

The magnetometer can be used to measure magnetic gradient by adding a second sensor. The two vertically separated sensors result in a measurement of the vertical magnetic gradient. Two 55 gallon steel drums are buried on campus for anyone who would like to try locating them by making gradient measurements.

WORKSHOP SCHEDULE
EXPLORATION GEOPHYSICS

The workshop on exploration geophysics will convene in the Geology Department at Potsdam College at 9:00 a.m., Sunday, September 26, 1993. Participants should report to Room 120 in Timerman Hall to have a brief discussion of the geophysical methods included in the workshop. Following the discussion geophysical equipment will be carried to the field behind Timerman Hall where the seismic refraction, electrical resistivity and magnetic surveys will be conducted.

After completion of the field survey, participants will return to Timerman Hall room 120 to analyze and interpret the field data. A demonstration of using the microcomputer to analyze and model the field data will be presented. Finally a discussion of the interpretation of the field data will be conducted. Participants will receive handouts on the geophysical methods of surveying conducted in the workshop. The workshop should be completed at 12:00 noon however participants are free to leave at any time.

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