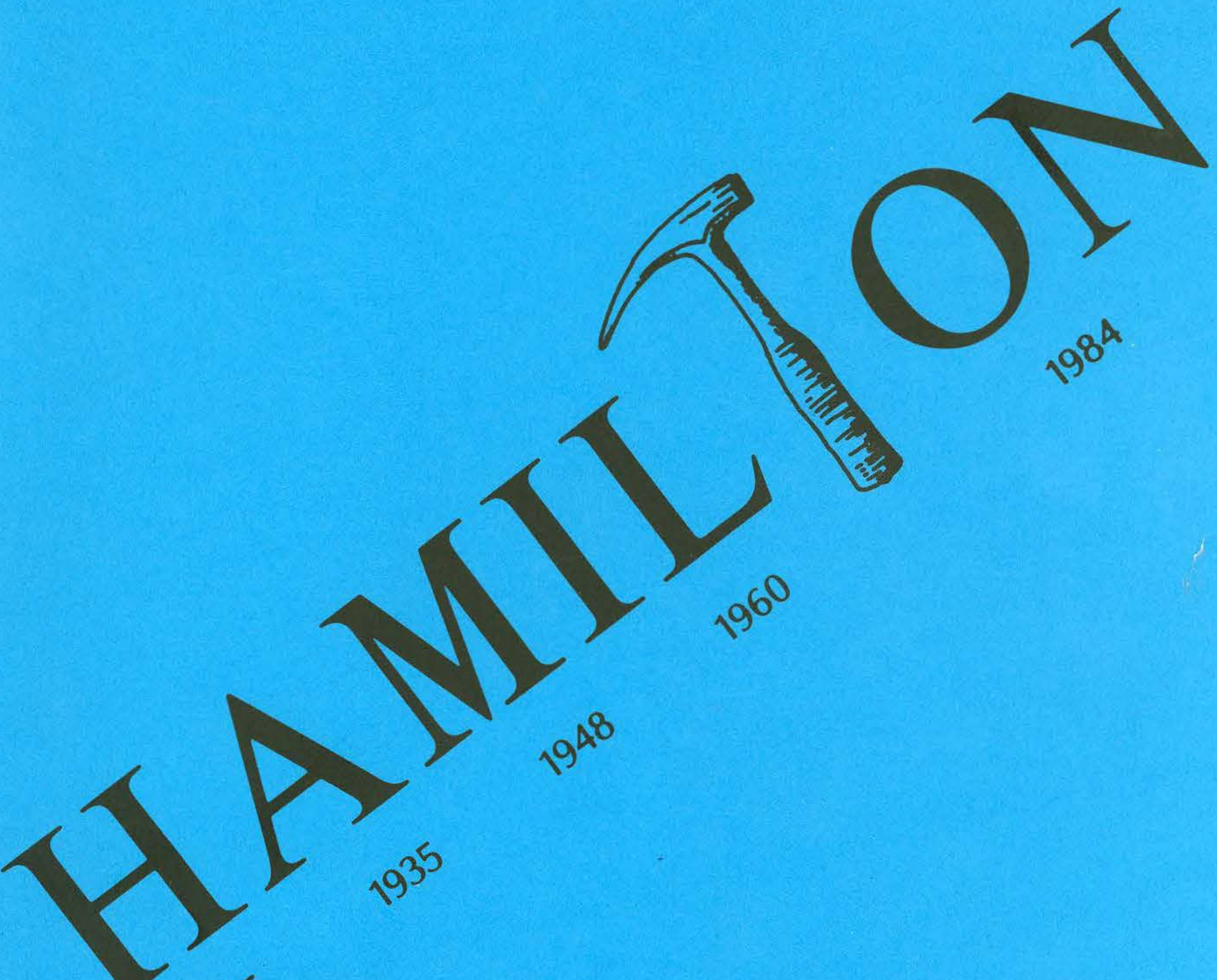
FIELD TRIP GUIDEBOOK

New York State Geological Association



56th Annual Meeting

September 21-23, 1984

Department of Geology - Hamilton College

NEW YORK STATE GEOLOGICAL ASSOCIATION

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FIELD TRIP GUIDEBOOK

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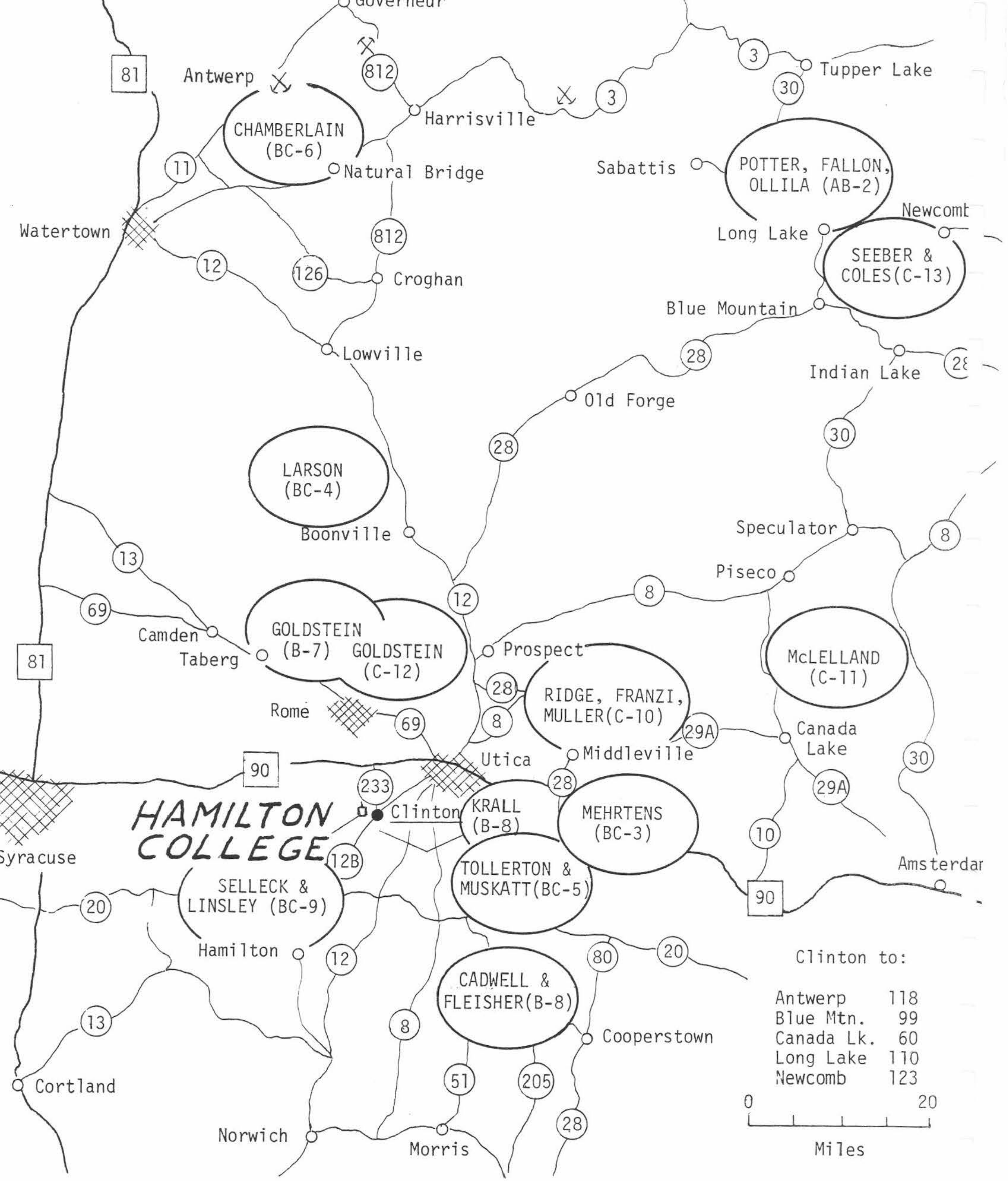

GENERAL LOCATION OF FIELD TRIPS FOR THE 56th ANNUAL MEETING OF THE NYSGA
(AB: Friday and Saturday; B: Saturday only; C: Sunday only; BC: Saturday and Sunday)

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WELCOME FROM THE ORIGINAL HOST

In 1925, on the initiative of Nelson C. Dale, Hamilton College hosted the first meeting of the NYSGA. Fifty-nine years later (with some interim visits in '35, '48, and '60) we enthusiastically welcome you all. Sorry it took so long.

FIELD TRIP LEADERS

We are grateful to our 22 colleagues who have prepared field trip guides and who are leading the 13 field trips. Tewksbury and Allers will examine some key bedrock and surficial exposures in the Black and Mohawk River valleys and reconstruct the geologic history of central New York for the benefit of secondary school earth science teachers. In the Precambrian terrane of the central Adirondacks, Potter and Fallon present the results of recent studies on the stratigraphy, structure, and metamorphism of metasedimentary units, and Ollila examines the nature of the contact zone of the Marcy anorthosite massif. Chamberlain carefully documents the mineral paragenesis of the classical Sterling iron mine at Antwerp. McLelland presents results of rock structure and fabric analysis, part of his well-known long-term study in the southern Adirondacks. Four trips are concerned with Paleozoic geology: Selleck and Linsley examine the sedimentology and fauna of the Hamilton group; Tollerton and Muskatt present new evidence for reconstructing the paleoenvironment of the Bertie Formation; Mehrtens presents new evidence for the foreland basin setting of the Trenton Group; and Larson looks at animal-sediment relations in the Black River Group. Significant new work on deglaciation, ice lobe advances, ice-dammed lakes, and paleomagnetic stratigraphy is being presented on different trips by Cadwell, Fleisher, Krall, Ridge, Franzi, and Muller. Seeber and Coles examine focal mechanisms and surface manifestations of recent seismic activity in the Newcomb region. And, Goldstein brings us face to face with our environmental responsibilities on a trip to a hazardous waste landfill which is contaminating groundwater, and examines a groundwater aquifer in glacial sediments.

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CROSS SECTION OF THE LOON POND SYNCLINE, TUPPER LAKE QUADRANGLE, NEW YORK

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INTRODUCTION

This field trip crosses the Loon Pond Syncline, an area mapped during the summers of 1978, 1979, and 1980 under the direction of Leo Hall at the University of Massachusetts. The syncline, a northeast-trending elongate basin, was named and first mapped by Buddington and Leonard (1962) as the northeast part of the Bog River Synclinorium (Figure 1). The syncline occupies most of the southern half of the fifteen minute Tupper Lake Quadrangle, which lies in the central Adirondacks within the Adirondack Highlands region. The Arab Mt. Anticline, composed of charnockitic gneisses, lies to the north, and the Salmon Lake Anticline lies to the south (Figure 1).

The stratigraphy of the syncline is distinctive. In contrast to most of the Adirondack Highland rocks which are chiefly orthogneisses and anorthosite, a great deal of quartzite, marble, and calc-silicate rock as well as quartzofeldspathic gneisses constitute the rocks here. In this respect the rocks of the Loon Pond area are similar to those of the Northwest Lowlands (Figure 1). Erosion of the multiply deformed rocks has exposed a structural basin which shows some similarity to the Darning Needle Syncline in the Cranberry Lake Quadrangle (Leavell, 1977).

Access: Route 10, linked to Route 30 by Route 10A, extends west across the field area to the old Sabattis Railroad station and provides the primary access to the sparsely populated region (Figure 2). With the exception of a clear area south of Sabattis that burned in 1903, the region is heavily wooded. Topographic relief is about 800 feet.

STRATIGRAPHY

General Statement. The Precambrian rocks of the Loon Pond Syncline have been divided into basement, cover, and intrusive rocks. The basement is formed by the East Charley Pond Gneiss, a hornblende-quartz-perthite rock (Figure 3). The overlying cover rocks are inferred to be of sedimentary and volcanic origin, and their interpreted relative ages, from oldest to youngest, are the Little Charley Pond Formation, the Bear Pond Gneiss, and the Lost Pond Marble.

There are two igneous plutons in the field area: the East Charley Pond Gneiss composes one and the Otter Pond Dioritic Gneiss the other. Their stratiform and generally conformable geometry helps to define the overall structure of the area. Some of the isolated bodies of the Otter Pond Dioritic Gneiss are due to erosion of the multiply deformed rocks. It can be shown at outcrop and map scale that all of the older rocks were intruded by the Otter Pond Diorite prior to the first phase of deformation. Intrusion persisted beyond F, and ceased prior

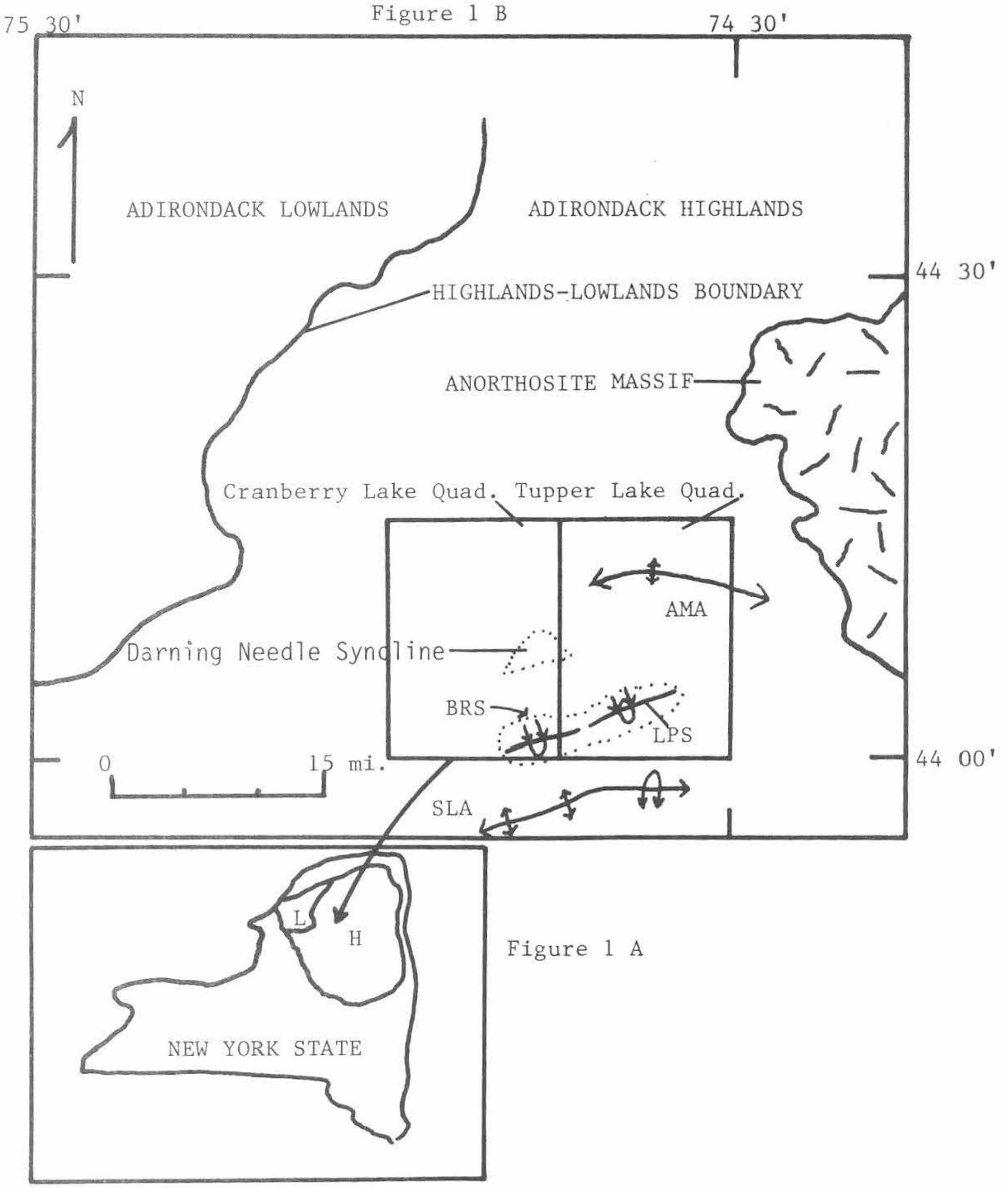


Figure 1 A. Index map of New York State and the Adirondacks. Highland and Lowland divisions of the Adirondacks are noted.

Figure 1 B. The Arab Mountain Anticline lies in the northern half of the Tupper Lake Quadrangle. The Bog River Synclinorium is enclosed by a dotted line (BRS), and the northeastern half of the BRS is the Loon Pond Syncline (LPS). The Salmon Lake Anticline (SLA) lies south of the Tupper Lake Quadrangle.

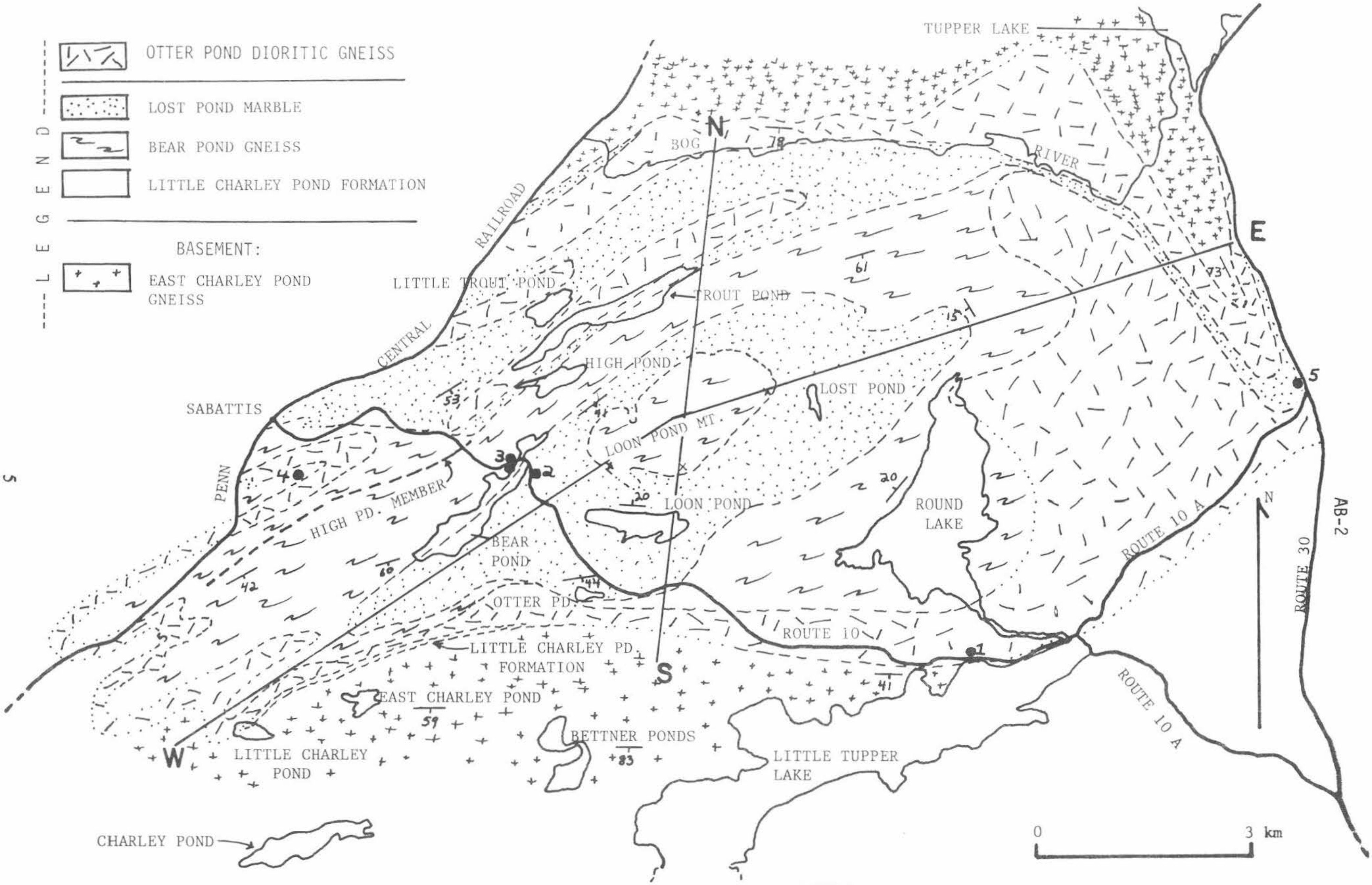
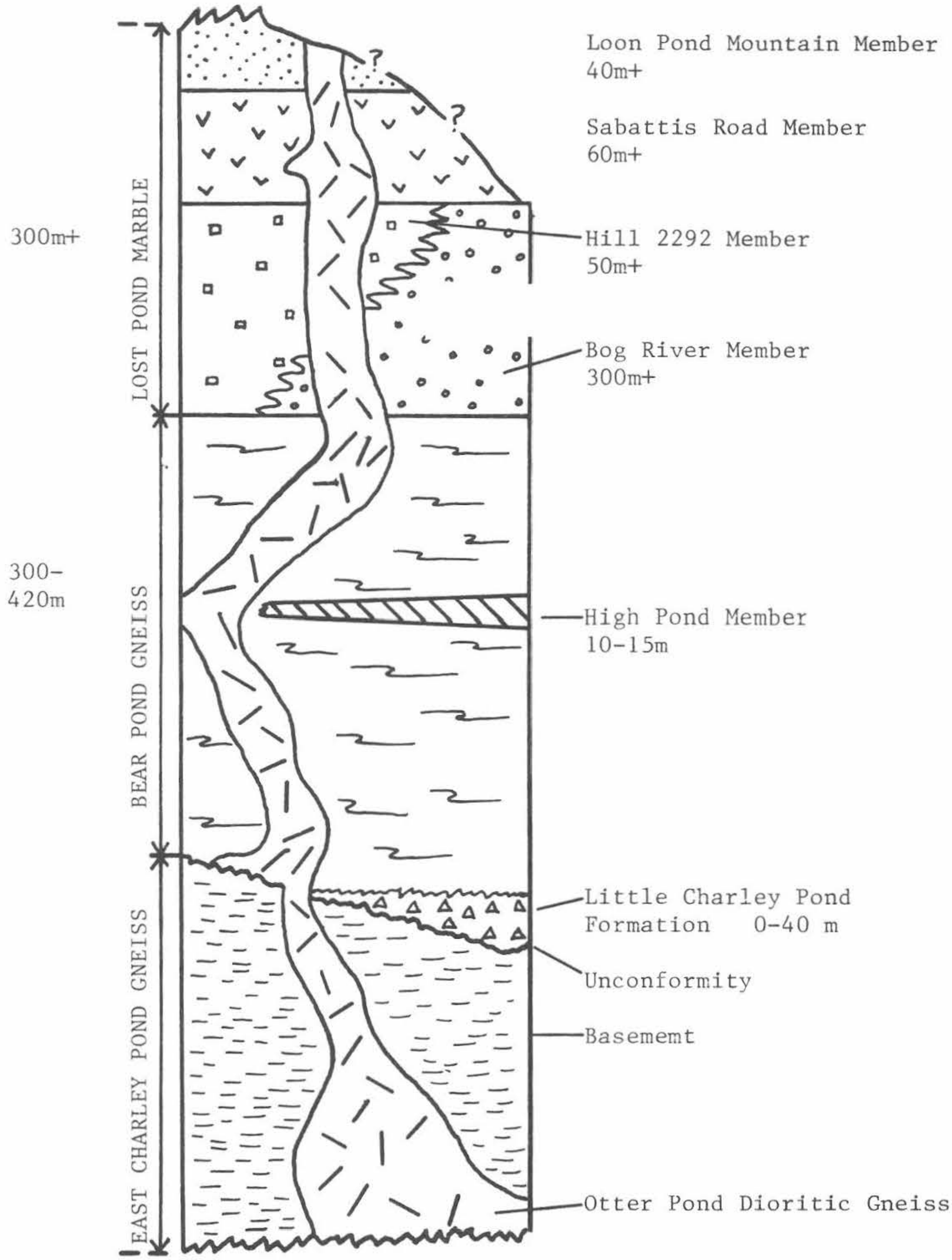


FIGURE 2

Figure 2. Generalized geologic map of the Loon Pond Syncline. Field trip stops along Routes 10 and 10 A are numbered.

Figure 3. Columnar section of rocks in the Loon Pond Syncline area. Relative ages are interpreted on the basis of structural position



to F_2. There is less compelling evidence for the igneous nature of the East Charley Pond Gneiss because crosscutting relationships are absent. The best evidence for an igneous origin are the homogeneous nature of the rock, the rather uniform rock texture and some partially assimilated quartz-rich sedimentary layers.

Several generations or a continuous sequence of granitic pegmatites are present throughout the area. The rocks are common in outcrop but not volumetrically significant. Most are a few cm thick and only a few are thicker than 10 m. They crosscut some second phase folds and are deformed by second phase folds in other places.

The structural interpretation of large-scale early folds in this area is based upon the repetition of the formations and intrusive rocks, and thus the accurate identification and local correlation of units is of greatest importance.

Basement: East Charley Pond Gneiss. This granitic igneous rock is a biotite-magnetite-plagioclase-hornblende-quartz-perthite gneiss with minor garnet in places. It is a ridge former and is exposed on some of the highest hills near East Charley Pond. The rock is considered to be basement, and no thickness has been determined. Outcrops are massive, poorly jointed, and have a homogeneous mineralogy. Coarse pink perthite commonly forms more than half of the rock, and where this occurs outcrops weather dull white or pink. Where hornblende content is more than a few percent the rock may weather brown.

A few thin veins and lenses of pegmatite with compositions similar to the rest of the pluton were noted. The thin pegmatite bodies are almost invariably parallel to foliation.

In many localities foliation is poorly developed or absent. In these places the rock has a coarse homogeneous porphyritic texture with large megacrysts of perthite. Foliation within the gneiss, well developed at some locations, was produced during the first phase of deformation.

Cover Rocks: Little Charley Pond Formation. This interlayered quartzite and diopsidic marble is exposed only in the southern portion of the field area in the vicinity of Little Charley Pond. The formation is up to 40 m thick and thins to zero both east and west along strike. The lower half of the formation consists of 12-18 cm-thick quartzites interbedded with 2-5 cm-thick beds of calcite-diopside-quartz granulite. A distinct foliation is formed by curved wisps of calcite strung out within the quartz matrix.

The upper part of the formation, more deeply pitted by weathering, is a diopsidic marble with scattered .5 m-thick beds of quartzite. This upper portion of the formation is relatively difficult to trace along strike.

Bear Pond Gneiss. The Bear Pond Gneiss, up to 420 m thick, overlies the Little Charley Pond Formation. It is extensively exposed and contains the High Pond Member. The gneiss is named for its excellent exposures at the type locality on the northwest shore of Bear Pond and along Route 10 to the west over a distance of .4 km. By correlation it is inferred to underlie the central and eastern summits of Loon Pond Mountain, and it also occurs as a layer of varied thickness that rings the central portion of the syncline. The rock is absent south of Loon Pond because of the intrusion of the Otter Pond Dioritic Gneiss.

The formation is composed of several different quartz-microcline gneisses with magnetite. Plagioclase, biotite, garnet and sillimanite are commonly present but do not occur in all of the rocks. Muscovite and chlorite are common secondary minerals.

Due to the resistance of various rock types to weathering, the Bear Pond Gneiss produces a variety of topographic features including high ridges, lakes, ponds, and swamps.

The High Pond Member, 10-15 m thick, is predominantly a quartz-rich granulite with a sugary texture. The rock contains magnetite, pyrite, garnet, biotite, sillimanite, quartz, and microcline. Where the pyrite content is high, the rock weathers a rich brown. These granulite beds of the High Pond Member closely resemble the Hill 2292 member of the Lost Pond Marble.

A garnet-quartz-augite-microcline gneiss, less commonly found in this member, is present in small dark-colored boudins. In several locations the contact between the High Pond Member and the Bear Pond Gneiss is transitional, with thin layers of augitic gneiss present in the Bear Pond Gneiss a few meters from the High Pond Member.

Lost Pond Marble. This formation, youngest of the cover rocks, overlies the Bear Pond Gneiss and is named for exposures near Lost Pond. It is divided into four members which from the base upward are: 1) the Bog River Member, a calcite-quartz-clinopyroxene marble with the type locality along the Bog River, 2) the Hill 2292 Member, a sillimanite-microcline-quartz granulite with the type locality along the steep southern slopes of Loon Pond Mountain, 3) the Sabattis Road Member, interbedded coarse graphitic marble, diopside-calcite granulite, and thin-bedded quartzites with the type locality .5 km east of Bear Pond along Route 10 (Stop 2), and 4) the Loon Pond Mountain Member consisting of massive pyritic quartzites. Not all of the members are laterally continuous (Figure 3).

The Lost Pond Marble forms two broad areas of outcrop, one surrounding Loon Pond Mountain and the other in a broad arc that includes Little Trout Pond (Figure 2). With the exception of the massive quartzite unit the formation is characterized by low topographic relief, and many of the swamps in the area are underlain by this marble-rich formation.

Intrusive Rocks: Otter Pond Dioritic Gneiss. This intrusive unit, named for its exposures along the high ridge west of Otter Pond, is extensively distributed throughout the field area. The diorite is exposed in map-scale folds in the southwestern part of the area and has a large areal exposure east of Round Lake. In addition, there are small bodies of dioritic gneiss scattered throughout the area whose discontinuous nature and spatial distribution suggest an igneous origin. The gneiss is a ridge former; outcrops are large and the rock is commonly massive. First described by Buddington and Leonard (1962), the unit has a much greater exposure than that shown on their map.

All of the rocks contain plagioclase, hornblende, and magnetite. Quartz is present in almost all of the thin sections, and biotite is also common. Other minerals not always present include augite, hypersthene, cummingtonite, and garnet. Modes for samples collected near Stop 1 appear in the Road Log.

Fresh samples are dark green, and the rock weathers to a black and white, salt and pepper texture where hornblende and pyroxene compose only a few modal percent. As mafic content increases, weathered surfaces are dark-brown and black. Foliation is well developed to absent. In most outcrops a fair to good foliation is defined by alternate layers of mafic and felsic minerals, and in some outcrops the foliation gives way to a strong lineation. In several locations foliation is absent, and rocks have a granitic texture.

Although the Otter Pond Diorite is concordant with surrounding units in most areas, crosscutting relationships can be demonstrated both at outcrop and map scales. On the ridge west of Otter Pond the hornblende gneiss cuts across the Bear Pond Gneiss at a low angle to layering. At Stop 1 along Route 10, a dike of hornblende diorite cuts the older, foliated dioritic gneiss.

METAMORPHISM

Regional metamorphism in the Loon Pond Syncline area is of the hornblende-granulite facies. Silver (1968) demonstrates that rocks now exposed in the Highlands underwent granulite-facies metamorphism between 1100 and 1020 million years ago. McLelland and Isachsen (1980) suggest that the metamorphism was caused by a doubling of crustal thickness that accompanied the Grenville Orogeny.

Timing of Metamorphism With Respect to Deformation. The dominant foliation, present nearly everywhere except near F_2 axial surfaces, is the product of the F_1 deformation. Extensive recrystallization of the rocks associated with F_1 formed texturally and mineralogically distinct layering on a 1 cm to 1 m scale, especially in the Bear Pond Gneiss. These layers are folded by subsequent deformations.

The second phase of deformation formed a pervasive mineral lineation in the form of sillimanite and streaks of biotite and garnet. The formation of at least some of the garnet lasted longer than that of the sillimanite, and garnet can be seen completely enveloping sillimanite grains in thin section. The recrystallization of magnetite and pyrite also postdates the growth of sillimanite. Both opaque minerals are seen in the field as coatings on large sillimanite needles.

The F_3 phase is associated with a weakly developed mineral lineation, and the F_4 deformation did not involve any observable recrystallization.

STRUCTURAL GEOLOGY

The concentric outcrop pattern (Figure 2) gives the initial impression that the basin is structurally simple, but stratigraphic correlation and field relationships indicate that the rocks have undergone a period of isoclinal folding that was associated with the development of minor structures and large nappes. Subsequent lesser deformations have produced the basin and many of the minor structural features. Deformation of the rocks in the Loon Pond Syncline has produced a pervasive northeast-southwest structural grain which is in contrast to the east-west strike and southerly dip of the rocks surrounding the structure.

Minor Structural Features: Bedding and Foliation. In the Bear Pond Gneiss, textural and mineralogical variations appear to represent relict bedding. In the Lost Pond Marble, interbedded diopside-calcite granulites and quartzites may represent original siliceous dolomite and sandstone beds. No primary sedimentary structures other than bedding are evident, and thus there is no sedimentary evidence for the direction of primary tops.

Foliation is expressed by: 1) individual biotite flakes, 2) sheets and lenses of quartz and sillimanite, 3) thin quartz discs in marble and 4) flattened and elongated granular aggregates of biotite, garnet, and quartz. Foliation and relict bedding are parallel except in hinge areas of early folds, and thus the major foliation is related to the first phase of deformation. A more subtle foliation is present in the axial regions of some second phase folds.

Early Folds. A few early isoclinal folds were found. Relict bedding defines folding, and axial planar foliation is defined by biotite grains. The folds are only a few cm across and in a few localities are refolded. Fold axes may be parallel to those of the second phase.

Second Phase Folds. These open to tight minor folds are related to the southwest-plunging map-scale folds. They fold the dominant foliation and in some cases display an axial planar foliation. Axial surfaces dip southeast and the folds plunge southwest except where modified by subsequent deformation. The consistent southwest plunge is accompanied by a strong parallel mineral lineation.

Third Phase Folds. These open to tight folds are well developed on Loon Pond Mountain and on the shores of Loon Pond. Here axial surfaces dip northwest and plunges range from northwest to southwest.

Fourth Phase Folds. These broad, gentle folds trend north-south and have vertical axial surfaces. They have wavelengths of up to 2 m in outcrop.

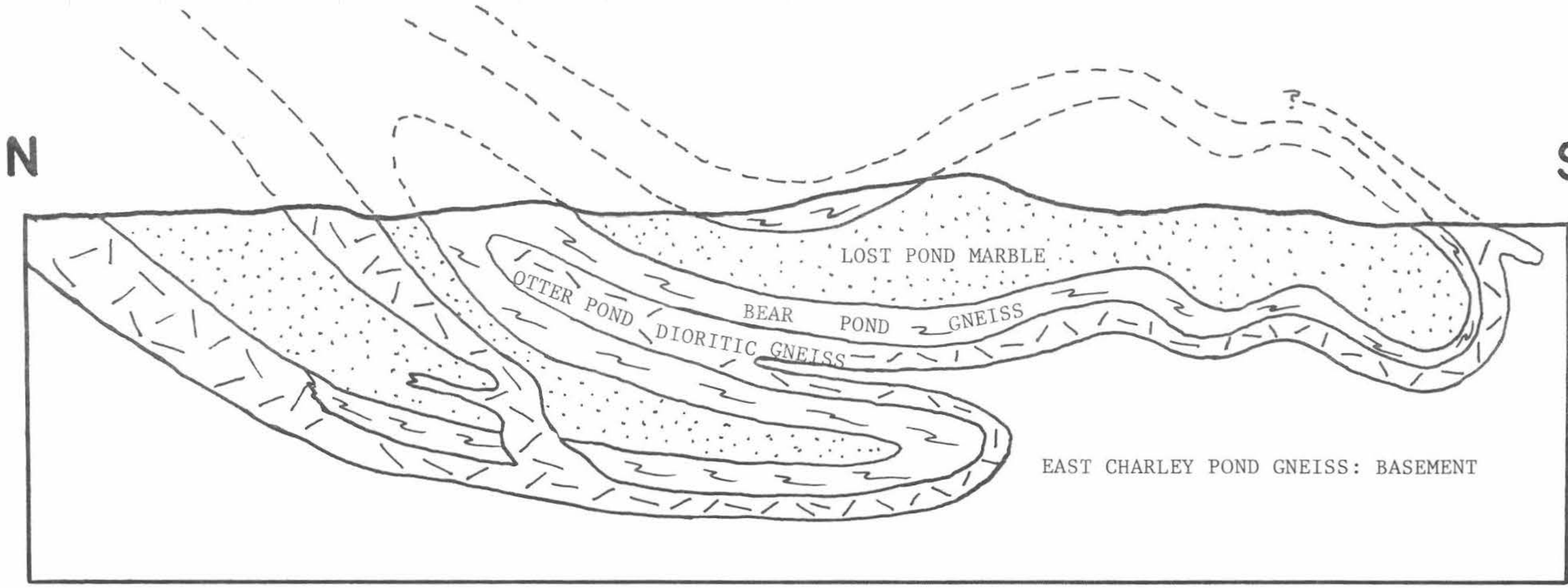


FIGURE 4

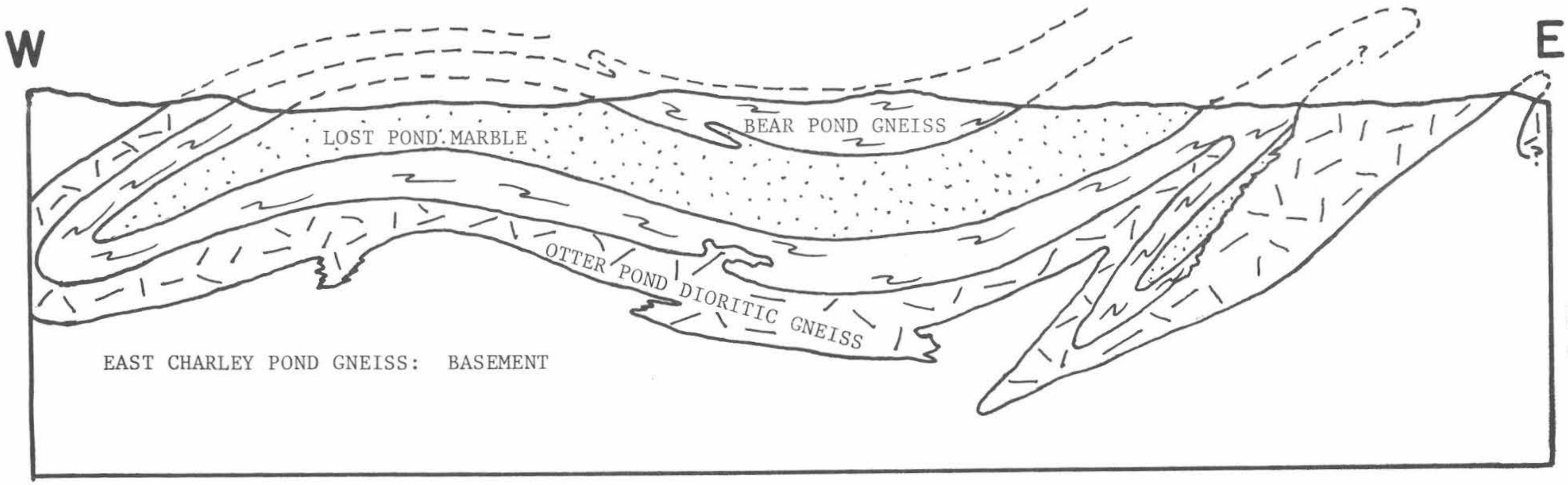


Figure 4. Generalized cross sections, east-west and north-south, across the Loon Pond Syncline. Not to scale.

Mineral Lineations. Most of the mineral lineations measured are sillimanite. Sillimanite needles and quartz form sheets up to 1 cm thick that define both foliation and lineation in the Bear Pond Gneiss. Magnetite, biotite, and hornblende also form clear lineations, as do streaks of garnet and quartz. The mineral lineations show a strong southwest-plunging concentration.

Additional linear features include quartz rods, feldspar rods, and boudin axes. The rods, a few mm to 2 cm wide and up to 1 m long may represent fold hinges separated from limbs by shearing. When rods occur with a mineral lineation in outcrop, the two are parallel.

Major Folds. Several major fold systems (structures of the same generation) and a fold complex (a map scale interference pattern of more than one generation of folds) are present.

The Sabattis Fold System consists of map scale and minor folds of the F_2 phase. Four overturned anticlines and corresponding synclines deform the Bear Pond Gneiss and the Otter Pond Dioritic Gneiss. The first syncline south of Sabattis is best exposed, with excellent outcrop control in the hinge area. Minor structural features, including folded quartz veins, feldspar rods, sillimanite and magnetite needles all closely mimic the planar and linear elements of the major folds.

The Trout Pond Fold System is also a product of the F_2 deformation. The system is a 3.5 km-long series of folds within one of the Otter Pond Diorite bodies. The folds, located southwest of Trout Pond, differ from the Sabattis Fold System in that most of them do not have overturned limbs. The southwest-plunging mineral lineation is present, however.

The Loon Pond Mountain Fold Complex is centered on the three peaks of Loon Pond Mountain. The complex has been affected by all four deformations.

F_1: The repetition of the Bear Pond Gneiss near the center of the structure is caused by the erosion of a large-scale fold of nappe proportions (Figure 4). Other evidence for the F_1 phase includes the digitations of the Bear Pond Gneiss - Lost Pond Marble contact at the center of the structure.

The second phase is responsible for the northeast trend of the complex. The major F_2 fold here is not overturned, although many minor structures of this generation have limbs overturned to the northwest. Minor structures of the F_2 phase plunge southwest except where modified by the F_4 phase.

Excellent examples of the third phase folds are found on the south-facing slopes of Loon Pond Mountain. Axial surfaces dip northwest. Fourth-phase folds are best expressed by the gentle warping in section E-W, Figure 4.

The Loon Pond Mountain Syncline is a basin formed by the interference of second, third(?), and fourth-phase synclines. Although many of the minor second and third phase folds have overturned limbs, the map-scale folds of these two phases at the present level of erosion are not characterized by overturned limbs near the center of the basin.

ROAD LOG

Miles

0.0 Depart Long Lake public beach and proceed north on Route 30.

4.4 Gatehouse for Whitney Park on left.

6.6 Turn left towards Sabattis on Route 10A.

8.9 View of Loon Pond Mountain to the northwest.

9.3 Small outcrops of East Charley Pond Gneiss on both sides the road.

9.5 Cross bridge at the head of Little Tupper Lake.

9.6 Turn left towards Sabattis on Route 10.

10.6 Stop 1. Otter Pond Dioritic Gneiss is exposed along the north side of the road. Near the east end of the outcrop, a late 1 meter-thick vertical dike cuts the foliated gneiss. Discontinuous quartz veins help define the edges of the north-trending dioritic dike.

Estimated modes from thin sections of the foliated gneiss at this outcrop (646) and one .3 miles east (647) follow:

	646	647
Quartz	1	12
Plagioclase	71	43
An percent	(42)	(38)
Biotite	-	1
Augite	13	-
Hornblende	10	36
Apatite	tr.	tr.
Zircon	1	tr.
Magnetite and Ilmenite	4	8

Foliation is N84W,24SW and parallels foliation in basement rocks to the south. The outcrop was mapped by Buddington and Leonard (1962) as a diorite gneiss, but they did not consider the unit to be as widespread as is shown in Figure 2.

10.9 Whitney Headquarters sign.

ROAD LOG

Miles

14.9 Watchung Boy Scout Camp sign.

15.1 Stop 2. The Sabattis Road Member of the Lost Pond Marble is exposed on the north side of the road. Most of the outcrop consists of brown-to gray-weathering marble beds up to 2 m thick. A few layers of quartzite, .5 m thick, are also present. The graphitic marble includes quartz, microcline, pale green diopside, phlogopite, calcite, sphene, and scapolite. The quartz and diopside stand out as small knobs and clusters on the weathered surface.

F_2 fold axes plunge S45W at about 10 degrees, and minor undulations in bedding define a pronounced lineation in places.

15.4 Step 3. The Bear Pond Gneiss is exposed along the northwest shore of Bear Pond. Layering and foliation strike northeast, parallel to the trend of the pond, and dip southeast beneath the Lost Pond Marble of Stop 2. Grain size in the magnetite-biotite-quartz-microcline gneiss varies from fine to pegmatitic.

A biotite-rich unit of the Bear Pond Gneiss lies on the north side of the road .1 mile west of Bear Pond. The gneiss is cut by a coarse microcline pegmatite body that is roughly concordant to foliation.

Near the crest of the hill .2 miles west of Bear Pond a third example of the Bear Pond Gneiss is exposed. The rock is a magnetite-biotite-zircon-garnet-sillimanite-microcline-plagioclase-quartz gneiss. Locally, both sillimanite and coarse garnet constitute more than 10% of the rock. A strong F_2 sillimanite lineation plunges S46W at 4 degrees. Garnet distribution in this "chicken pox" rock is locally random, but in some places the garnet is concentrated in layers parallel to foliation.

16.3 Swamp to north is underlain by Lost Pond Marble.

16.4 Bear Pond Gneiss exposed on left side of road.

17.6 Turn left on dirt road to Charlie Pond Club land. Proceed 100 yards to gate. This area is not generally accessible to the public. Proceed .1 mile to a log landing on the west side of the road. Stop 4. Beginning at the southwest corner of the log landing, walk 400 yards at S40W across uneven terrain to the outlet of a small pond. Here the Bear Pond Gneiss is exposed in the axial region of a map-scale F_2 syncline. Both limbs strike NE and dip SE. Many minor structures exposed along the outlet illustrate the second phase of deformation. Fold axes and mineral lineations form a tight southwest-plunging concentration on an equal area net. If water levels permit, a faint F_3 magnetite lineation will be seen.

The Loon Pond Dioritic Gneiss is exposed on the other side of the outlet .1 mile southwest along the edge of the pond.

17.8 Return to Route 10. Turn left.

18.0 Bear Pond Gneiss on south side of road.

18.2 Abandoned Sabattis Railroad Station. Turn around and return to Little Tupper Lake.

19.8 Trout Pond trailhead on north side of road.

26.4 Turn left at fork and proceed northeast along Route 10A to Route 30.

29.5 Intersection with Route 30. Turn left.

29.6 Stop 5. Overgrown quarry on west side of road. The old quarry is in the Bog River Member of the Lost Pond Marble and contains graphite, calcite, sphere, phlogopite, quartz, calcite, and diopside. Diopside varies in color from pale to dark green and locally forms more than 50% of the rock. The rock is moderately well to poorly foliated, and grain size varies from medium to coarse. There is great variation in the relative proportions of calcite and quartz in the rock.

Rocks at this location bear a strong resemblance to the diopsidic marbles of the Sabattis Road Member of the Lost Pond Marble (Stop 2).

End of this portion of the trip. Turn around and proceed south to Long Lake public beach (mile 40.1) or proceed to next leg of field trip.

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TRIP AB-2

BEDROCK GEOLOGY OF THE GRAMPUS LAKE AREA, LONG LAKE QUADRANGLE, NEW YORK

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INTRODUCTION

What follows is a brief description of the geology of the Grampus Lake area. These are the preliminary results of a study conducted under the guidance of Professor Leo M. Hall at the University of Massachusetts. The map area lies in the southwest part of the Long Lake quadrangle and includes a portion of the southeast corner of the Tupper Lake quadrangle (see Figure 1). Previous work in the area was done by Cushing (1907) in the Long Lake quadrangle and Buddington and Leonard (1962) in the Tupper Lake quadrangle.

Stratigraphy

Stratified rocks in the Grampus Lake area include quartz-rich calcareous gneisses, quartzites, and subordinate amounts of pelitic gneisses and marbles, locally interlayered with granitic gneisses. These rocks overlie a basal gneiss, undivided in this study, which includes granitic, charnockitic and syenitic gneisses. Metamorphosed intrusive rocks are abundant in the area (see Figure 2).

Basal gneiss

This unit consists of undivided granitic, charnockitic, and pyroxene syenitic gneisses. The granitic gneiss is a pinkish gray or tan weathering hornblende-biotite-othoclase-microperthite granitic gneiss. It is commonly felsic, with less than four percent dark minerals, and is poorly foliated. The pyroxene microperthite syenitic gneiss is well foliated. It weathers orange, buff, or tan. Aggregates of feldspar and quartz, or microperthite augen lie in the foliation. An orange brown rind or "maple sugar" brown weathering is common. A well foliated charnockitic gneiss is present in the basal gneisses. This is a deeply weathering orange-brown or gray gneiss. Lenses of coarse quartz and feldspar are aligned in the foliation. A biotite-antiperthite gneiss weathers gray, pink, or tan and is poorly foliated. It is commonly equigranular. Locally abundant feldspar augen are present.

Interlayered granitic and quartz granular gneisses

The basal gneisses are overlain by pink weathering, locally garnetiferous granitic gneisses interlayered with subordinate amounts of quartz granulites. The granitic gneisses occur in well-foliated or layered textures and in more massive equigranular textures. Layers of pink and white-weathering feldspar, coarse quartz which forms ribs on the outcrop

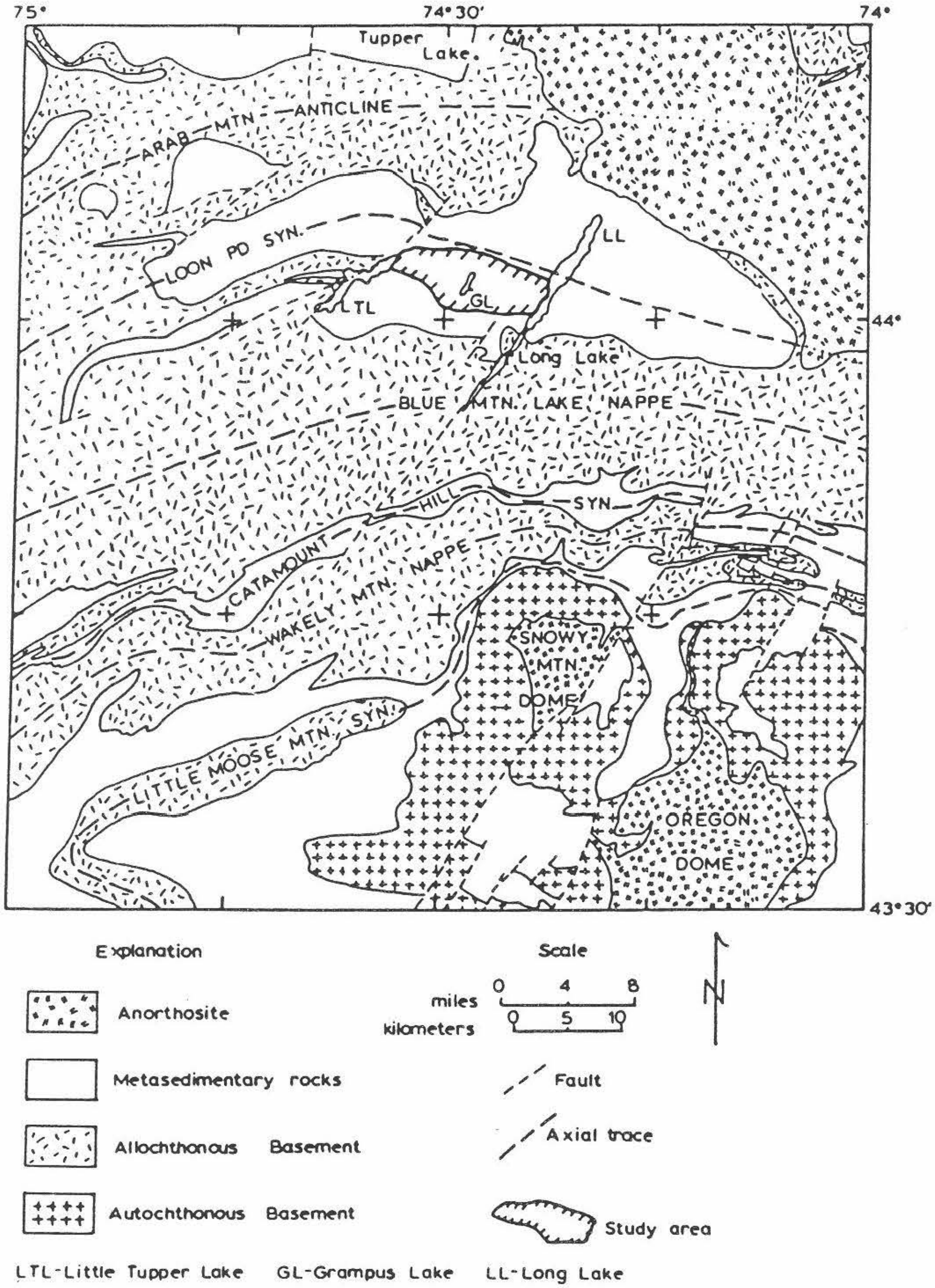

Figure 1. Generalized regional geology of a portion of the central Adirondacks. Compiled from Walton and DeWaard (1963), Fisher et al. (1971), McLelland and Isachsen (1980), Turner (1980), and Ollila (pers. comm., 1981).

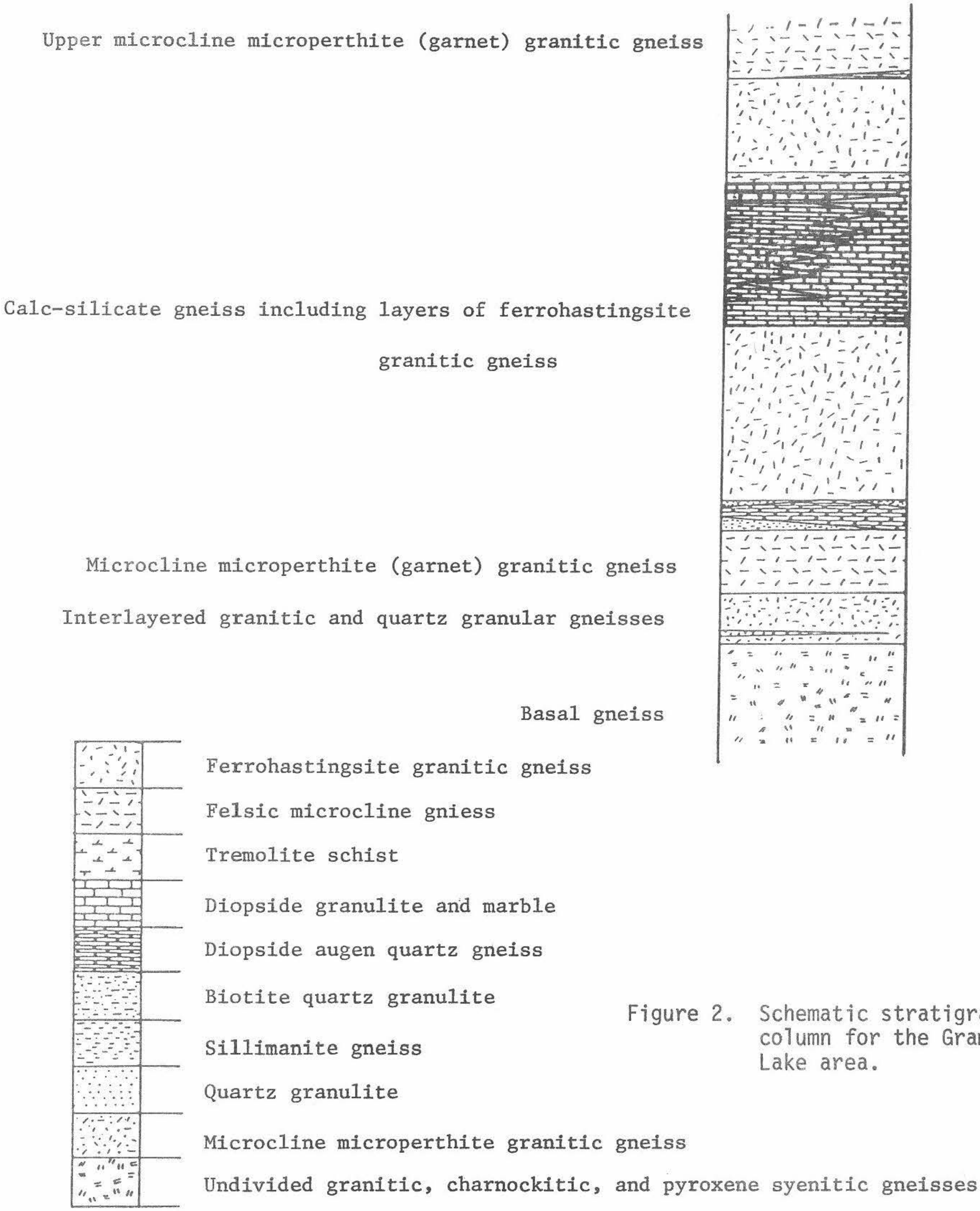

Figure 2. Schematic stratigraphic column for the Grampus Lake area.

surface, and anastomosing layers of hornblende are locally present. Thin amphibolites are locally present, as are leaves of bluish-gray quartz in the felsic gneisses. The massive granitic gneiss is a moderately fine-grained equigranular rock.

Microcline microperthite (garnet) granitic gneiss

This gneiss weathers pink, brown, or gray. A moderately fine-grained equigranular texture is most common. This texture consists of interlocking microcline and quartz grains with tiny disseminated garnet present locally. In places the gneiss assumes a more granitic, even pegmatitic, appearance. Augen of pink microcline and coarse leaves of bluish quartz are locally present.

Calc-silicate gneisses

The most abundant lithic type in the calc-silicate gneisses is diopside augen quartz gneiss. Diopside augen, commonly several centimeters long, weather into the outcrop leaving a characteristic rough network of quartz selvages. Glassy quartzites, of similar mineralogy but lacking diopside augen, are also present. Scapolite is common in this rock. Wollastonite is rare. Diopside granulites are abundant in the upper part of the unit. These are commonly buff or rusty weathering, poorly foliated rocks consisting of diopside and microcline, diopside and quartz, or diopside and calcite. Microcline augen are present locally. Gray weathering gneisses consisting of diopside and phlogopite and others consisting of tremolite, phlogopite, and diopside are found in the upper part of the unit. Rare outcrops of very coarse marbles and of sugary textured, fine-grained marbles are present. Minor biotite-feldspar-quartz granular gneiss is present in the lower part of the unit. This is a fine-grained rock which weathers black and commonly is thinly layered. Sillimanite-rich gneisses are also present in the lower part of the unit.

Upper microcline microperthite (garnet) granitic gneiss

This is a felsic, pink weathering, poorly foliated gneiss. Small amounts of biotite and locally hornblende are present. The gneiss is locally garnetiferous. In places interlayered quartz granulites are present.

Metamorphosed Igneous Rocks

General Statement

Metamorphosed plutonic igneous rocks include ferrohastingsite granitic gneiss, pyroxene quartz syenitic and augite syenitic gneisses, well foliated pyroxene gneiss, pyroxene (garnet) gneiss, gabbroic gneiss, and andesine augen dioritic gneiss. All are moderately to well foliated and commonly appear concordant to bedding and the early foliation in the stratified rocks.

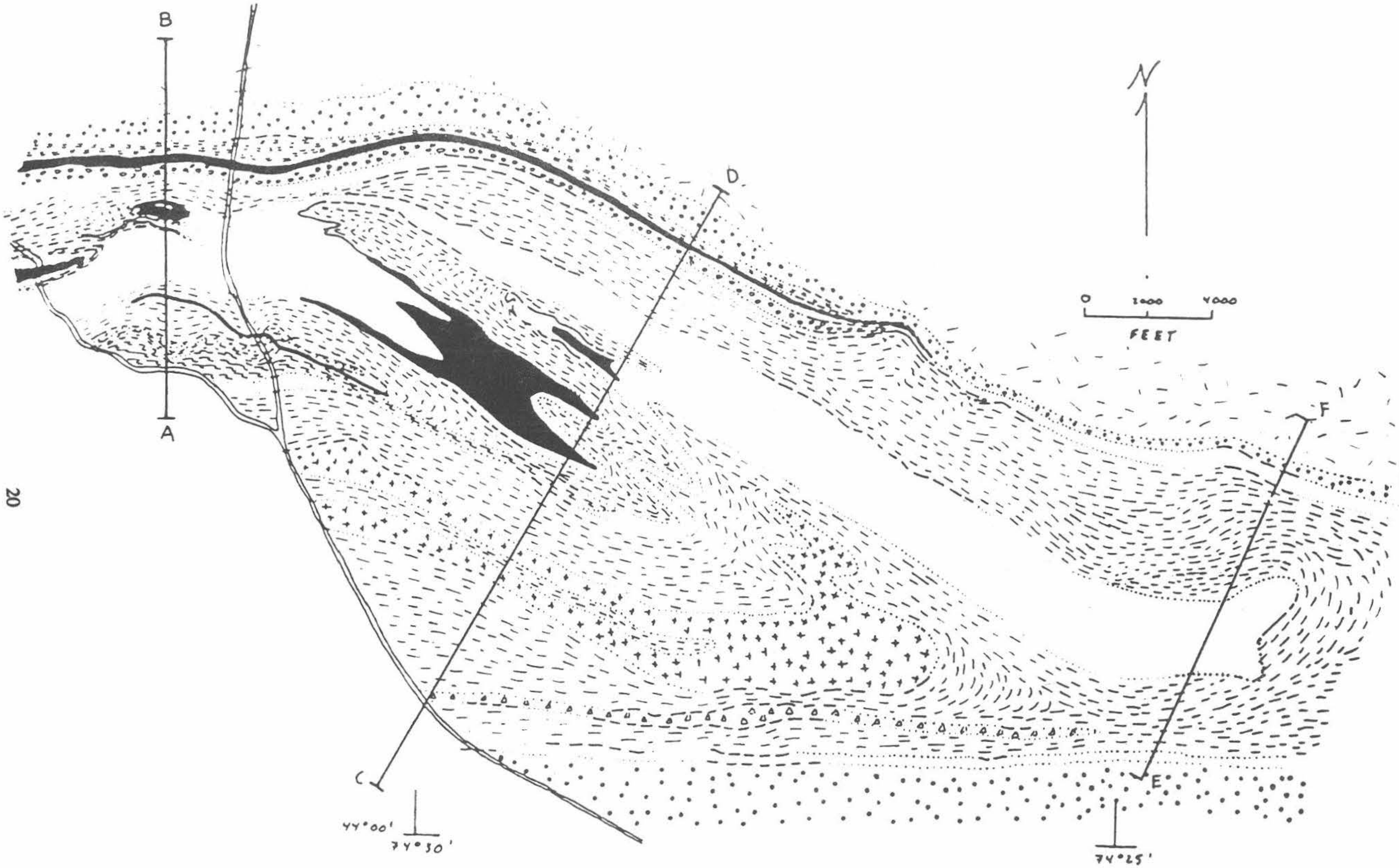

Figure 3. Geologic map of the Grampus Lake area.

Stratified Rocks

Upper microcline microperthite (garnet) granitic gneiss

Calc-silicate gneisses

Lower microcline microperthite (garnet) granitic gneiss

Interlayered granitic and quartz granular gneisses

Basal gneisses

Contacts: Known

Approximate

Inferred

Generalized local strike and dip of bedding or regional foliation 60

Intrusive Rocks

Andesine augen dioritic gneiss

Well foliated pyroxene gneiss

Pyroxene (garnet) gneiss

Augite syenitic gneiss

Pyroxene quartz syenitic gneiss

Ferrohastingsite granitic gneiss

Figure 3. (continued) Explanation of geologic map.

Ferrohastingsite granitic gneiss

This is a moderately to poorly foliated pink, tan, or gray weathering gneiss. Sparse biotite is an additional mafic mineral. Augen of microperthite under 1 cm in diameter are common. Where observed, the contacts and foliation in the granitic gneiss are parallel to the compositional layering or foliation in the metamorphosed sedimentary rocks, which is associated with the early deformation.

Pyroxene quartz syenitic gneiss

A hornblende ± augite ± hypersthene syenitic or quartz syenitic gneiss. The rock weathers white or tan and displays the characteristic "maple sugar" brown weathered rind. K-feldspar is orthoclase microperthite or mesoperthite which is olive green on a fresh surface. A strong linear fabric of coarse hornblende is locally present. The foliation in this gneiss is folded by third phase hinges, therefore the intrusion of the syenite must have occurred before or during the second phase of deformation.

Augite syenitic gneiss

The augite syenitic gneiss is a gray weathering moderately well foliated gneiss which commonly contains abundant discontinuous layers of coarse feldspar. A felsic and locally massive granitic gneiss is locally interlayered in the syenitic gneiss. Where both are present, an apparent gradational relationship obscures the distinctions between the two. The augite syenitic gneiss, however, is more strongly foliated, has a higher color index, and commonly shows dark colored feldspar, either dark greenish brown or reddish brown, on a fresh surface. The granitic gneiss is commonly more massive, more felsic, and lighter in color on a fresh surface. The augite syenitic gneiss predates third phase deformation.

Gabbroic gneiss

The gabbroic gneiss is a black weathering rock which contains plagioclase, augite, and hypersthene in a subophitic texture. Hornblende, biotite, and garnet form corona textures. Massive and well foliated varieties are present. Thin sheets of gabbroic gneiss are present in the basal gneiss and are not shown on the geologic map. The contacts of the gabbroic gneiss locally truncate the regional foliation.

Pyroxene (garnet) gneiss

The garnetiferous variety is a white and black weathering moderately well foliated rock. Clots of pinhead garnets 1 cm to 3 cm across are common. Plagioclase and hornblende are commonly equigranular and medium grained. Angular augen of plagioclase are present locally. The variety which lacks garnet contains elongate patches of fine grained black hornblende and augite which weather out in slight relief against the white, equigranular andesine. Centimeter size augen of andesine are present.

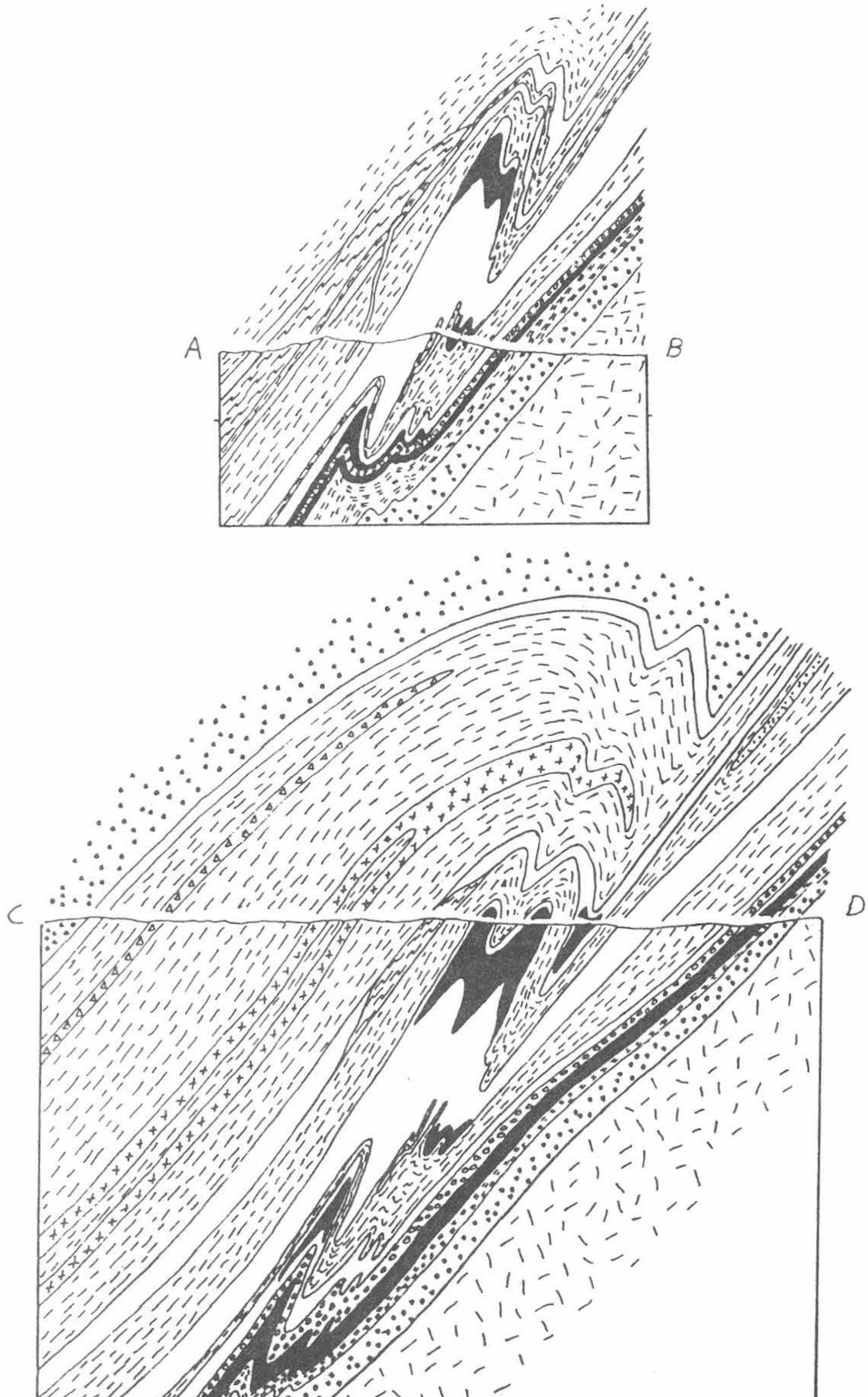

Figure 4. Geologic cross sections.

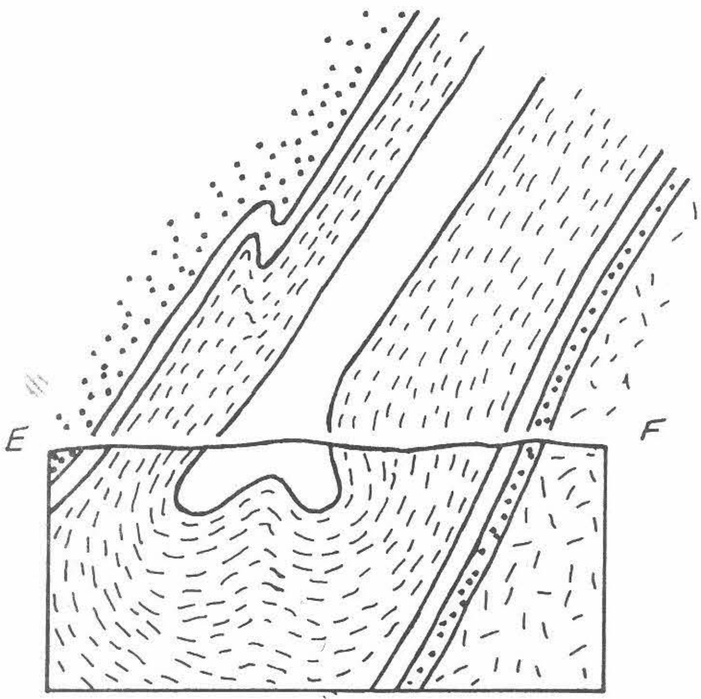

Figure 4. Continued.

Well foliated pyroxene gneiss

The well foliated pyroxene gneiss is black and tan weathering, locally with a pinkish caste. More mafic portions weather black and orange while felsic parts are beige. This unit is characterized by strong foliation and locally abrupt variation in color index. The timing of intrusion of this gneiss is uncertain.

Andesine augen dioritic gneiss

A black and white weathering, well foliated hornblende ± augite andesine dioritic gneiss which commonly intrudes as thin layers and in thicker layers which are shown on the geologic map. White weathering augen of andesine commonly comprise up to thirty percent of the rock. Hornblende is the major dark mineral, clinopyroxene is subordinate, and orthopyroxene is locally present. A thin sheet of this gneiss truncates a second phase fold. The gneiss appears to predate third phase deformation.

Structural Geology

General Statement

The geologic map and structure sections of the Grampus Lake area are dominated by the hook patterns formed by interference of second and third phase folds (see Figures 3 and 4). Gentle folding on northerly trending axial surfaces is late. Evidence of four phases of folding has been observed in the area.

No map scale early fold has been observed in the study area. The prominent regional foliation is axial planar to minor first phase folds. These folds are tight isoclinal folds with narrow, pointed profiles.

Second phase folds are isoclinal folds in the regional foliation which have rounded hinges and lack a prominent axial plane feature. Second phase axes plunge moderately to the south at high angles to the trend of the axial surfaces. The Rock Pond Syncline is a major second phase fold with an axial trace 20 km long within the area (see Figure 5).

Third phase folds have tightly appressed profiles and an intense axial plane foliation developed locally in the hinge area. Axes plunge gently southeast. Axial surfaces trend west or northwest and dip steeply south. The prominent convex up shape of the structure (see Figure 4) and the local trends of contacts and foliations are due to the third phase folds.

Fourth phase folds are locally developed, low amplitude folds with open parallel profiles. Axial surfaces trend northerly and have steep dips. Local changes in plunge are caused by fourth phase folding.

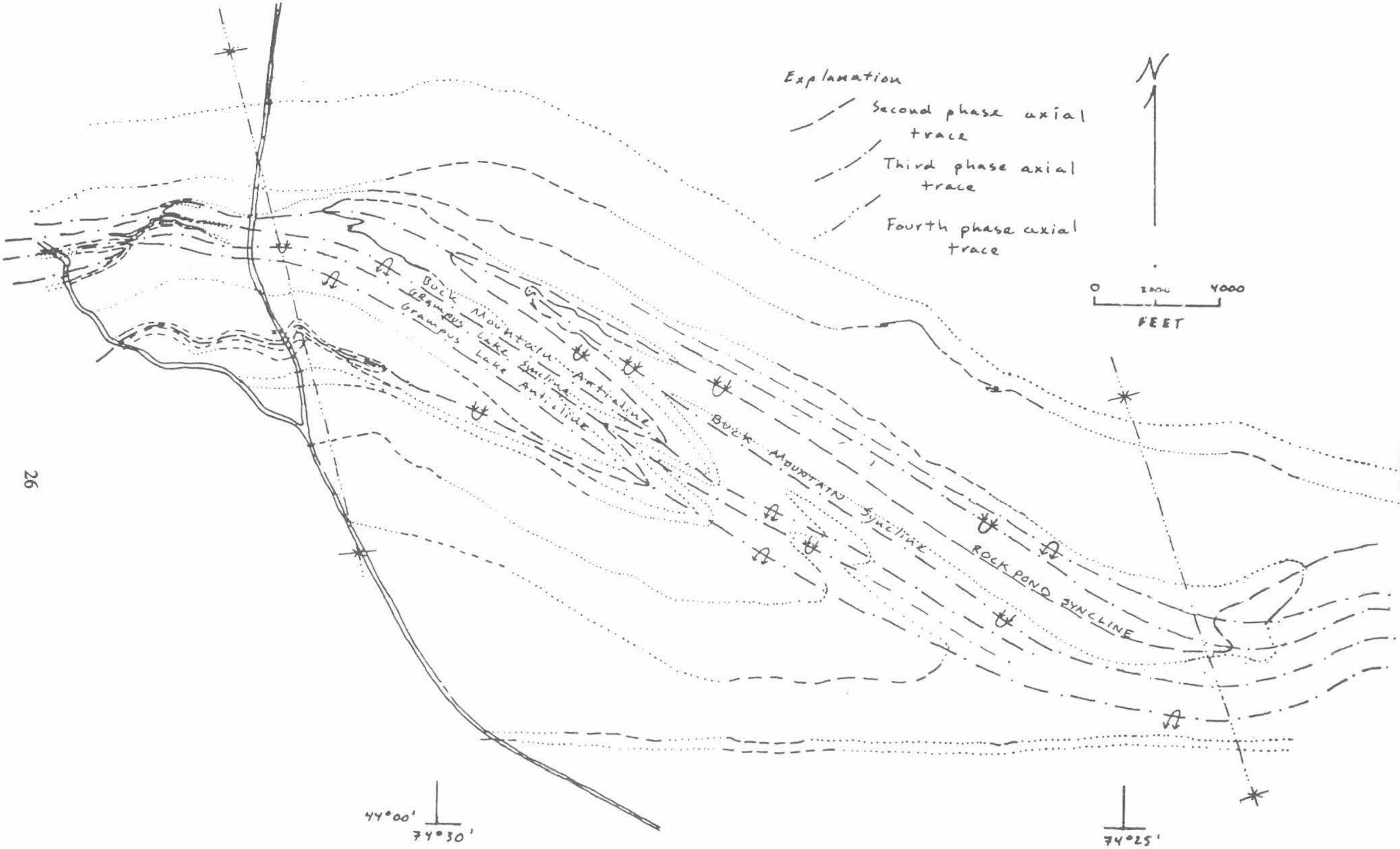

Figure 5. Generalized geologic map showing locations of axial traces of major local folds.

Major Folds

Map-scale folds of the second, third, and fourth phases are present in the area. The Rock Pond Syncline is a large second phase fold. The limbs of this fold dip moderately to steeply south and the axes plunge south or southwest. The fold is approximately reclined over much of the area. The axial trace of this fold is shown on Figure 5. In the east, the core of fold is occupied by diopside granulites. Further west and higher in the section, diopside quartz gneisses become more abundant. The fold is refolded by third phase folds in the central part of the area (Figures 3 and 4). The granitic gneiss which occupies the core of the fold in this region is interpreted as a folded layer, thinner on the southern limb of the fold than on the northern. A thin layer of quartzites and diopside quartz gneisses sporadically occurs between the granitic gneiss and the overlying microcline microperthite (garnet) granitic gneiss which lies in the core of the Rock Pond Syncline in the western part of the area.

Pairs of synclines and anticlines associated with third phase folding have axial traces which cross the area from northwest to southeast (see Figure 5). These folds have axial surfaces which dip steeply south and axes which plunge gently southeast. In the center of the map area a hook pattern results from interference of third phase folds and the Rock Pond Syncline. The convex-up shape of the structure and the orientations of strikes and dips in the area are due to third phase deformation (see Figure 4).

The axial traces of only two fourth phase folds are shown on Figure 5. Other map scale fourth phase folds are known or inferred in places. These folds dip steeply to moderately southwest. Plunges to the south are due to the orientation of the regional foliation. In this interpretation, fourth phase folds are responsible for gradual changes in strike of bedding and foliation shown on the geologic map (Figure 3).

Metamorphism

Mineral assemblages present in the rocks of the Grampus Lake area record metamorphism in the intermediate pressure granulite facies (Green and Ringwood, 1967). Temperature estimates of 730 ± 30°C have recently been made by garnet-clinopyroxene exchange thermometry on a sample from the nearby Newcomb quadrangle (Johnson _et al_. 1983). Pressures of 7.7 ± 1 kb have been estimated from the _equilibrium_ forsterite + anorthite = garnet in the metagabbro from the Newcomb quadrangle (Johnson and Essene, 1982). Mineral assemblages in the study area are consistent with these estimates. Undersaturated rocks contain the assemblage clinopyroxene-almandine, consistent with the clinopyroxene almandine isograds of Buddington (1963) and DeWaard (1968). Sillimanite is the stable aluminosilicate. Kyanite has not been found. The assemblage diopside-calcite-quartz is common in the calc-silicate gneiss. The assemblage wollastonite-calcite-quartz is observed at one locality (Stop 1). Disequilibrium textures present there suggest that wollastonite may have formed at fairly constant, moderate temperatures as fluid composition became progressively more water rich. The same assemblage with the addition of graphite and sulfides is present at another location (Stop 2). The buffering effect graphite exerts on fluid composition suggests a possible lower limit of temperature, given a pressure (Figure 6).

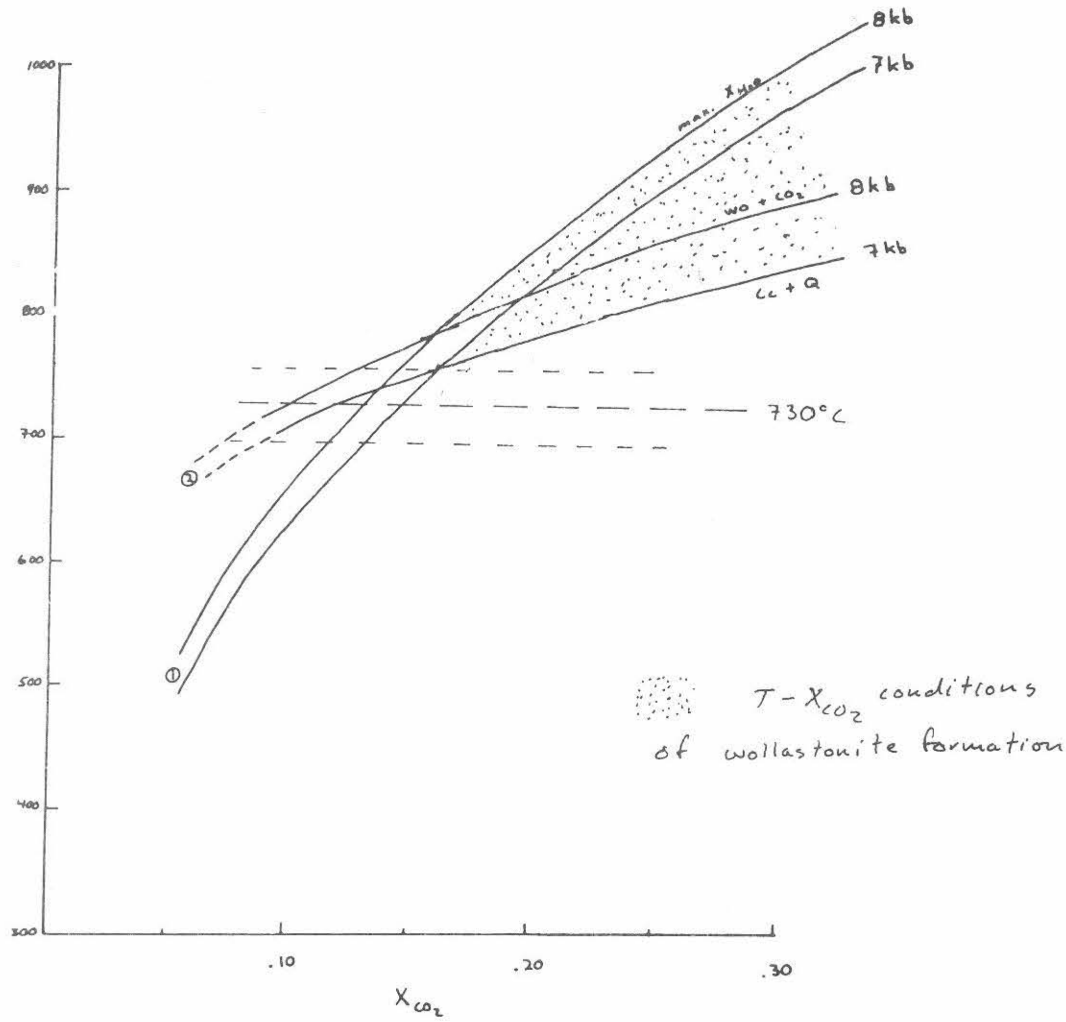

Figure 6

1.) Ohmoto and Kerrick, (1977). 2.) calculated from data of Yu. P. Mel'nik (1972) and G. B. Skippen, (1977). 3.) Johnson <u>et al</u>., (1983).

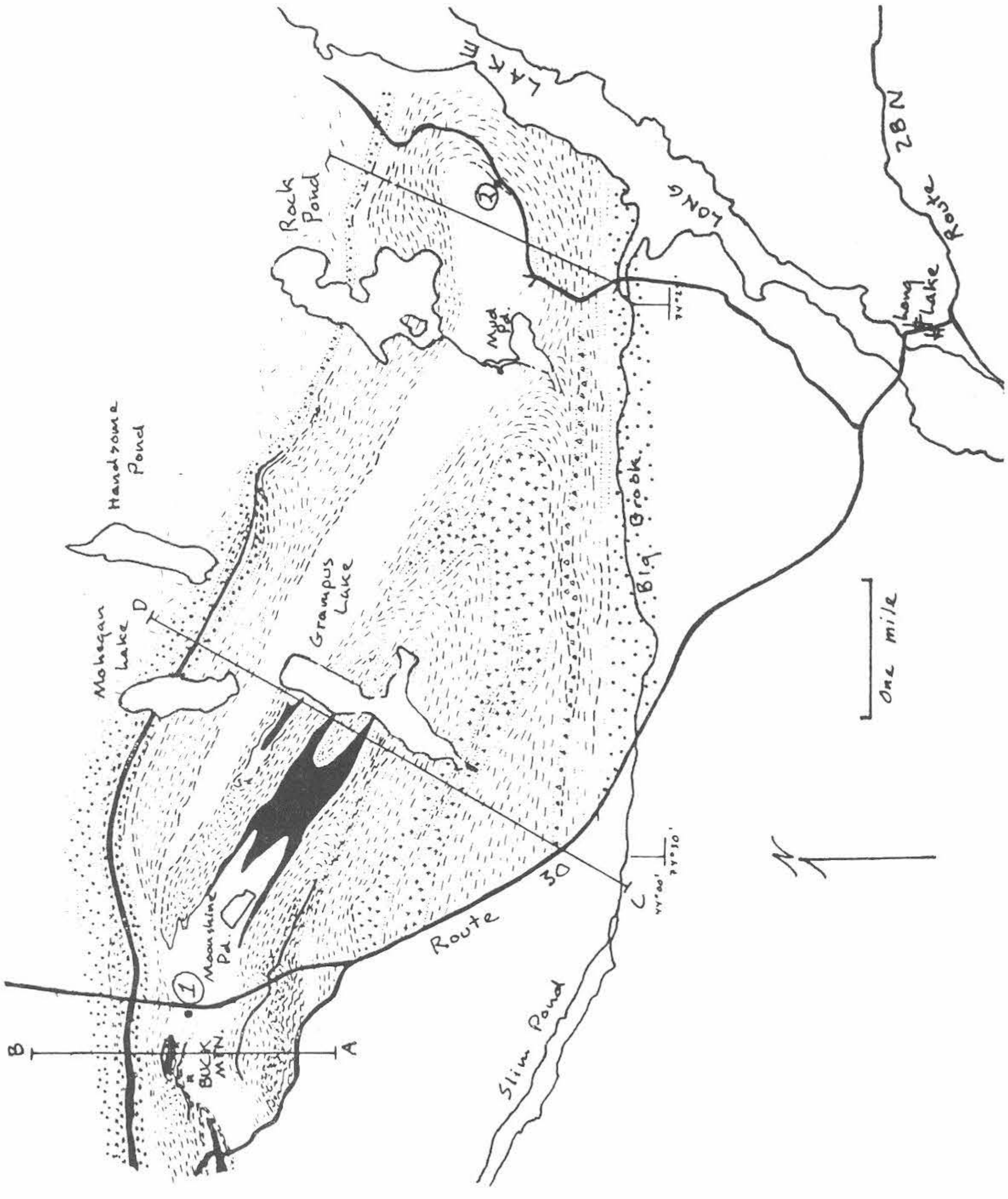

Figure 7. Field trip stops in the Grampus Lake area.

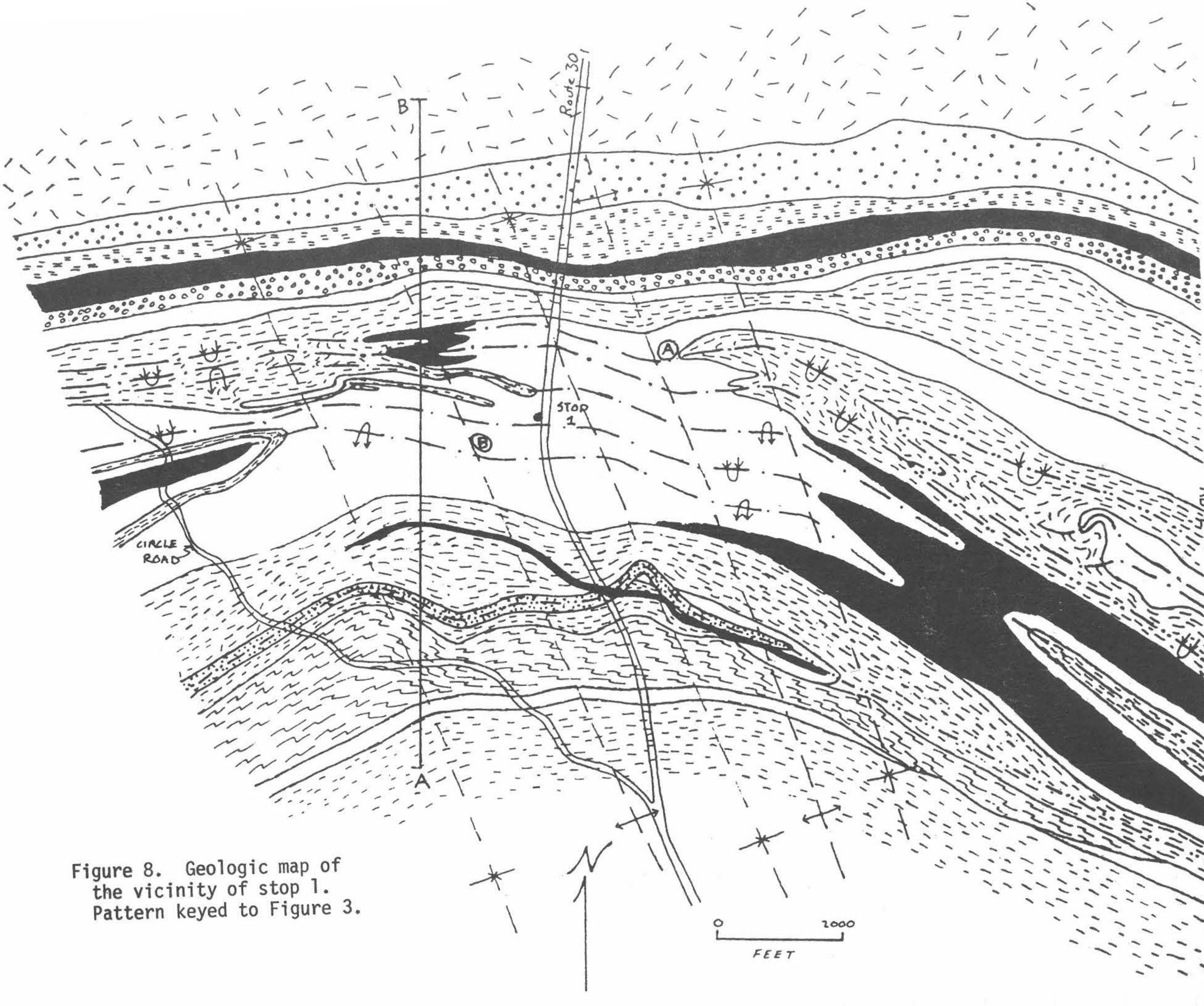

Figure 8. Geologic map of the vicinity of stop 1. Pattern keyed to Figure 3.

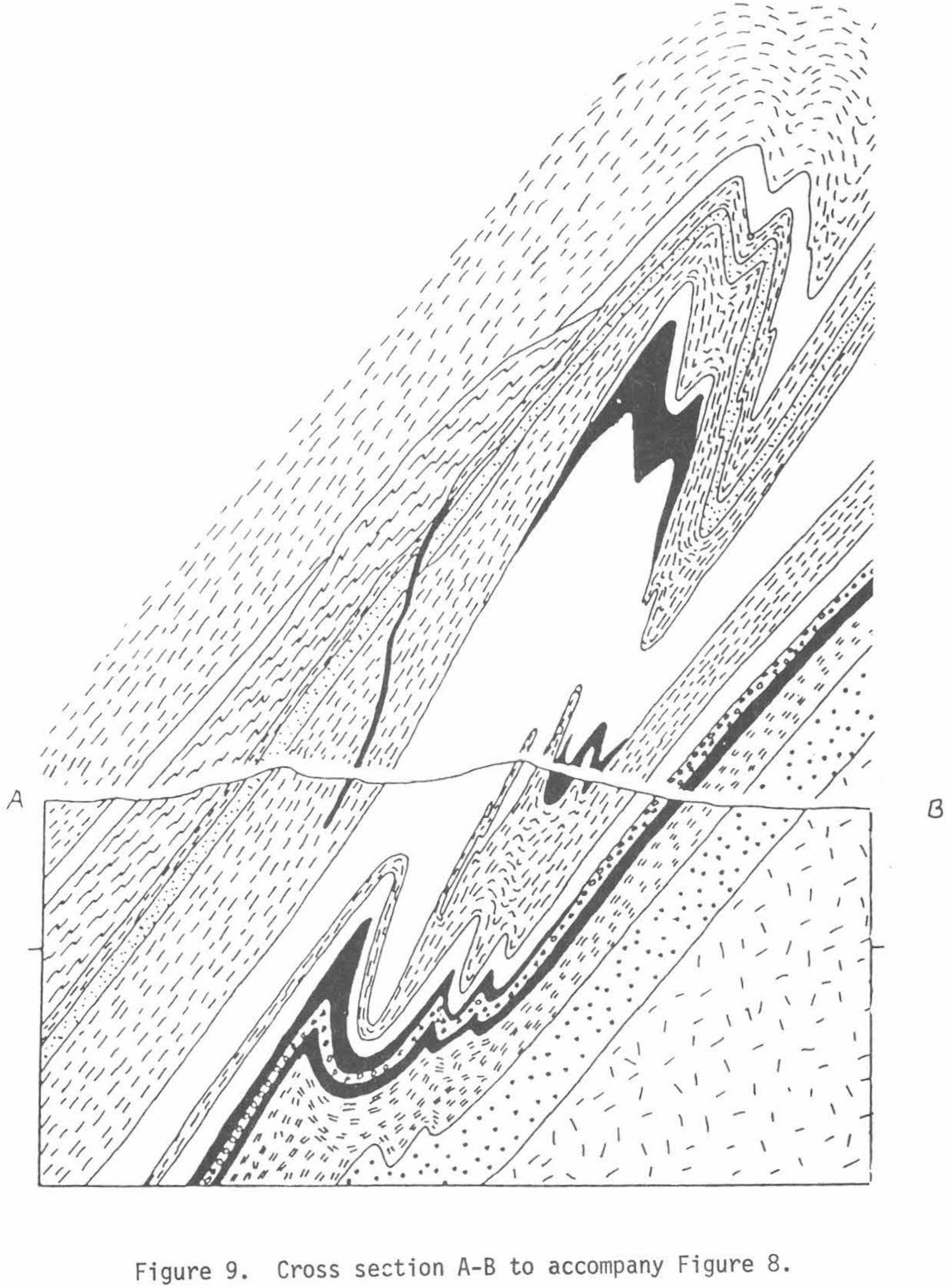

Figure 9. Cross section A-B to accompany Figure 8.

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Road Log

Distance given in miles from last stop in Potter's log.

Miles

0.0 Proceed south on Route 30.

2.7 Buck Mountain Fold Complex

This stop offers the opportunity to see the diopside augen quartz gneiss and other quartz rich gneisses and well exposed examples of second and third phase folds.

Road Cuts: Glassy quartzites and diopside augen quartz gneiss with small sparse diopside augen are exposed on the western side of Route 30. The small outcrop on the eastern side of the highway contains a few large dark green diopside porphyroblasts.

1-A.) Buck Mountain Syncline. Walk east from the highway 600 feet down the logging road. Bear left at a small cabin onto an overgrown skidder trail. Walk 1000 feet approximately N45E along a contour. The outcrop on the steep slope to the southeast exposes the hinge of the Buck Mountain Syncline, a large third phase fold (see figs. 8 and 9). This is a tight fold which plunges moderately to the southeast on a northwest trending, steeply south dipping axial surface.

1-B.) Proceed west from the highway on an overgrown trail from the south end of the western road cut for about 1000 feet. Turn north and climb the hill to a large pavement outcrop.

The rock is typical diopside augen quartz gneiss. The augen which have weathered out, giving the rock this distinctive texture, are large single crystals or aggregates of smaller grains of green to white diopside. Scapolite is common. Minor K-feldspar, and traces of calcite, sphene, and zoisite are present.

A second phase isoclinal fold, outlined by a thin amphibolite, is refolded by a tight third phase fold, forming a hook interference pattern. Note the lack of an axial plane foliation in the hinge of the second phase fold. A more prominent axial plane feature is present in the hinges of third phase folds. Other folds, and a nice view to the south, are visible higher on the hill.

Return to highway and proceed south.

7.0 Turn left at Cheerio's Cabins.

10.0 Bear right at the three-way intersection.

10.8 Stop 2. Rock Pond Syncline

The low outcrop to the left is a buff weathering diopside granulite which contains, in addition to diopside, calcite, plagioclase, and minor microperthite and quartz. Sulfides are present and graphite is abundant.

The structure of this area is poorly understood. The Rock Pond Syncline closes around the calcsilicate gneisses just to the northwest. Refolding in the third phase produced north-side up folds (see fig. 4, Section E-F). The inappropriate rotation sense of the third phase fold is attributed to original curvature of the hinge line of the Rock Pond Syncline. Late folding about northwest trending axial surfaces causes local changes in the direction of plunge of the third phase fold.

End of road log.

Thank you for comments and criticisms.

A TRAVERSE ACROSS THE SOUTHERN CONTACT OF THE MARCY ANORTHOSITE MASSIF, SANTANONI QUADRANGLE, NEW YORK

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INTRODUCTION

The 15' Santanoni quadrangle is located between 44°00' and 44°15' north latitude and 74°00' and 74°15' west longitude in what is known as the High Peaks region of the Adirondack Mountains of New York State (Figure 1). The nearest towns of any size are Lake Placid, which is northeast of the quadrangle, and Saranac Lake, which is north of the quadrangle (Figure 2).

Approximately 80% of the quadrangle is underlain by anorthosite or anorthosite-related rocks. The remainder of the area is underlain by pyroxene syenite gneisses (commonly ascribed to a mangerite-charnockite series), amphibolites, granitic, pelitic and calc-silicate gneisses, marbles and quartzites. All of these rocks have been affected by the approximately one billion year old Grenville metamorphism and form the southeastern part of the Grenville province of the Canadian shield. Bedrock geologic mapping of the Santanoni quadrangle at a scale of 1:62500 was carried out between 1978 and 1982 under the direction of Howard Jaffe from the University of Massachusetts.

Metamorphism

A number of workers have attempted to estimate metamorphic conditions for the area surrounding the Santanoni quadrangle. The quadrangle lies entirely within the 750° C isotherm of Bohlen et al. (1980). This isotherm is based on both magnetite-ilmenite and feldspar thermometry. Valley (1980) suggested maximum temperatures of just under 700° C for the area immediately south of the Santanoni quadrangle (northern part of Newcomb quadrangle) on the basis of the assemblage tremolite-calcite-quartz. Jaffe et al. (1978) estimated metamorphic temperatures between 760 and 790° C in the Mt. Marcy quadrangle on the basis of coexisting orthoferrosilite-ferroaugite compositions, and Valley and Essene estimated temperatures to be 750° C ± 30 ° based on akermanite bearing assemblages at Cascade Slide (Mt. Marcy quadrangle).

Pressure estimates based on the stability or iron-rich orthopyroxene suggest metamorphic pressures of approximately 8 kbar (Jaffe et al. 1978, Bohlen and Boettcher 1981). Reinterpretation of textures of iron-rich pyroxenes of Ollila et al. (1984) suggests that these may be igneous as well as metamorphic pressures. Valley and Essene (1980) estimated metamorphic pressures of 7.4 ± 1 kbar based on akermanite bearing assemblages at Cascade Slide in the Mt. Marcy quadrangle.

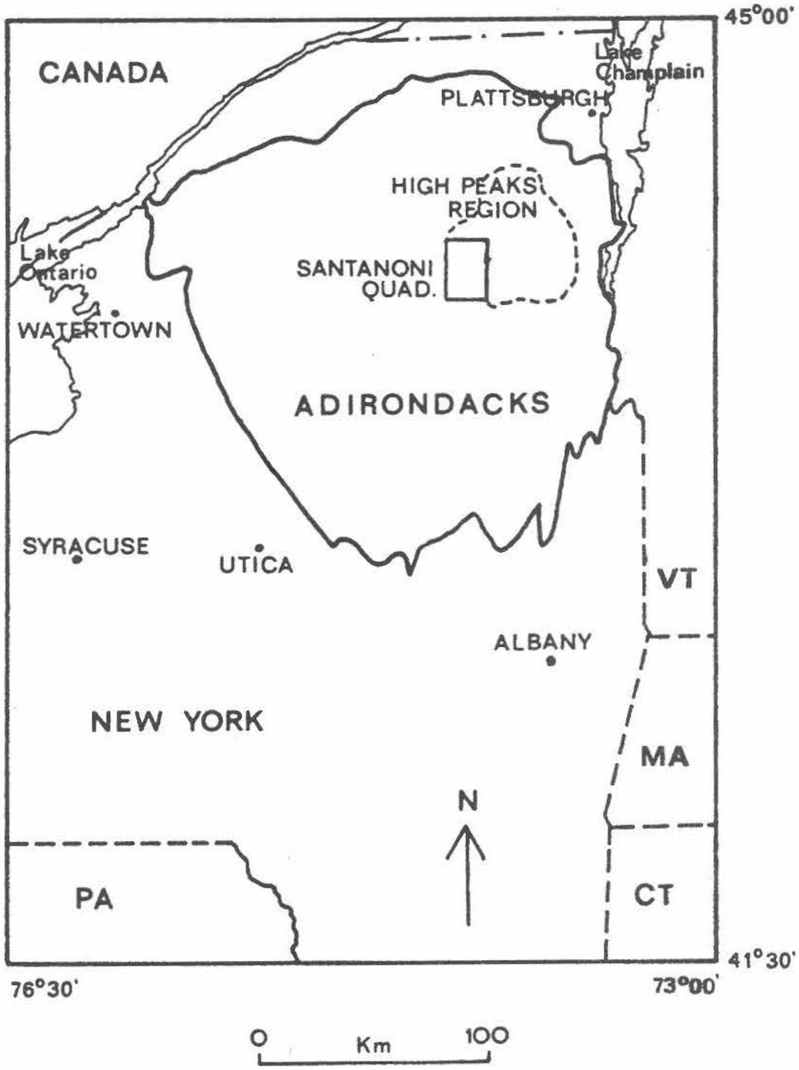


Figure 1. Index map showing location of the Santanoni quadrangle (modified after Buddington, 1953).

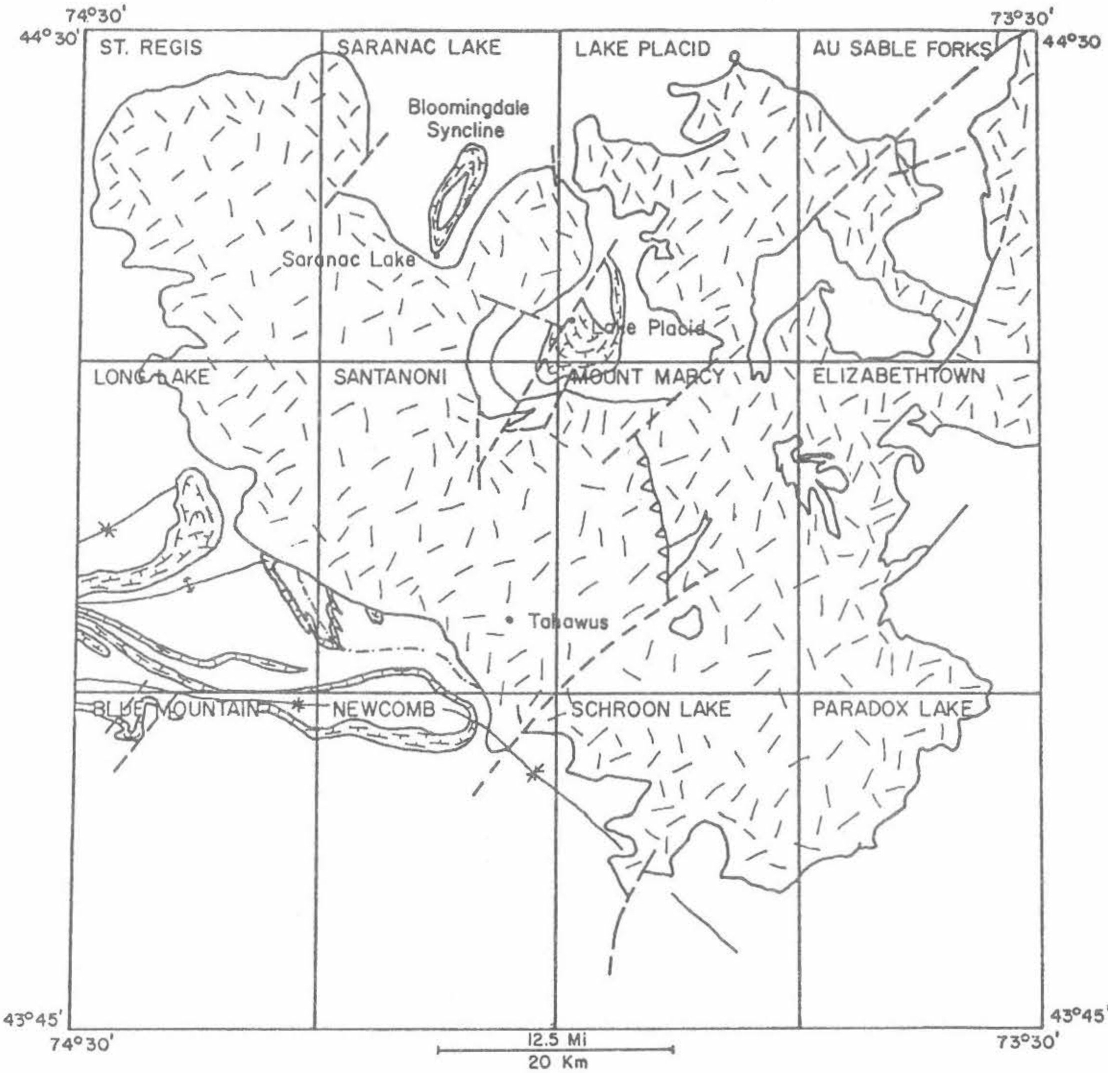


Figure 2. Generalized geologic map showing anorthosite, patterned, calc-silicate rocks and other gneisses . Axial traces of third phase synclines from Wiener et al., 1984, a third phase anticline and the second phase Wolf Pond syncline are also shown.

Use of the Newton and Perkins (1982) garnet-plagioclase-orthopyroxene quartz geobarometer on mineral compositions from the Mt. Marcy and Elizabethtown quadrangles published by Kretz (1981) gives pressures of approximately 9 kbar. Although these pressures are slightly higher than those given by Valley and Essene or Bohlen and Boettcher, the uncertainties are such that they are not significantly different.

Metamorphic assemblages in the Santanoni quadrangle are consistent with a regional metamorphic maximum occurring near or below 750° C. This conclusion is based upon the common occurrence of biotite + quartz and the single occurrence of tremolite + calcite + quartz in the southern part of the quadrangle. These assemblages place upper limits on metamorphic temperatures whereas the occurrence of orthopyroxene, both in granitic gneisses and amphibolites and the occurrence of sillimanite + potassium feldspar place minimum limits on the temperature of metamorphism. The temperatures at which biotite + quartz or muscovite + quartz react to produce orthopyroxene + potassium feldspar or sillimanite + potassium feldspar respectively are highly dependant on the composition of the fluid phase but it is perhaps reasonable to assume that the temperatures (650° C to 750° C) determined for muscovite-absent rocks in central Massachusetts by Robinson et al. (1982) are reasonable minimum estimates for metamorphic conditions in the Santanoni quadrangle. A single specimen from the southwestern part of the quadrangle contains both wollastonite and calcite + quartz. This is consistent with a high-grade regional metamorphism but suggests variable fluid compositions over short distances.

High-grade metamorphism of anorthositic rocks has produced garnet, hornblende and scapolite. These minerals are most common in highly deformed anorthositic rocks and all of these minerals could have formed during the same high-grade regional metamorphism that has affected all Adirondack rocks. The possibility, however, that garnet formed during isobaric cooling of anorthosite cannot be ruled out. Locally retrograde assemblages characterized by minerals such as prehnite, pumpellyite and chlorite can be found in anorthositic rocks.

Geochronology

Ashwal and Wooden (1983) have reviewed the various age determinations for Adirondack anorthosite and other rock types and provide evidence that Adirondack anorthosite may be as old as 1288 ± 36 Ma. This age is based on Sm-Nd data from mineral separates and whole rock samples from a layered sequence of rocks near Tahawus in the Santanoni quadrangle. Metamorphic ages based on whole rock and mineral separate data using both Sm-Nd and Rb-Sr range between 995 and 915 Ma. Ashwal and Wooden conclude that the most likely explanation for these ages is that anorthosite intrusion and crystallization was followed by distinctly later prograde metamorphism.

Basu and Pettingill (1983) concluded that anorthosite crystallization in the Snowy Mountain dome in the south-central Adirondacks took place at approximately 1100 Ma. This age is based on Sm-Nd data from whole rocks and mineral separates. The age is heavily dependent on garnet which is a metamorphic mineral in anorthositic rocks.

Basu and Pettingill's age corresponds closely with ages determined by Silver (1969) on zircon separates from a variety of Adirondack rocks. According to Ashwal and Wooden these ages represent metamorphic rather than igneous crystallization ages. The degree to which Basu and Pettingill's age depends on garnet supports the argument of Ashwal and Wooden.

The deposition age of metamorphosed sedimentary rocks in the Adirondacks is uncertain, but Grant et al. (1981) determined an age of 1265 ± 25 Ma for leucogneisses in the northwest lowlands of the Adirondacks. These rocks are interpreted to be metamorphosed felsic volcanics, are the oldest rocks in the Northwest Lowlands stratigraphy and were interpreted to be basement by Wiener et al. (1984).

Structural Geology

The Marcy anorthosite massif is roughly heart shaped and covers approximately 5000 km^2 (Figure 2). There are several hypotheses concerning the general shape of the massif. Bowen (1917) suggested that the main body of the Adirondack anorthosite massif is a laccolith. Buddington (1939) believed that what he called the St. Regis-Marcy anorthosite unit is a northwest-southwest elongate domical structure. He did not believe that the floor of this unit was exposed and suggested that it everywhere dipped underneath overlying metamorphosed sedimentary rocks. A similar hypothesis was presented by Davis (1971), who stated that foliated anorthosite dips under overlying syenitic rocks and metamorphosed sedimentary rocks in the St. Regis quadrangle. He described the anorthosite as a layered sheet with the more mafic portions at the top and reported that a few hills show upward increase in mafic minerals. Folding complicates these relations so that leucocratic anorthosite forms topographic highs in anticlinal areas and gabbroic anorthosite is found in synclinal topographic lows.

The domal hypothesis has most recently been re-expressed by Whitney (1982, p. 68) who suggests that anorthosite rose as "relatively rigid domes through the supracrustal rocks sweeping them into pseudoconformity with the anorthosite contact and disrupting earlier structural trends." Whitney's hypothesis differs significantly from Buddington's, however, in that he suggests that the anorthosite domes were emplaced in the solid state.

Crosby (1966) presented a different interpretation for anorthosite in the Jay-Whiteface sheet in the Lake Placid and Ausable Forks quadrangles. He interpreted anorthosite here as being in northerly to northeasterly transported nappes that emplace anorthosite over metamorphosed sedimentary and syenitic rocks. Jaffe et al. (1983) have described a similar nappe or thrust in the Mt. Marcy quadrangle that emplaces undeformed anorthosite over syenitic and anorthositic gneisses.

Balk (1930, 1931) believed the Adirondack anorthosite massif to be a lenticular sheet 10 to 12 miles thick tilted to the northeast at 20° to 30°. Balk believed the roof of this sheet was exposed in the north and south and the floor in the east. Buddington (1939) disagreed with this. He suggested that the features Balk observed in the east were sills of anorthosite and that the floor of the main body was never exposed.

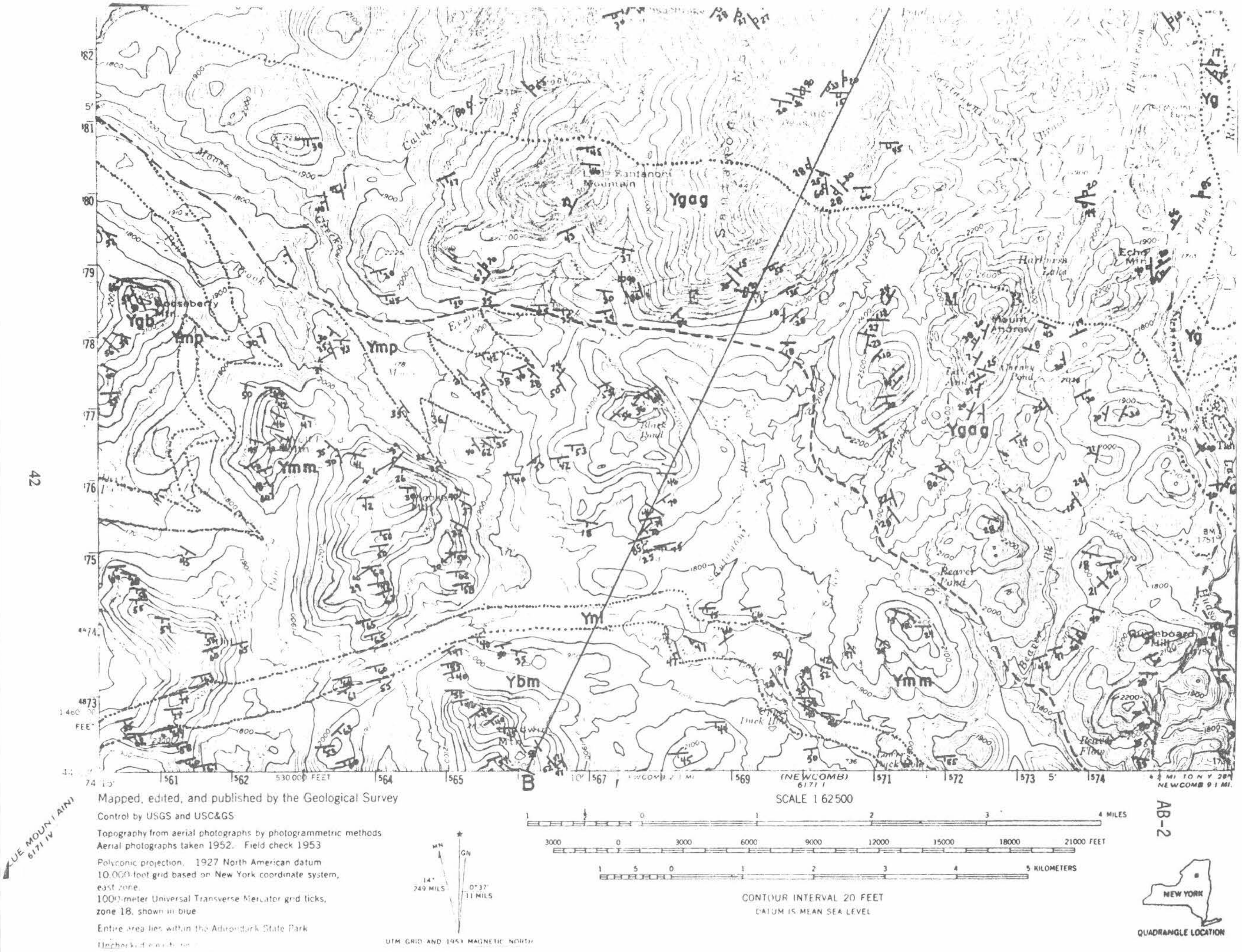
Simmons (1964) used gravity measurements to model possible size and shapes of the anorthosite massif. He concluded that the anorthosite is a slab 3 to 4.5 km thick with two roots extending downward to 10 km or more. Simmons' model is based on a density contrast of .010 between anorthosite and country rocks. The average density for country rocks of 2.82 implies a higher percentage of amphibolites in the country rocks than has been observed in the Santanoni quadrangle. The density contrast used by Simmons, however, is supported by the half width of the anomaly associated with the anorthosite body.

A more detailed gravity survey by Mann and Revetta (1979) detected five gravity lows within the anorthosite massif. They suggest that gravity highs between the lows may either represent thinning of the anorthosite or bodies of gabbro or amphibolite within the anorthosite.

As can be seen from this summary, there is no consensus on the shape, depth or extent of anorthosite in the Adirondacks. Most recent mapping, however, suggests that Buddington was correct in concluding that the floor of the main body of anorthosite is not exposed. This conclusion is supported by mapping in the Mt. Marcy quadrangle (Jaffe et al. 1983) and by relationships in the Santanoni quadrangle. Jaffe et al. (1983), Crosby (1966), and Buddington (1939) all have recognized anorthosite bodies that overly metamorphosed sedimentary rocks but interpret these as either thrust sheets, nappes, or sills.

One critical problem yet unsolved, is the exact nature of the contact of anorthosite with surrounding rocks. Davis found that the contact to be parallel to foliation in anorthosite and surrounding rocks. While this may be locally true, the foliation is secondary and need not be parallel to contacts. Detailed mapping of large areas combined with geophysical studies may be needed to delineate the shape of the anorthosite massif.

Within the Santanoni quadrangle, foliated anorthosite (gabbroic anorthosite gneiss) dips under other gneisses in the southwestern part of the quadrangle. In the northeastern part of the quadrangle, gabbroic anorthosite gneiss is limited in extent and has only been observed as sills or layers separated from the main anorthosite body. The main anorthosite body in the northeast corner of the quadrangle retains igneous textures up to its contact with pyroxene monzonite granulites which are intrusive into anorthosite. Foliated rocks away from the contact do, however, dip away from the anorthosite and imply


Ygag
Ymp
Ygb
Ymm
Yml
Ybm
Ymm
Yg
Little Santanoni Mountain
Gooseberry Mtn
Black Pond
Mount Andrew
Harkness Lake
Echo Mtn
Beaver Pond
Beaver Flow
Duck Hole
B
74°15'
530 000 FEET
(NEWCOMB)
6171 I
SCALE 1:62 500
4 MILES
21000 FEET
5 KILOMETERS
CONTOUR INTERVAL 20 FEET
DATUM IS MEAN SEA LEVEL
MN
GN
14°
249 MILS
0°37'
11 MILS
UTM GRID AND 1951 MAGNETIC NORTH
NEW YORK
QUADRANGLE LOCATION
NEWCOMB 9.1 MI.
Mapped, edited, and published by the Geological Survey
Control by USGS and USC&GS
Topography from aerial photographs by photogrammetric methods
Aerial photographs taken 1952. Field check 1953
Polyconic projection. 1927 North American datum
10,000-foot grid based on New York coordinate system, east zone
1000-meter Universal Transverse Mercator grid ticks, zone 18, shown in blue
Entire area lies within the Adirondack State Park
(BLUE MOUNTAIN)
6171 IV

AB-2

Figure 3. Geological map of the southwestern part of the Santononi quadrangle.

EXPLANATION

Intrusive Igneous Rocks

Ypqs Pyroxene quartz syenite gneiss; tan-weathering, brown on a broken surface, hornblende-pyroxene-quartz-microperthite gneiss.

Ypmg Pyroxene monzonite granulite; tan-to pink-weathering, gray to brown on a broken surface, fine grained granulite. Commonly contains bluish gray plagioclase megacrysts.

Ygag Gabbroic anorthosite gneiss; Hornblende-garnet-pyroxene-plagioclase augen gneiss, includes minor amounts of anorthosite, gabbro, pyroxene quartz syenite and monzonite gneiss.

Yg Gabbro; medium to coarse grained sub-ophitic gabbro and fine grained rusty-brown-weathering gabbro granulite. C.I. > 35

Yga Gabbroic-noritic anorthosite; medium to coarse grained sub-ophitic, C.I. between 10 and 35, includes minor amounts of anorthosite and gabbro.

Ya Anorthosite; medium to coarse-grained, C.I. < 10, includes minor amounts of gabbroic-noritic anorthosite and gabbro.

Metamorphosed Sedimentary and Interlayered Igneous Rocks

Ybm Baldwin Mountain Gneiss; gray-to pink granitic gneiss, amphibolite, thin calc-silicate rocks, and rusty-brown to gray-weathering sillimanite or biotite-garnet-sillimanite quartzite.

Ynl Newcomb Lake Formation; coarse grained marble, calc-silicate rocks, calcareous quartzite, minor interlayered granitic gneiss, amphibolite and rusty-brown granulites.

Ymm Moose Mountain Gneiss; gray-to brown-weathering hornblende granitic gneiss with thin (< 1m) amphibolite layers, hornblende (locally orthopyroxene) granite gneiss, minor interlayered pink granitic gneiss, calc-silicate rocks, biotite-quartz plagioclase gneiss and rusty-weathering plagioclase rich granulites.

Ymp Moose Pond Formation; coarse grained marble, calcareous quartzite, rusty-brown weathering diopside quartz-microperthite granulites, and fine grained biotite-quartz-microperthite gneiss.

Ygb Gooseberry Mountain Gneiss; pink-weathering medium-grained granitic gneiss containing 1 to 3 mm thick layers rich in sillimanite or sillimanite, biotite and quartz and thicker (1m) amphibolite layers.

Geological contact (dashed; approximate, dotted; inferred)

Fault (solid; approximate, dashed; inferred

Strike and dip of foliation

Strike and dip of plagioclase megacryst foliation

Strike and dip of compositional layering

Strike and dip of axial surface

Fold hinge line

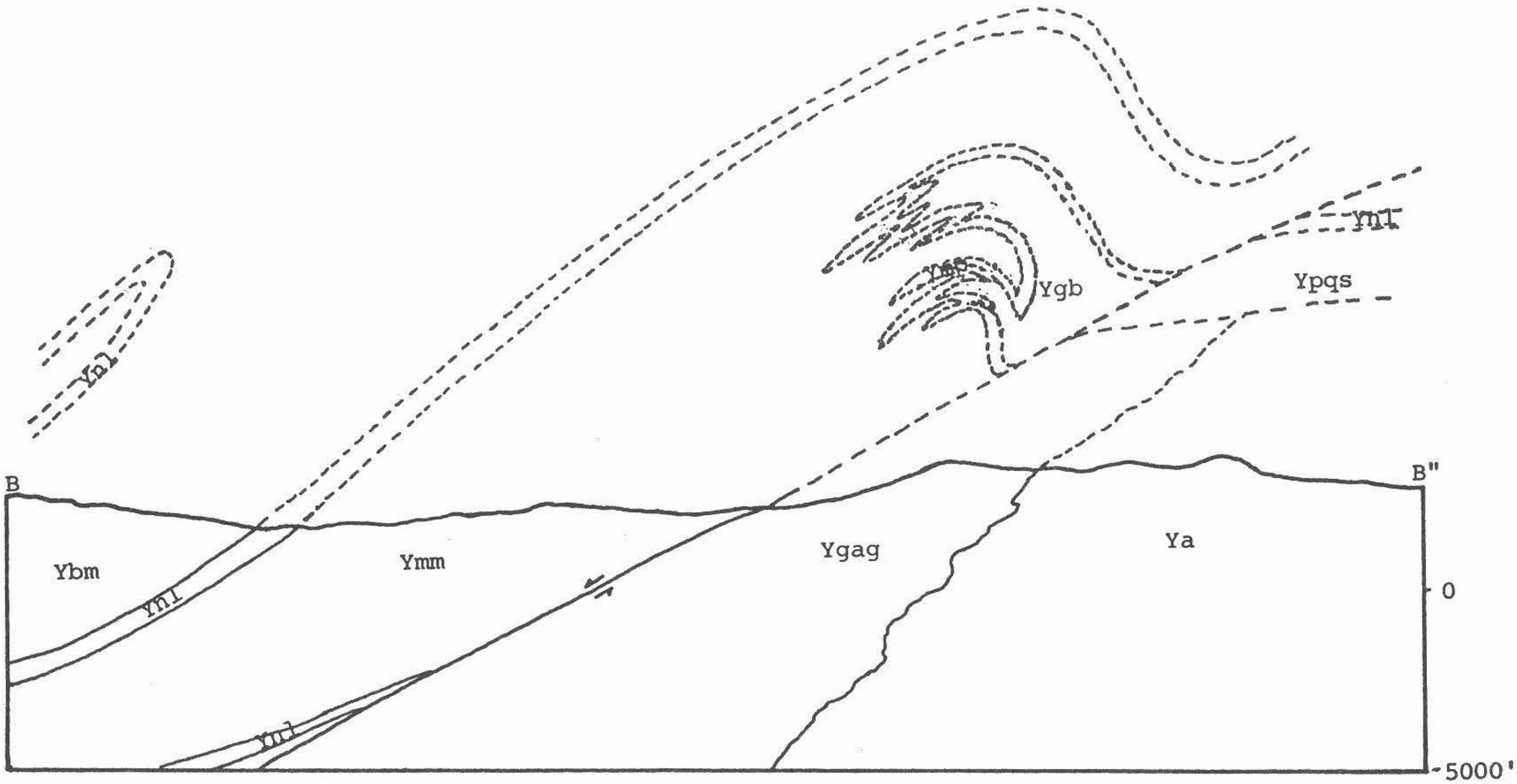


Figure 4. Schematic cross section along B-B'(Figure 3) illustrating the relationship between second and third phase folds and the Ermine Brook fault.

that the anorthosite dips beneath the surrounding rocks so that only the upper contact of the anorthosite is present in the Santanoni quadrangle.

Gneisses surrounding the major northwest trending lobe of the Marcy massif (Figure 2) define an open northwest trending doubly plunging anticline. A saddle in this anticline, caused by a northeast trending open syncline, occupies most of the Santanoni quadrangle. The part of this syncline that is located near the boundaries between the Santanoni, Mt. Marcy, Lake Placid and Saranac Lake quadrangles (Figure 2) was named the Newman syncline by Buddington (1953).

Northwest and northeast trending open folds refold three earlier phases of folding. Map scale folds related to two of these earlier phases of folding are evident in and around the southern part of the Santanoni quadrangle. The earliest map scale folds (F_2) are tight to isoclinal in nature and fold a foliation related to earlier isoclinal folds. Axial surfaces of second phase minor folds trend E-W to SE-NW and dip to the S. The Wolf Pond syncline (Figures 2 and 3) is a second phase fold. Other map scale second phase folds occur near Moose Pond.

Second phase folds are refolded by E-W trending folds (Figures 2, 3 and 4) whose axial surfaces dip to the south. Three large scale third phase folds are shown in Figure 2. These are from south to north the Newcomb syncline, Long Lake anticline and the eastern extension of the Loon Pond syncline of Potter (this guide book). The location of the axial traces of the Newcomb and Loon Pond synclines has been taken from the regional synthesis of Wiener et al. (1984). The location of the axial trace of the Long Lake anticline is uncertain. Contacts shown in eastern part of the Long Lake quadrangle in Figure 2 are highly speculative.

One of the principal reasons for mapping the Santanoni quadrangle was to try and constrain the timing of anorthosite intrusion relative to the regional deformation history. A sill of gabbroic anorthosite gneiss lies parallel to and slightly below the contact between the Moose Mountain Gneiss and the Newcomb Lake Formation across the entire southwestern part of the quadrangle. The foliation in the sill is concordant to that of surrounding rocks and this foliation and the sill are folded by third phase folds. Intrusion of the sill must have taken place prior to or during second phase folding. Jaffe et al. (1983) report xenoliths in anorthosite that contain second phase folds. This suggests that anorthosite intrusion was taking place during second phase folding. McLelland and Isachsen (1980) reached a similar conclusion on the basis of anorthosite dikes that are both involved in and cross cut second phase folds.

The gabbroic anorthosite gneiss contact in the southwestern part of the quadrangle cuts across both second and third phase folds (Figures 2 and 3) and at the one locality (Ermine Brook), where rock is exposed near the contact, rocks are intensely foliated and lineated and are finer grained than normal. This evidence and the map pattern has led to the conclusion that the contact is a fault.

The main purpose of this trip is to walk across this fault (named the Ermine Brook Fault). Along the way we will look at some of the gneisses that underlie most of the southwestern part of the quadrangle. These rocks have been subdivided into five units which are briefly described in the explanation for Figure 3.

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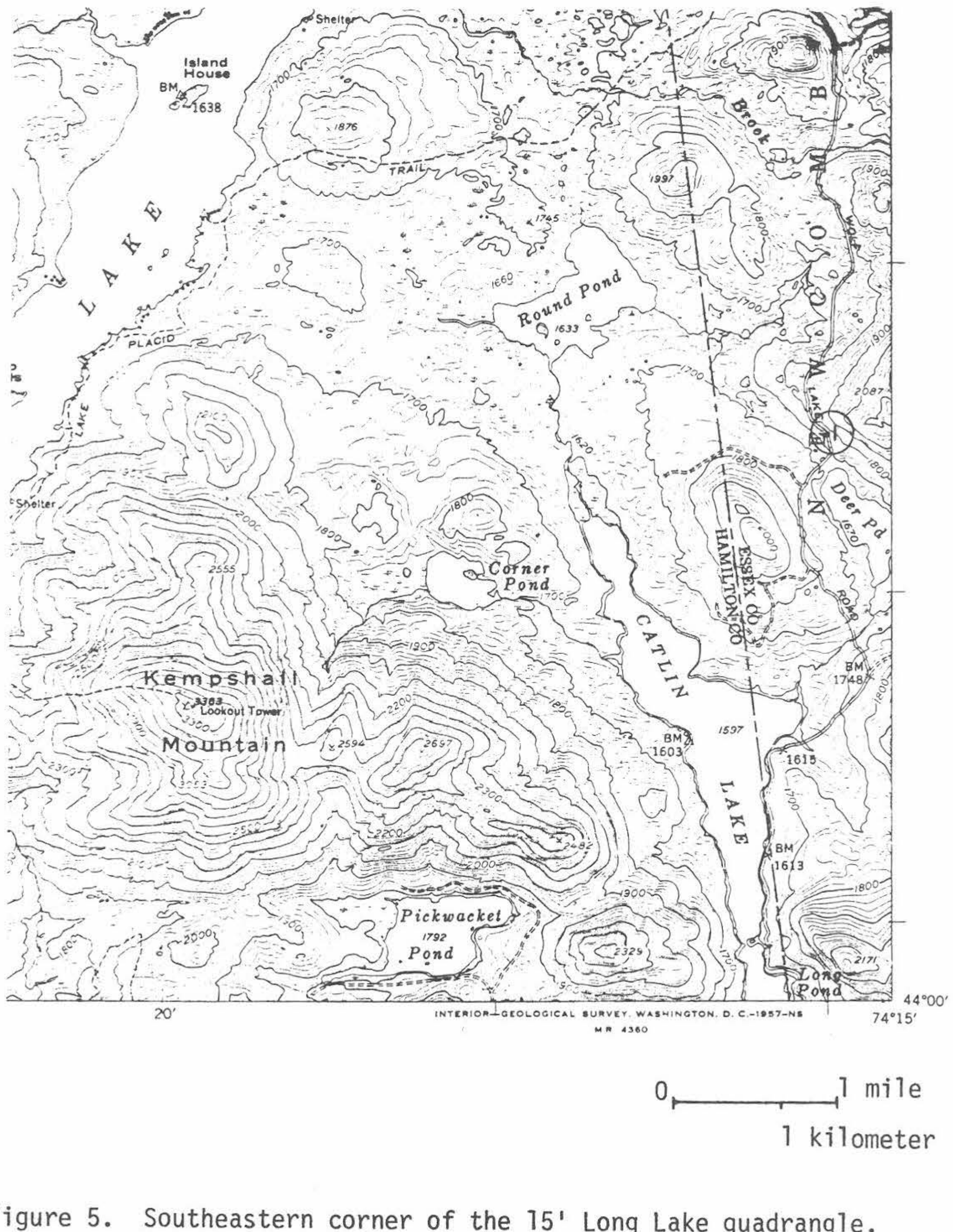


Figure 5. Southeastern corner of the 15' Long Lake quadrangle. Stop 1 is north of Deer Pond.

ROAD LOG

Mileage

0.0 Depart Long Lake public beach and proceed south on Route 30.

0.7 Turn left on Route 28N towards Newcomb.

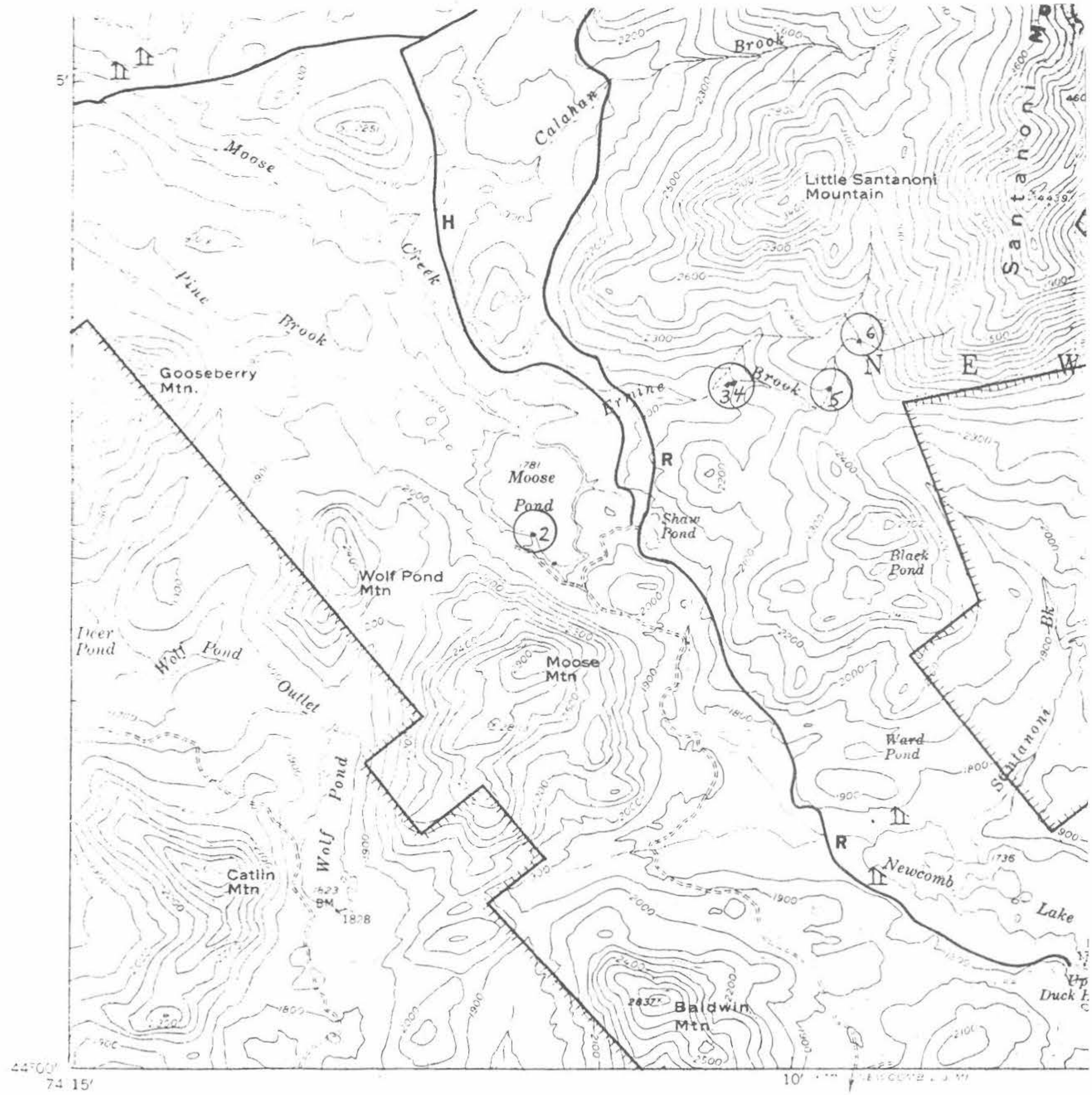
11.0 Proceed about 10 miles, turn left into Huntington Forest. From here on we will be traveling on private roads of the Huntington Forest, some of which are not on the topographic maps. We will walk to Moose Pond via the saddle between Moose Mountain and Wolf Pond Mountain. For those interested in visiting these localities at some other time, an excellent trail leaves from the Santanoni Preserve in the village of Newcomb. It is about a 7-mile walk to Moose Pond and 8 miles to Ermine Brook. Stop locations are shown on the 15' Long Lake (Figure 5) and Santanoni (Figure 6) quadrangle maps.

STOP 1 Gooseberry Mountain Gneiss

The Gooseberry Mountain Gneiss consists of pink-weathering, medium-grained granitic gneiss, black-and white-weathering amphibolites that average 1 m in thickness and gray-weathering layers 1 to 3 mm thick rich in sillimanite, biotite and quartz. Sillimanite gneisses are the key to identifying the Gooseberry Mountain gneiss. Other rocks included within this unit are quite similar to rocks included in the Moose Mountain Gneiss.

Sillimanite gneisses form only a small part of the unit but have been found on the north side of Gooseberry Mountain, on the southern end of a hill just east of Deer Pond, and on the southwestern side of Gooseberry Mountain in the Long Lake quadrangle.

North of Gooseberry Mountain, zones characterized by abundant sillimanite rich layers occur within pink granite gneiss. Within these zones, sillimanite-rich layers are separated by 1 to 2 cm layers of quartz and pink microperthite augen. Granitic gneiss within the Gooseberry Mountain gneiss is typically composed of anhedral quartz, and microperthite, brown biotite, magnetite, and may contain green hornblende and orthopyroxene. Sample 17-1 (Table 1) is fairly representative except that the granitic gneisses commonly contain higher modal proportions of magnetite. Average grain size is 1 to 2 mm and foliation is defined by oriented biotite. Sillimanite rich layers and amphibolite layers parallel foliation. Both weathered and broken surfaces of granitic gneiss are pink to brown. Locally the granite gneiss is garnetiferous and garnetiferous granitic pegmatites are fairly abundant on the southwest side of Gooseberry Mountain.



0 1 mile

1 kilometer

Figure 6. Southwestern corner of the 15' Santanoni quadrangle. Stops 2-6 are at Moose Pond and along Ermine Brook.

Table 1. Mineral Assemblages and Estimated Modes for Rocks within the Gooseberry Mountain Gneiss.

Sample No.	17-1	LL-4	17-21
Quartz	39	X	
Microperthite	57	X	+
Plagioclase	3		X
Hornblende			X
Biotite	1	X	+
Orthopyroxene			+
Sillimanite		X	
Opaques	+		+
Apatite			+

X = Major constituent

+ = Minor constituent

Hand specimen descriptions:

17-1 Medium-grained pink granitic gneiss.
Layer in granitic gneiss rich in sillimanite biotite and quartz.

LL-4 Sillimanite + biotite + quartz rich layers in granitic gneiss.

17-21 Medium-grained, amphibolite gneiss.

STOP 1 (cont'd)

Amphibolite layers are abundant at low elevations on the northern side of Gooseberry Mountain and also north of Deer Pond in the Long Lake quadrangle. The amphibolites are medium-grained but range in texture from well-foliated to granular. They are black-and white-weathering and typically have a color index near 50. Hornblende and plagioclase are the dominant minerals but amphibolites also contain brown biotite, orthopyroxene and oxides. Both garnetiferous and non-garnetiferous varieties have been observed. At location 17-22, granitic gneiss and amphibolite layers average 1 m in thickness and granitic gneiss forms about 60% of the outcrop. Going south from this location, the amount of amphibolite decreases, until near the summit of Gooseberry Mountain, the rock is almost entirely granitic gneiss. This sequence is reversed down the south side of Gooseberry Mountain. This repetition, and the occurrence of sillimanite gneisses on the north and south side of Gooseberry Mountain, suggest that this unit has been doubled by folding.

STOP 2 Moose Pond Formation-Ymp

The Moose Pond formation is extremely poorly exposed. One probably loose block of coarse-grained calcite marble (sample 17-6, Table 2) is located at the south end of Moose Pond. A mixture of rusty-brown medium-to-fine-grained diopside-quartz-microperthite granulites (17-7, Table 2) and fine-grained biotite-quartz-microperthite gneisses cropsout along the western shore of Moose Pond at location 17-7. Although there is abundant quartzite and diopsidic quartzite float northwest of Shaw Pond a nearby outcrop northwest of Shaw Pond contains only a few thin diopside-rich layers. Most of the rock at this location is a medium-to fine-grained brown augite-biotite-quartz-plagioclase-microperthite gneiss.

No outcrops were observed in the region between Gooseberry Mountain and Wolf Pond Mountain. The Moose Pond marble is inferred to go through this region on the basis of topography and the geometry of folds in the Moose Mountain gneiss and the Gooseberry Mountain Gneiss. The aeromagnetic map (1978) shows a magnetic low east of Gooseberry Mountain which is consistent with the presence of marble in this valley. Mineral assemblages in rocks from the Moose Pond marble are listed in Table 2.

Ermine Brook

This is the only place in the quadrangle, yet found, where rock is exposed near the gabbroic anorthosite gneiss contact. We will walk upstream along Ermine Brook and see the sequence of rocks shown at Stops 3, 4, 5, and 6.

Table 2. Mineral Assemblages from the Moose Pond and Newcomb Lake formations.

	Moose Pond		Newcomb Lake						
Sample No.	17-7	17-6	21-28	21-18	22-53	21-40	22-56	22-59	22-54
Calcite		X				+	X	X	
Clinopyroxene	X	X	X	X	X	X	X	X	
Tremolite							X		
Wollastonite						X			
Biotite	X								X
Phlogopite		X					X		
Quartz	X	X	X	X		X			X
K-feldspar		X	X	X	X	X		X	
Plagioclase	X	X				X			
Scapolite		X						X	
Sphene	X	X	X	X	X	X		X	
Apatite	X	X		X		X			
Graphite		X	X				X		
Tourmaline							X		
γ Cpx.	1.723	ND	1.697	ND	1.718	ND	1.697	1.697	

X = Major constituent

\+ = Minor constituent

ND = Not determined

STOP 3 Calcereous quartzites of the Moose Pond Formation.

These rocks contain quartz, diopside, calcite, sphene, microcline and scapolite. The scapolite is partially altered to prehnite.

The dominant foliation in these rocks trends 276°, 26 s and a strong lineation defined by trains of diopside trends 213°, 20°.

The rotation sense of minor folds at this outcrop indicates south side down movement.

STOP 4 Mylonitic pyroxene quartz syenite gneiss.

This rock is more intensely foliated and finer grained than normal for pyroxene quartz syenite gneiss. Foliation trends 90°, 35 s and locally a strong lineation defined by quartz ribbons trends 222°, 25°. A few asymmetric augen and what are possibly S and C foliations (Berthe et al. 1979, Simpson and Schmid, 1984) suggest south side down movement.

The mineralogy of pyroxene quartz syenite gneiss along Ermine Brook is typical of pyroxene quartz syenite gneiss found elsewhere in the quadrangle. Modes of typical pyroxene quartz syenite gneisses are listed in Table 3.

STOP 5 Gabbroic Anorthosite Gneiss (Ygag)

The gabbroic anorthosite gneiss map unit trends NW-SE in a zone across the southwestern corner of the quadrangle (Figure 3). It lies between metamorphosed sedimentary and igneous rocks to the southwest and anorthosite to the northeast. The dominant rock type within the unit has a color index (C.I.) equivalent to gabbroic anorthosite but minor amounts of gneissic anorthosite, noritic anorthosite, gabbro, and syenite are included as well. Forty-four samples from the gabbroic anorthosite gneiss map unit were examined in index of refraction oils or in thin-section. Of these 27 have a C. I. equivalent to gabbroic anorthosite, 7 are monzonites, 5 are anorthosites, 2 are gabbros, 1 is noritic anorthosite and 1 is a quartz syenite. The syenitic rocks and the gabbros generally occur as concordant layers or sills. Most of the rock in the unit has a composition similar to gabbroic anorthosite except that it contains more garnet, hornblende, alkali feldspar and quartz. Modes of four samples of gabbroic anorthosite gneiss (Table 4) are representative of the amount of hornblende, garnet, K-feldspar, and quartz typically found in these rocks. Orthopyroxene, although present in all

Table 3. Modes of Pyroxene Quartz Syenite Gneisses.

Sample No.	3-2	4-13	4-25	Al-4
Quartz	6.5	6.8	5.4	15.9
Microperthite	49.1	53.1	53.7	61.3
Plagioclase[1]	24.0	18.6	19.1	14.8
Augite[2]	6.9	8.6	8.5	2.5
Orthopyroxene[3]	3.4	6.3	2.9	1.4
Hornblende	3.9	3.3	6.7	3.0
Garnet	3.3	0.9	0.6	0.3
Opaques	1.9	1.5	1.9	0.3
Apatite	1.1	0.6	0.8	0.4
Zircon	+	0.3	0.1	0.1
Biotite	--	--	0.1	--
1 An_{plag}	18	21.9	--	--
2 MG_{cpx}	14	23	13	--
3 MG_{opx}	9	16	13	11
C.I.	20	22	22	8

Table 4. Modes of Gabbroic Anorthosite Gneiss

Sample No.	4-38	17-28	17-13	18-30
Plagioclase[1]	69.4	67.5	64.3	76.8
Alkali-feldspar*	6.3	7.4	5.8	2.7
Quartz	1.2	1.3	0.7	0.4
Augite[2]	7.5	10.3	9.2	4.9
Orthopyroxene[3]	2.0**	3.9	2.1	8.0
Hornblende	4.6	2.7	2.5	1.7
Garnet	3.4	2.5	9.7	3.6
Opaques	5.2	4.2	3.8	1.4
Biotite	--	--	0.1	0.4
Apatite	0.5	0.2	1.9	0.1
1 An	36.9	40.8	36.9	42.8
2 MG	68	58	58	54
3 MG		47	44	48

* Both microcline and orthoclase microperthite.

** Extremely altered.

STOP 5 (cont'd)

four modes, is not commonly present in anorthosite gneiss. At this locality pyroxene is commonly rimmed by hornblende which locally defines a strong lineation that trends 225°, 24°.

Rock at this locality is heterogeneous, consisting of gabbroic anorthosite gneiss with lesser amounts of gabbro, anorthosite, and syenitic gneiss. This is a good area to look at both intrusive relationships between the different rock types and the intense deformation typical of gabbroic anorthosite gneiss. Note also the different response of the various rock types to deformation.

STOP 6 Anorthosite.

This anorthosite is typical of anorthosite in much of the interior of massif. This rock is probably part of a layer or large xenolith of anorthosite in the gabbroic anorthosite gneiss map unit. More gabbroic anorthosite gneiss crops out to the north along the other branch of Ermine Brook and on Little Santanoni Mountain.

Walk down Ermine Brook and return to cars. If there is time, typical outcrops of Moose Mountain gneiss are easily accessible along the way.

FORELAND BASIN SEDIMENTATION IN THE TRENTON GROUP OF CENTRAL NEW YORK

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Purpose

The purpose of this field trip is to re-examine several of the well-known exposures of the Trenton Group in the Mohawk Valley. This re-examination is necessary due to the recent reinterpretations of the depo-tectonic setting and sedimentology of these units. Specifically, the Trenton Group in central New York has been redefined as a foreland basin and a tectonically active shelf edge/trench slope environment (Cisne, et al., 1982). The sedimentology of the Trenton Group has been shown (Mehrtens, in review) to evolve from in situ carbonates to dominantly turbiditic in nature. This field trip will visit several localities where participants will be able to view: (1) variations in the thickness of units over short lateral distances; (2) rapid facies changes within individiual stratigraphic sequences; (3) syn-depositional slump fold horizons; (4) internal structures within limestone beds, identifiable as Bouma sequences. These features are all characteristic of sedimentation in the tectonically active foreland basin. The text for this field trip will review the stratigraphic model presented by Cisne and Rabe (1978) and Cisne, et al (1982). Evidence for the interpretation of the limestone and shale couplets of the Denley Limestone and Dolgeville Facies as bioclastic turbidites will also be reviewed. Previous interpretations of the Trenton Group depo-tectonic setting will be presented and compared to the foreland basin model. It is hoped that this field trip will demonstrate, if nothing else, the strong role that a model or paradigm can play in the interpertation of stratigraphic sequences.

Introduction

Recent publications by Rowley and Kidd (1981) and Cisne and Rabe (1978) and Cisne, et al., (1982) have revised our understanding of the tectonic setting in which Cambro-Ordovician sediments of the Taconic System have accumulated. These workers have reviewed past interpretations that include considering the sedimentary facies associated with the Taconic Orogen to have accumulated in mio- to eugeosynclinal settings (Kay, 1937) and those which recognize convergent margin features associated with continent-arc collision (Zen, 1961, 1967). The original continent-arc model

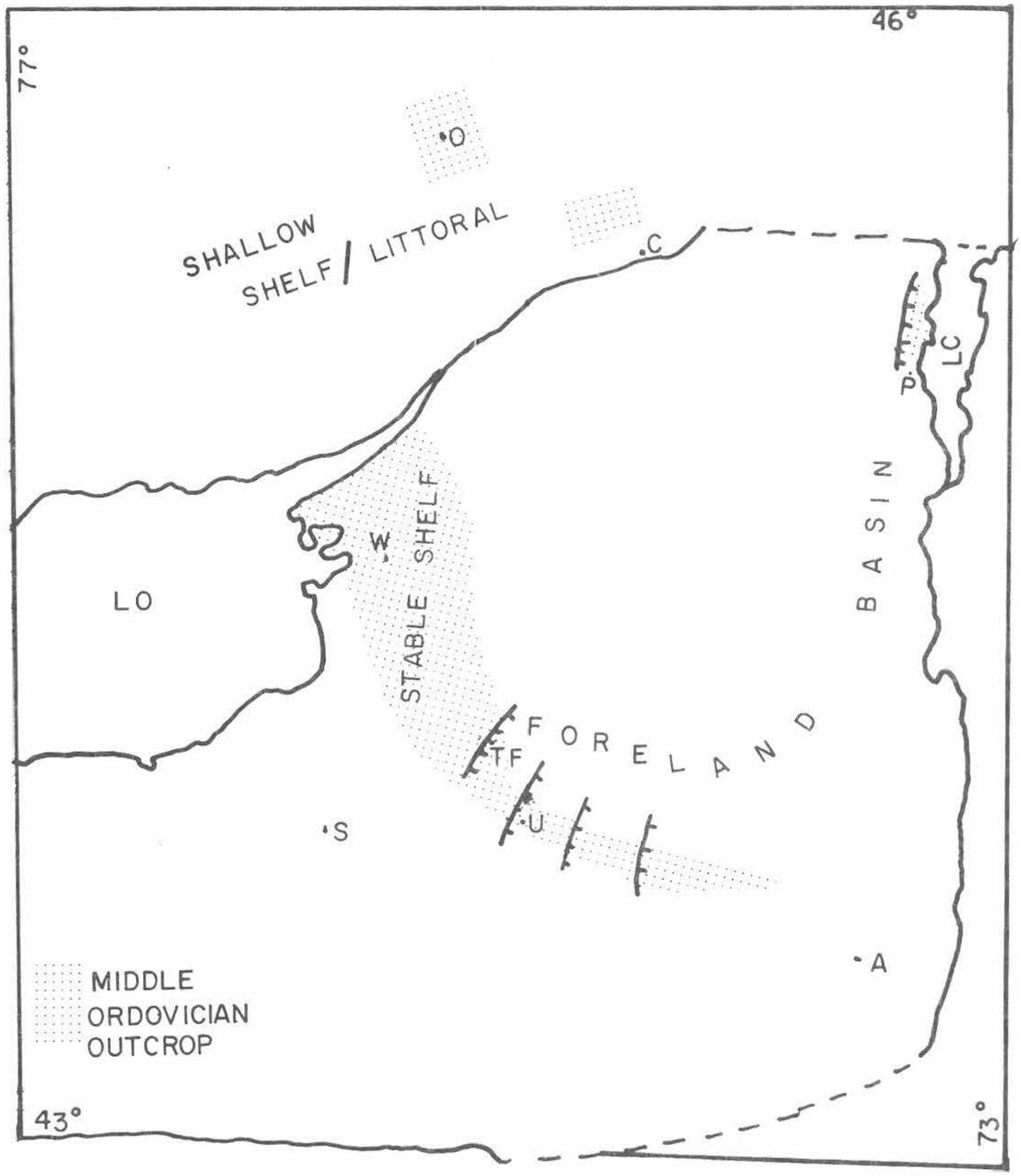

Figure 1. Paleogeographic map of the Trenton Group in New York and adjacent southern Ontario. Illustrated are the three main subdivisions of the depo-tectonic environment: the inner shelf or shallow shelf/littoral; the stable shelf; and the foreland basin. LO = Lake Ontario; LC = Lake Champlain; A = Albany; S = Syracuse; U = Utica; TF = Trenton type section at Trenton Falls Gorge; W = Watertown; C = Cornwall; O = Ottawa/Hull.

for the evolution of the Taconic Orogeny described by Bird and Dewey (1970) was refined by Rowley and Kidd on the basis of detailed petrographic and stratigraphic studies of the flysch associated with allochthon emplacement. Although the major contribution of their study is in the refinement of the timing and mechanism of allochthon emplacement, it also provides a concise summary of the tectonic settings of other portions of the continent-arc collisional belt.

Cisne, et al. (1982a) have also recognized that a continent-arc collisional model explains the stratigraphic sequence and facies succession of Middle Ordovician units in the Mohawk Valley, central New York. They compared this stratigraphic sequence and tectonic setting to that of the Australian flank of the Timor Trough, in other words, a continental shelf to outer trench slope setting. Their study, based on the detailed stratigraphy provided by bentonite horizons, generated paleobathymmetry and paleotopography of the cratonward portion of the shelf to trench transition.

The works of Cisne and Rabe (1978) and Cisne, et al. (1982) are important because they provide a major advance in our understanding of the sedimentary and biofacies of the Middle Ordovician sequence of the Mohawk Valley. In particular, it explains the origin and significance of: 1) numerous unconformities within the sequence; (2) attenuated and compressed stratigraphic sequences; (3) numerous syndepositional block faults; (4) major facies changes over short distances of the outcrop belt; (5) apparent anomolous juxtaposition of shallow and deep water sediments and sedimentary structures. These features have all been explained by recognition that the Middle Ordovician sediments were deposited on an actively fragmenting shelf in a foreland basin, and that each fault block within the shelf was accumulating a unique stratigraphic sequence. By using bentonite horizons to correlate fault blocks, Cisne and his coworkers constructed paleotopographic maps which illustrated the existence of an overall slope into the outer trench, superimposed on local topographic highs and lows. The extensional tectonics needed to generate this type of faulting in a compressional margin was described by Chapple (1973) as resulting from the passage of the shelf through the peripheral bulge. Similar syn-depositional extensional faults were described from the Middle Ordovician of southern Quebec by St. Julien and Hubert (1975) and their significance in producing localized stratigraphic sequences in the Middle Ordovician was mentioned by Mehrtens (1979).

The analogy of the Australian carbonate platform/Timor trench transect with that of the Trenton Group depotectonic setting is important. It suggests that there will be significant changes in paleoenvironments in the outcrop belt. Hypothesizing on what paleogeographic changes might be seen in the Trenton outcrop belt (Figure 1), the more cratonward exposures of the Trenton Group (west and northwest around the

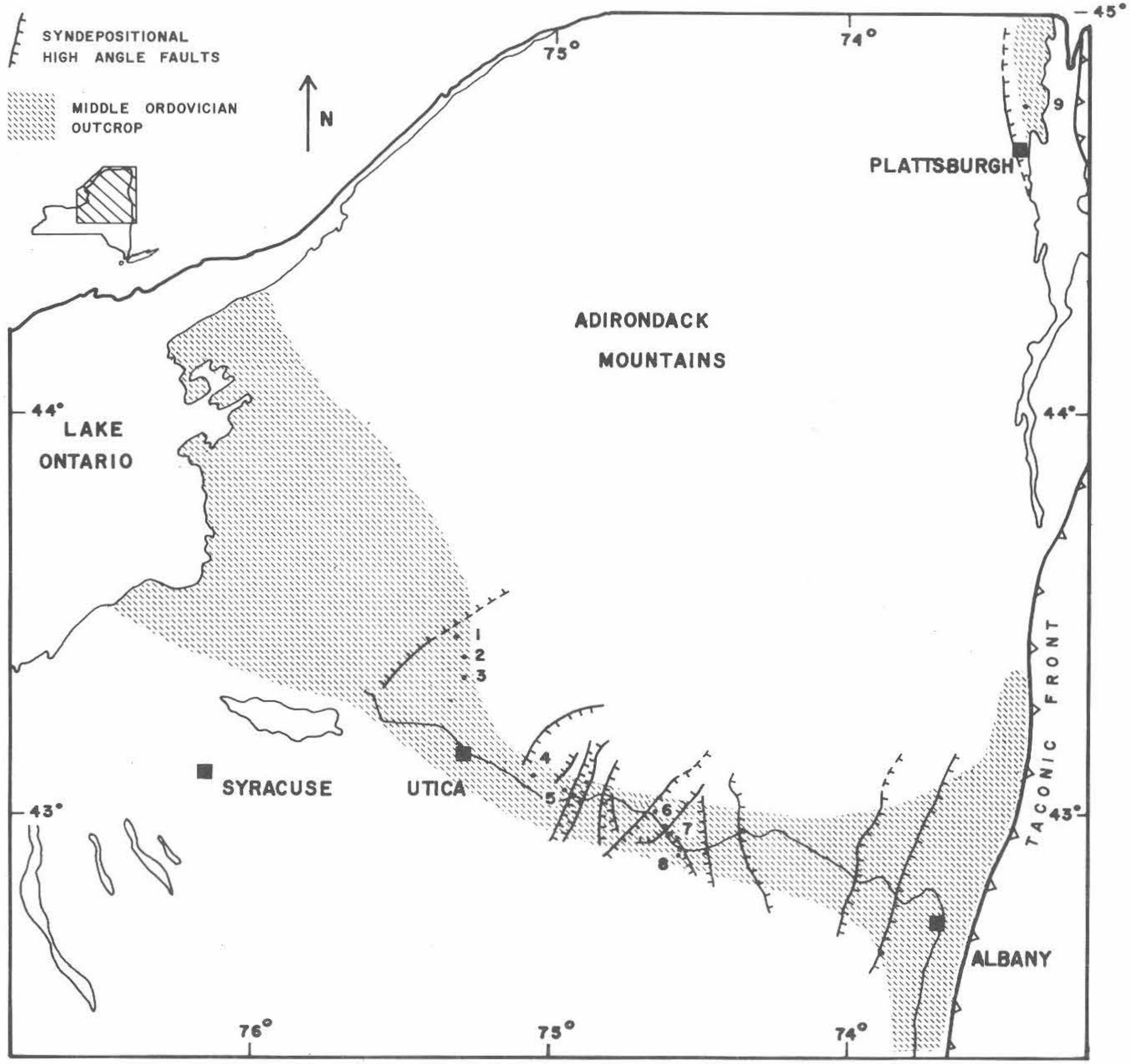

Figure 2. Locality map of northern New York State illustrating the outcrop belt of Trenton Group rocks. Key locations cited in the text include: 1 -Trenton Falls Gorge; 2 - Buttermilk Creek; 3 - City Brook; 4 - Herkimer; 5 - Little Falls; 6 - Caroga Creek; 7 - Palatine Bridge; 8 - Canajoharie Creek; 9 - Beekmantown.

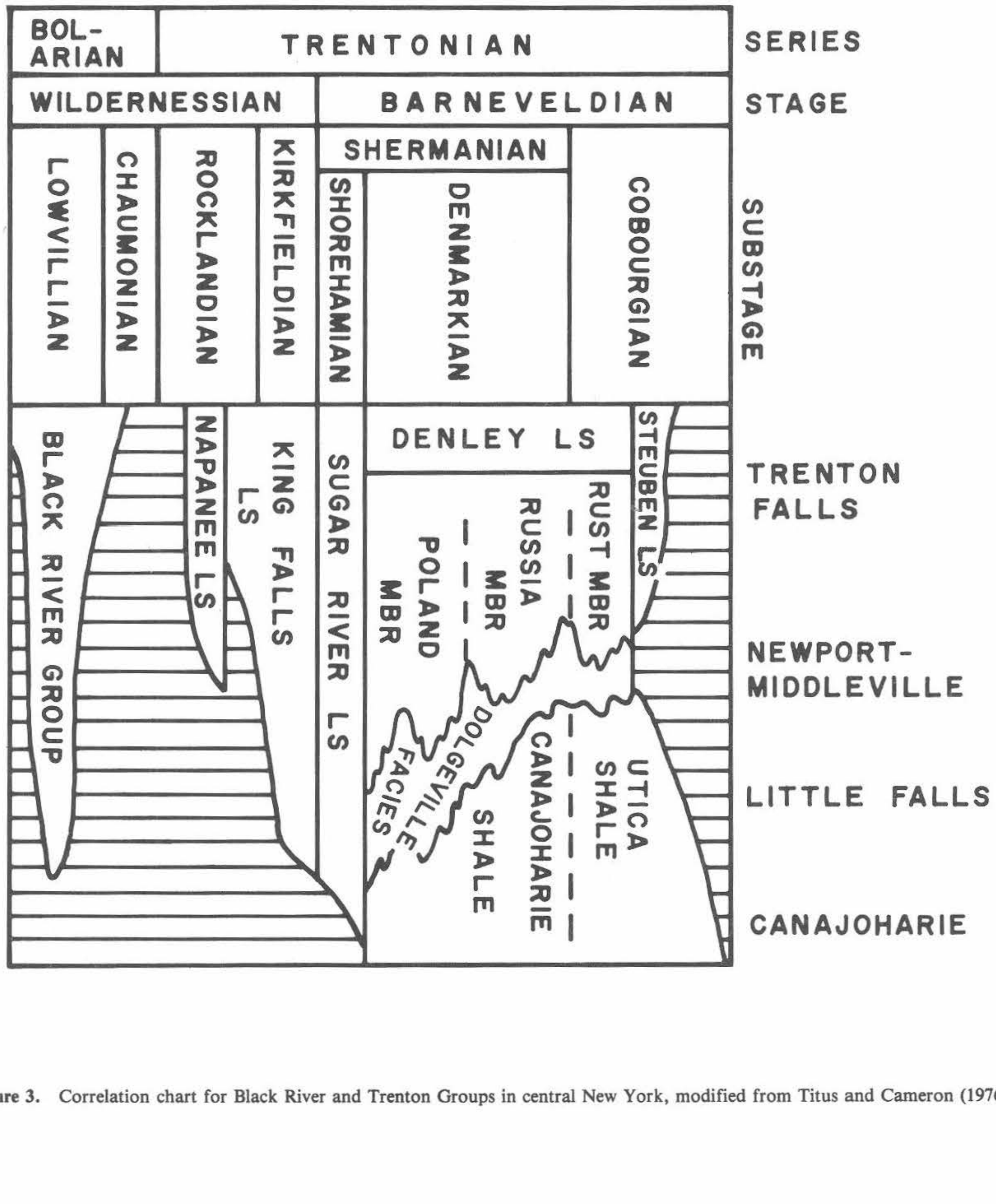

Figure 3. Correlation chart for Black River and Trenton Groups in central New York, modified from Titus and Cameron (1976).

Adirondacks) should be more removed from the trench setting, and therefore exhibit lithofacies that are shallow shelf in origin (ex. calcarenites and calcirudites exhibiting sedimentary structures diagnostic of wave-reworking) throughout the bulk of their stratigraphic sequence. Those portions of the outcrop belt which runs west-east in the Mohawk Valley are within the shelf to trench transition (Cisne and Rabe, 1978) and will exhibit features diagnostic of being below wave base and on a slope (slumps, turbidites, etc). A critical juncture then, would be the area of the outcrop belt where the transition from the stable, unfragmented shelf to that of the shelf/outer trench slope occurs. The most cratonward high angle fault in the outcrop belt occurs immediately to the northwest of the Trenton type section at Trenton Falls Gorge. This should also be the position of the stable shelf to foreland basin transition. It should be stressed that even within the foreland basin, shallow shelf (wave reworked) sediments will be deposited until the fault block founders. It is not unlikely then, to find shallow shelf sediments overlain by deeper water deposits. In this field trip we will be viewing outcrops that are all well within the fragmented foreland basin.

Stratigraphy

Pre-Trenton: The sedimentology of the Cambrian to Lower Ordovician siliciclastic and carbonate sequence (Little Falls Dolomite, Tribes Hill Dolomite, Beekmantown Dolomite) in the Mohawk Valley region of New York has been presented by Braun and Friedman (1969) and Mazzullo and Friedman (1975). These units represent peritidal to shallow shelf sediments and are, in the Mohawk Valley, unconformably overlain by the shallow shelf sandstones, dolostones, and limestones of the Black River Group. The nature of the Black River-Trenton contact is not diffinitive, and is thought by Cisne and Rabe (1978) to be a minor unconformity in central New York.

Trenton Group: The outcrop belt of Trenton Group rocks in central New York State extends through the Mohawk and Black River Valleys (Figure 2). This field trip will visit localities from northwest (Prospect, N.Y.) to southeast (Canajoharie Creek, N.Y.). The Trenton outcrop belt continues to the northwest and southeast direction (Figure 2), with isolated outcrops also present in the Champlain Valley of northern New York and western Vermont.

The stratigraphy of the Trenton Group was presented by Titus and Cameron (1976) who updated earlier versions by Kay (1937, 1953). The general stratigraphy of the Trenton Group is presented in Figure 3, which is modified from Titus and Cameron (1976).

The sedimentology of the basal units of the lower Trenton Group, the Selby and Napanee Limestones, have already been

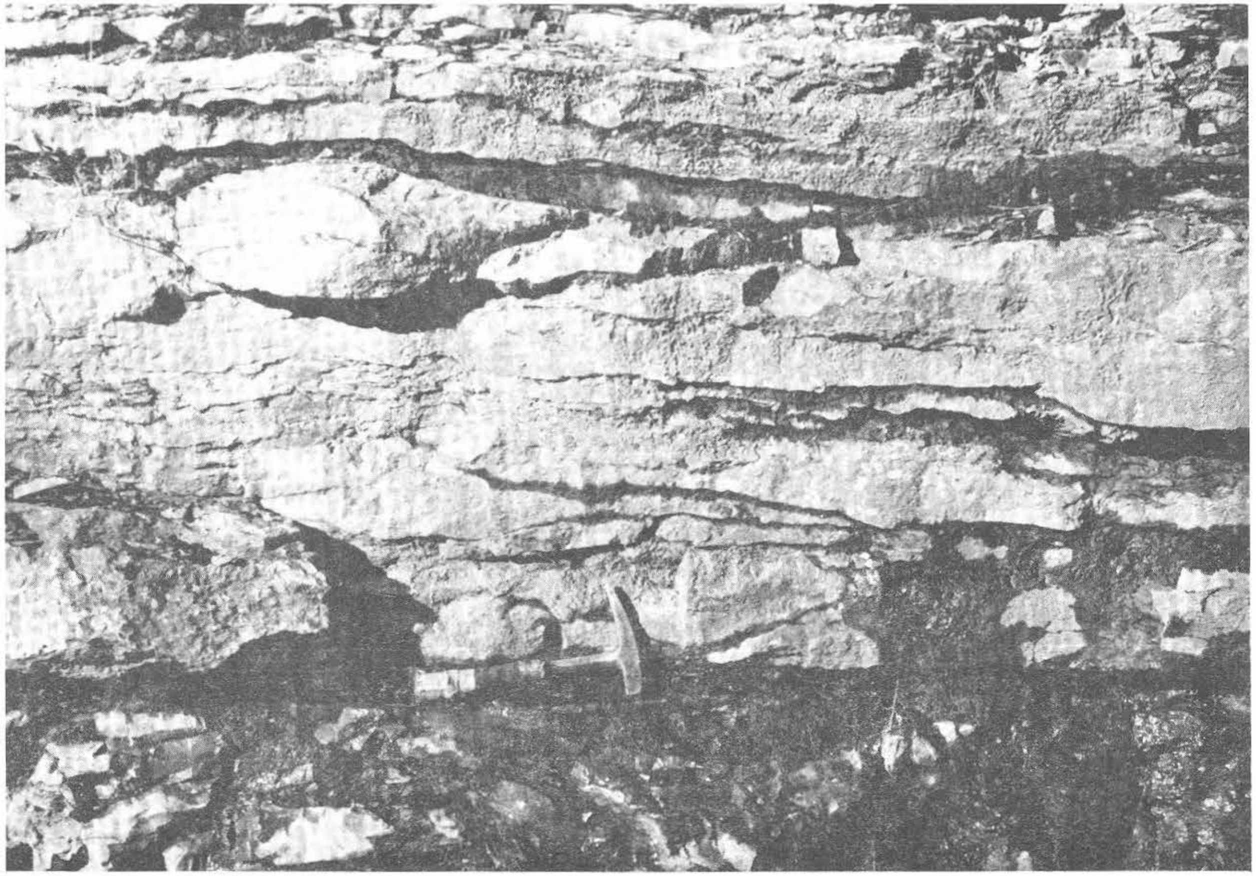

Figure 4. Outcrop photograph illustrating the typical bedding styles and textures of upper Kings Falls lithologies. Note the irregular bed thicknesses (lateral thickening and thinning) and wavy or hummocky bedding surfaces.

described in detail by Titus and Cameron (1976) and were the subject of previous NYSGA field trips (Cameron, Mangion and Titus, 1973 and Cameron, 1972). They are only briefly reviewed here. Essentially, they form a continuum of environments representing progressively increasing water depth from "lagoonal facies" (Napanee and Selby) to "offshore shoal facies" (lower Kings Falls), "shallow shelf facies" (middle and upper Kings Falls) and "deep offshore shelf" (lower and middle Sugar River) facies (Titus and Cameron, 1976, pp. 1211-1212). The lithofacies represented by these units include, for the Napanee and Selby, "relatively thick bedded calcilutites and calcsiltites with shale interbeds, horizontally and inclined current laminae and ripples, mudcracks, birdseye and vertical burrows". The shoal facies of the Kings Falls is a "thick bedded skeletal calcarenite" with "coquinal calcirudites, calcsiltites and interbedded calcareous shale" also present. The shallow shelf facies of the Kings Falls is a thinner bedded "sparry coquinal calcarenite" while the deep offshore shelf facies of the lower and middle Sugar River is a "thin bedded calcarenite with thin calcareous-shale interbeds" (Titus and Cameron, 1976, pp. 1211-1212). The typical bedding styles of the Kings Falls and lower Sugar River Limestones are shown in Figure 4. Note that the beds are laterally discontinuous, exhibiting lensing. This hummocky bedding style is a characteristic of wave reworked substrates, and has been noted by Aigner (1982) as diagnostic of storm reworked sediments. The Kings Falls and lower Sugar River lithologies will be seen at Shedd Brook (Stop 3) and Buttermilk Creek (Stop 1). The transition from this amalgamated bedding style to the more planar beds of the upper Sugar River and Denley, is significant and is interpreted as representing the transition from near wave base to below-wave base depths. Compare the relative thicknesses of the lithologies exhibiting amalgamated bedding among outcrops. Another characteristic of the Kings Falls and lower Sugar River Limestones is the relative absence of shale. This feature distinguishes the lower Sugar River from the upper Sugar as the contact with the Denley Limestone is approached.

The Denley Limestone overlies the Sugar River Limestone and is divisible at the type section in Trenton Falls Gorge (locality 1, Figure 1) into three members: Poland, Russia and Rust. The Denley/Sugar River contact is completely gradational in nature. On this field trip it will be seen at Stop 3 (Shedd Brook), Stop 1 (Buttermilk Creek) and Stop 2 (City Brook). The basal Denley differs from the Sugar River in that it: (1) contains a lower ratio of calcarenite to calcilutite (is finer-grained); (2) has slightly thicker shale interbeds; (3) contains horizons of barren, unfossiliferous, bioturbated micrite; (4) contains micrite beds with planar bases, bioturbated tops and a sequence of sedimentary structures recognizable as Bouma sequences in between. It was the recognition of these beds as turbidites that led to a re-evaluation of the depositional environment for the Denley Limestone. Also present within the Denley Limestone are

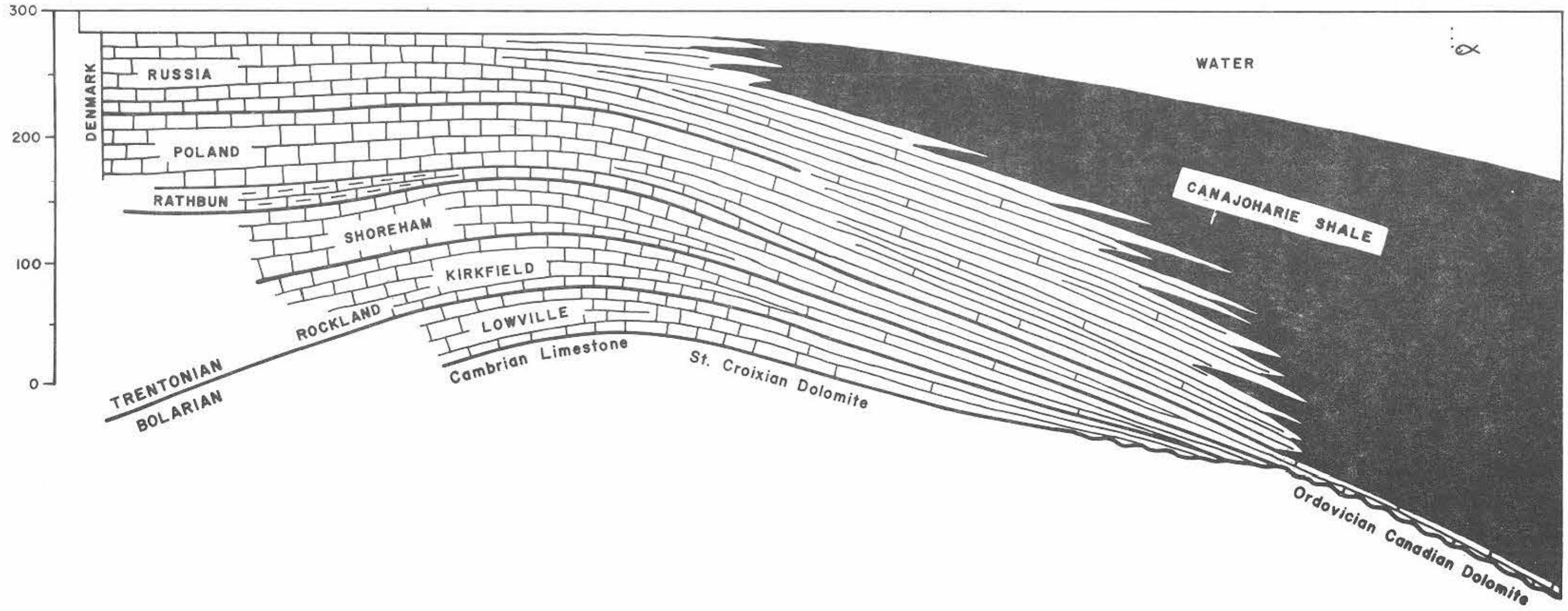

Figure 5. The restored section of the Trenton Group, after Kay (1965, Fig. 8.12). Essentially, Kay interpreted the Trenton units to be deposited on a shelf adjacent to a shale basin with a simple west to east transition from shallow shelf to transitional to deep water facies change. Compare this to Figure 6.

brecciated limestone horizons containing slump folds. The slump fold horizon visible in Figure 16 will be seen at Stop 4 (West Canada Creek).

Uppermost Trenton: Within the outcrop belt the Denley is overlain by a variety of units. East of the Trenton type section at Trenton Falls Gorge, the Dolgeville Facies was recognized by Kay (1953) as representing the transition between the Denley Limestone and the Canajoharie Shale. It is characterized by its fine-grained micritic composition, planar bases, laterally continuous beds and ubiquitous horizontal laminations (Figure 5). The Dolgeville will be seen at Stop 2 (City Brook). At the Trenton type section, the Steuben and Hillier Limestones overlie the Denley and have been described by Kay (1937, 1953) as a "thick bedded cross laminated calcite sandstone or calcarenite" and "argillaceous calcarenite and calcsiltite with interbedded shale", respectively (1953, p. 61). These limestones are conformably overlain by the Utica and Canajoharie Shales (Riva, 1972, fig. 16). The Steuben and Hillier Limestones are not present southeast of localities 1-3, Figure 2, and will not be seen in this field trip. In this region of the outcrop belt the shales of the Utica and Canajoharie overlie the Denley Limestone. The sedimentology of the Steuben and Hillier equivalents in the Trenton Group in Ontario have been studied by Brookfield (1983). Brookfield recognized lithofacies characteristic of supratidal, intertidal, lagoonal and shoal environments. Considering again Trenton paleogeography, the presence of these extremely shallow water environments in the most northwesterly portion of the outcrop belt is in keeping with the depositional model presented earlier.

To past workers it has always been somewhat problematical as to why the depositional environments represented by the lower to upper Trenton sediments should show a progressive deepening throughout most of the interval (Napanee and Selby Limestones), a shoaling (Steuben and Hillier Limestones) and ultimately, a rapid deepening (Utica and Canajoharie Shales). This oscillation was most recently attributed by Titus (1983) to be the result of facies mosaicing and oscillating eustatic sea level. As Cisne and Rabe's work demonstrates, extensional block faulting would explain why some of these facies (Steuben and Hillier, for example) are so localized in their occurence. The distribution pattern for Trenton units was incorporated into a model by Kay (1937, 1953), and is shown in Figure 5. As Figure 5 illustrates, the Trenton sediments were thought to have accumulated cratonward of the hinge of a subsiding shale basin. This model has persisted to the present day (Titus and Cameron, 1976). What differentiates Kay`s model from the one presented by Cisne and Rabe (1978) and adopted here, are the age relationships of the rock units from west to east. Kay`s (1937) use of the time-stratigraphic concept (rock unit=time unit) clouded the true chronostratigraphic relationships of the Trenton units. It is important to note that the bentonite stratigraphy presented by Cisne and Rabe demonstrates that the

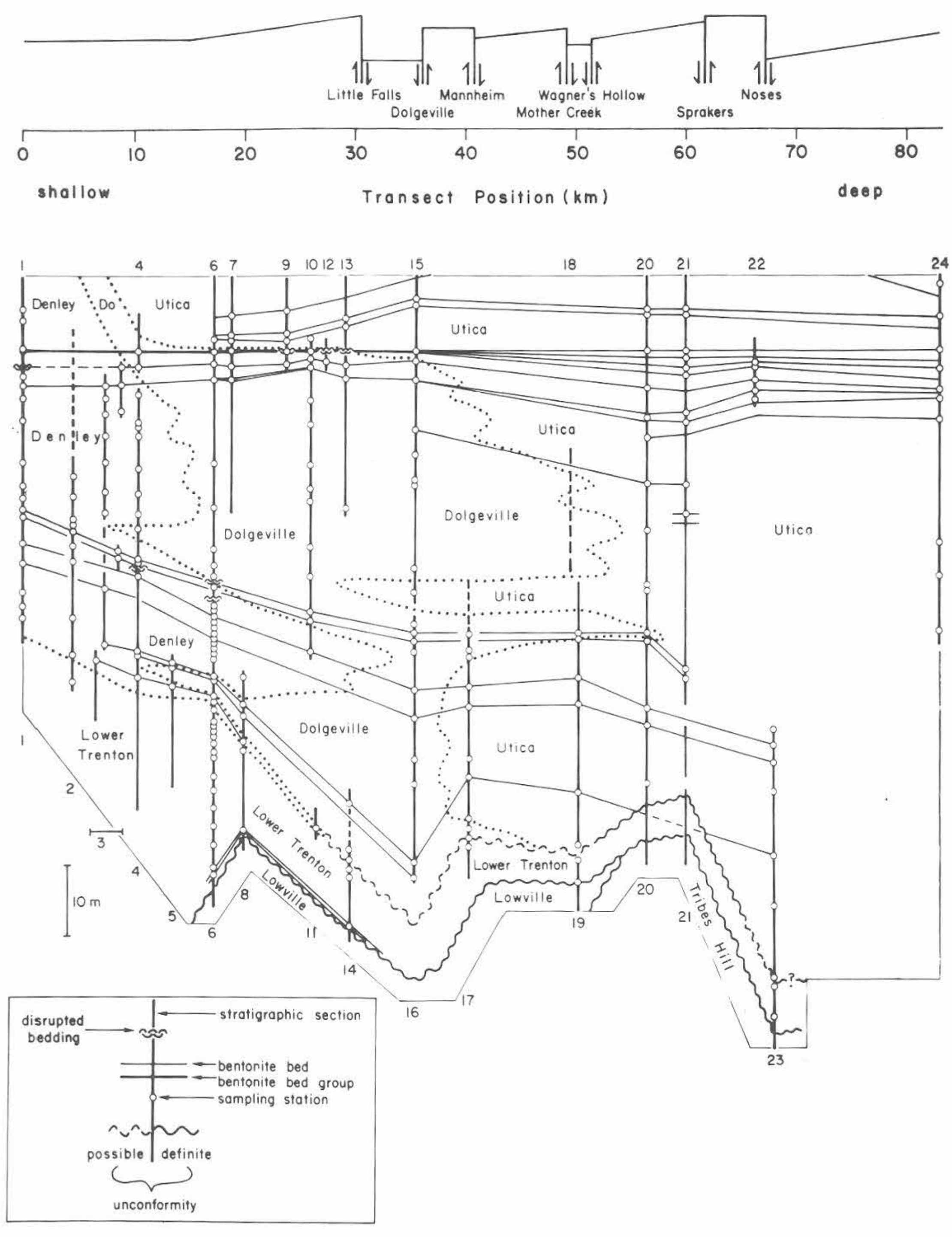

Figure 6. Stratigraphic sections of the Trenton Group and over- and underlying units are represented by vertical lines and have been correlated by bentonites (sub-horizontal lines). Sampling horizons for faunal studies are indicated by circles on the vertical lines. The structural cross section at the top of the diagram indicates that the Trenton strata thicken across basins and thin across basement highs, indicating that faulting is syndepositional. The sections are numbered: 1 - Trenton Falls; 2 - Gravesville-Mill Creek; 3 - Poland; 4 - Rathbun Brook; 5 - Shedd Brook; 6 - Buttermilk Creek; 7 - Farber Lane; 8 - Norway; 9 - Miller Rd.; 10 - North Creek; 11 - Gun Club Rd.; 12 & 13 - New York State Thruway; 14 - Burrell & Bronner Rds.; 15 - W. Crum Creek; 16 - Dolgeville Dam; 17 - E. Canada Creek; 18 - Mother Creek; 19 - Caroga Creek; 20 - Canada Creek; 21 - Flat Creek; 22 - Currytown Quarry; 23 - Van Wie Creek; 24 - Chuctanunda Creek. From Cisne and Rabe, 1978.

first appearance of a lithofacies in a stratigraphic sequence is not correlative with the first appearance in any other locality. Cameron (1972) also recognized the disachronous nature of formational contacts and attributed the rapid facies changes in an outcrop and the changes in thicknesses of units between outcrops to be the result of onlap resulting from transgression. These models differ primarily in the mechanisms invoked to explain the observed stratigraphic sequences. The foreland basin model presented by Cisne and Rabe (1978) and Cisne, et al (1982) suggests that foundering of the shelf was a more important factor than facies mosaicing associated with onlap. Their model is supported by the recognition of bioclastic turbidites from the Denley Limestone and Dolgeville Facies, suggestive of deeper water sedimentation characteristic of the outer shelf and slope environments.

The revised stratigraphy presented by Cisne and Rabe (1978) is based on the measurement of 24 stratigraphic sections in the Mohawk Valley. Examination of Figure 6 reveals several important features: (1) most of the 24 measured sections have a unique stratigraphic sequence; (2) that although there is a general eastward increase in water depth (as recognized by the Dolgeville and Utica Shale facies) there are exceptions to this patter (note Dolgeville tongues in sections 5-8 and Denley tongue in sections 8-14; (3) within any one stratigraphic section there is not a uniform increase in water depth; (4) note in Figure 6, the irregular topography of the underlying surface; (5) some stratigraphic sections are significantly thinned (example section 23 in Figure 6 where the upper Trenton is missing and the lower Trenton is immediately overlain by Utica). We will be visiting several of these exposures on this field trip. We will see: (1) section 5 on Figure 6, Shedd Brook (Stop 3), where the upper Kings Falls, entire Sugar River and Denley Limestones, and lower Dolgeville Facies are exposed; (2) section 6 on Figure 6 (Buttermilk and City Brooks, Stops 1 & 2), where the Black River/Lower Trenton contact, all of the Lower Trenton, Sugar River and Denley Limestones, and the base of the Dolgeville Facies are exposed; (3) section 1 on Figure 6, West Canada Creek (Stop 4), where the upper Denley Limestone exhbiting slump fold horizons can be seen; (4) section 16 on Figure 6, Dolgeville (Stop 5), where the best exposures of Dolgeville Facies can be seen; (5) section 20 on Figure 6, Canajoharie Creek (Stop 6), where the unconformity between the Lower Ordovician Tribes Hill Dolomite and the Lower Trenton is exposed, and the thinned Lower Trenton (Kings Falls) is immediately overlain by Utica Shale; (6) Caroga Creek (Stop 7), section 19 on Figure 6, another exposure of the Black River/Lower Trenton contact, thinned Lower Trenton lithofacies and overlying Utica Shale.

Turbidite Sedimentology

Research on the depostional environment of the Denley Limestone was begun as part of the data collection and

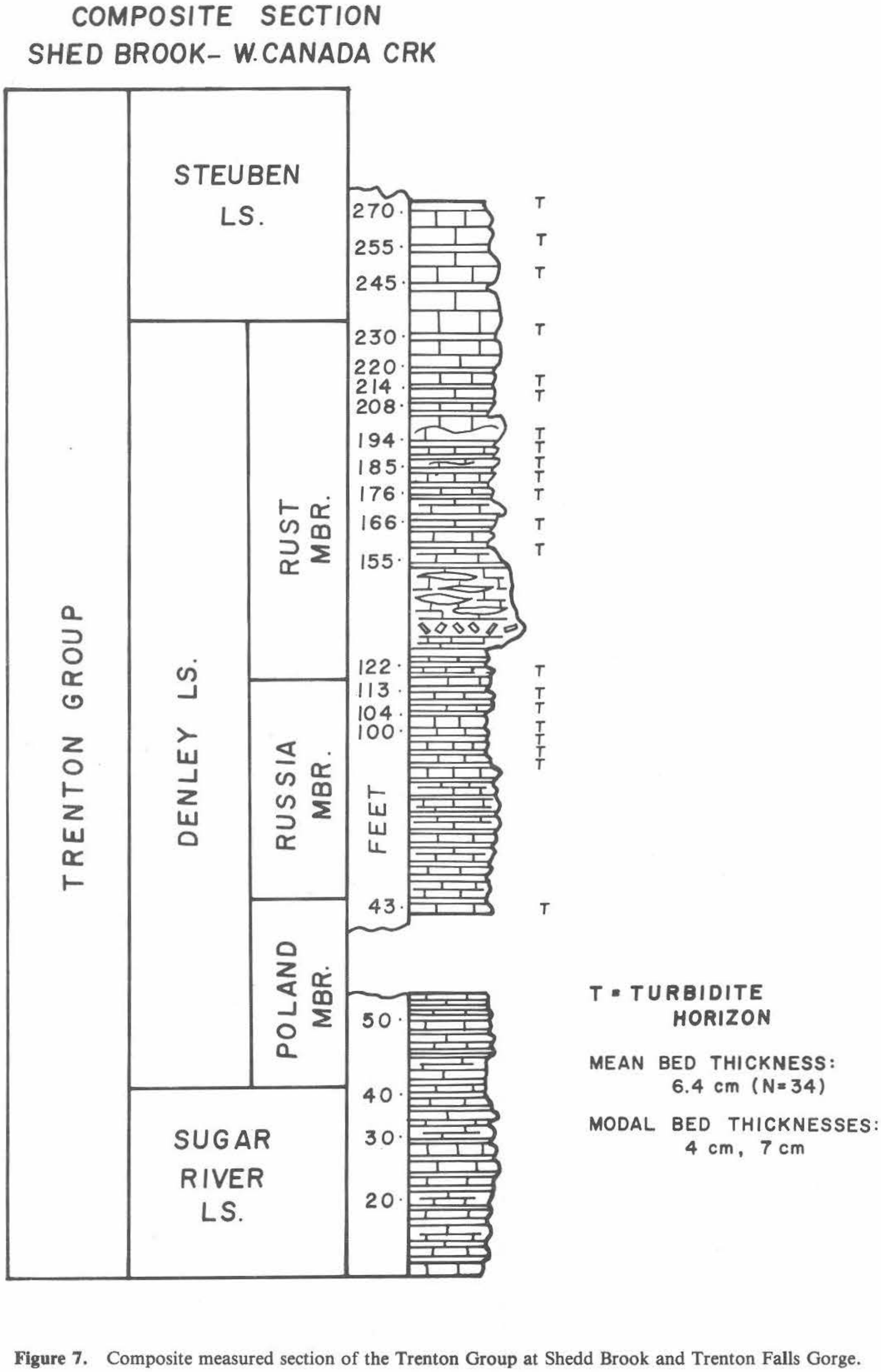

Figure 7. Composite measured section of the Trenton Group at Shedd Brook and Trenton Falls Gorge.

sampling for a study on the morphology and environmental preferences of the trepestome bryozoa, Prasopora (figure 7). Prasopora, a gum-droped shaped colony, covers bedding planes in the Sugar River and Denley Limestones, and is the index fossil for the Sugar River. Entire limestone beds below the colonies were collected and thin sectioned for petrographic analysis. The 250 foot thick Trenton type section at Trenton Falls Gorge (Denley Limestone: Poland, Russia,and Rust Members, Steuben and Hillier Limestones) was sampled in this fashion. Subsequent outcrops were sampled in a more methodical manner. Large thin sections (5 x 7 inch up to 5 x 14 inch) on 1/4 inch glass plates were made to preserve entire limestone beds and the upper and lower contacts with interbedded shale horizons.

The Denley Limestone is characterized by the repetitive interbedding of limestone and shale couplets. What is the origin of this cyclic bedding? Do the shale horizons represent episodic events of deeper water sedimentation or are they the result of dewatering and compaction of an original mixed carbonate-terriginous clay sediment? Are the carbonate beds resedimented and the shale autochtonous? These questions, orignally asked as part of the Prasopora paleoenvironment study, became important in their own right for the information they might contain on the origin of limestone-shale couplets in general. Mehrtens (in press) presented the evidence that 70% of the limestone beds sampled in the Denley Limestone at the Trenton Falls Gorge type section (Figure 7) were bioclastic turbidites. Subsequent studies of other stratigraphic sections in the Mohawk and Champlain Valleys indicate that bioclastic turbidites are a common occurence in the Denley Limestone, Dolgeville Facies and undifferentiated Trenton sediments. The bioclastic turbidites are recognizable by internal structures identifiable by Bouma Units. Comparison of the Denley Limestone turbidites to other examples of bioclastic turbidites cited in the geologic literature (Thompson and Thommasson, 1969; Cook and Taylor, 1977; Yurewicz, 1977; Scholle, 1971; Garrison and Fisher, 1975, among others) indicates that they share many characteristics: (1) division of beds into a lower graded and upper mudstone portion; (2) graded carbonate sand units capped by laminated or cross laminated units; (3) laterally extensive beds; (4) associated slump fold horizons; (5) planar bases with bioturbated tops; (6) ubiquitous grading and horizontal laminae. Figure 8 compares the internal structures of bioclastic turbidites to those of a clastic turbidite.

Although recognition of bioclastic turbidites is dependent on petrographic study it will be possible to see some features associated with turbidite sedimentation on this field trip. Examination of several of the large thin sections illustrates many of the features visible with petrographic study.

Examination of Figure 9 illustrates a typical bioclastic

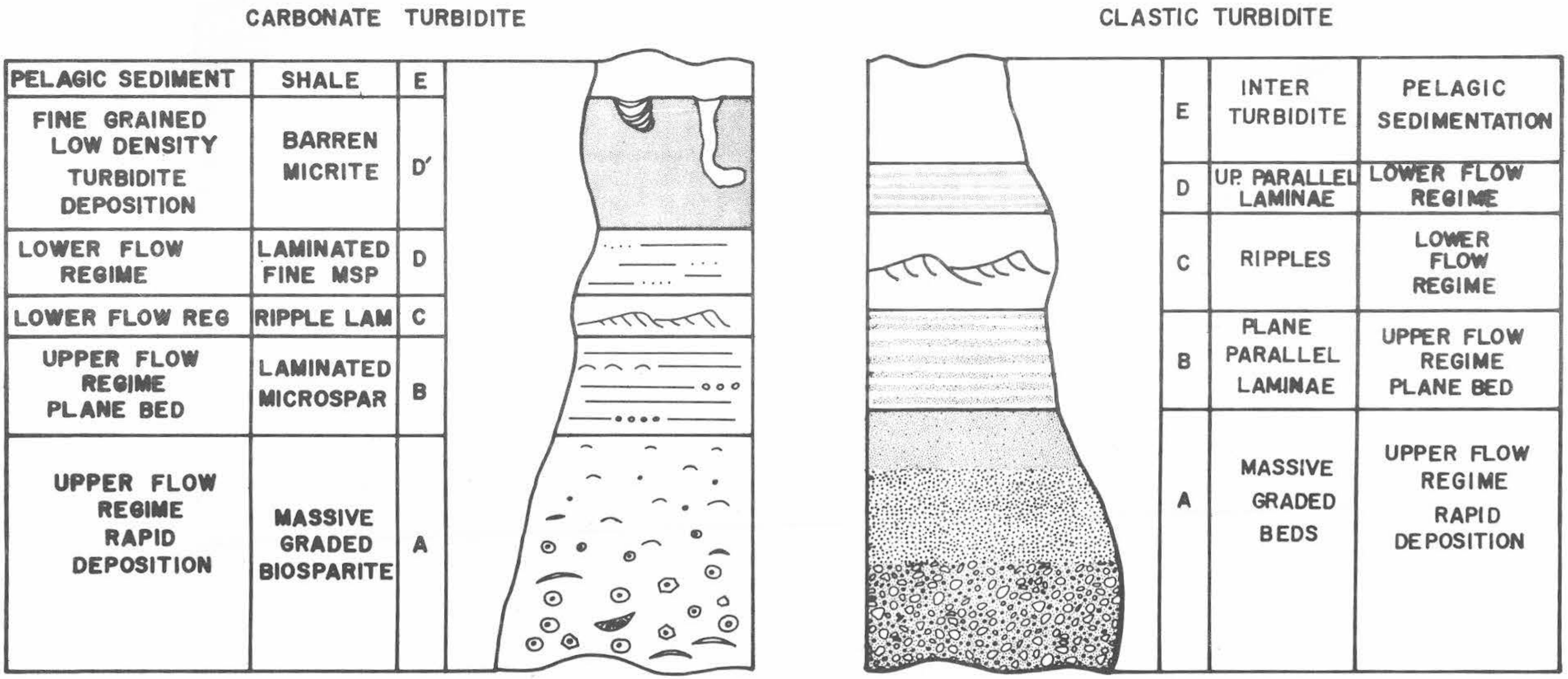

Figure 8. Comparison of the internal sedimentary structures and their hydrodynamic interpretations for clastic and carbonate turbidites. After Walker (1965).

Figure 9. A large (8 x 8 inches) thin section on a glass plate which illustrates a Tabd ′ sequence. Note the descending burrows.

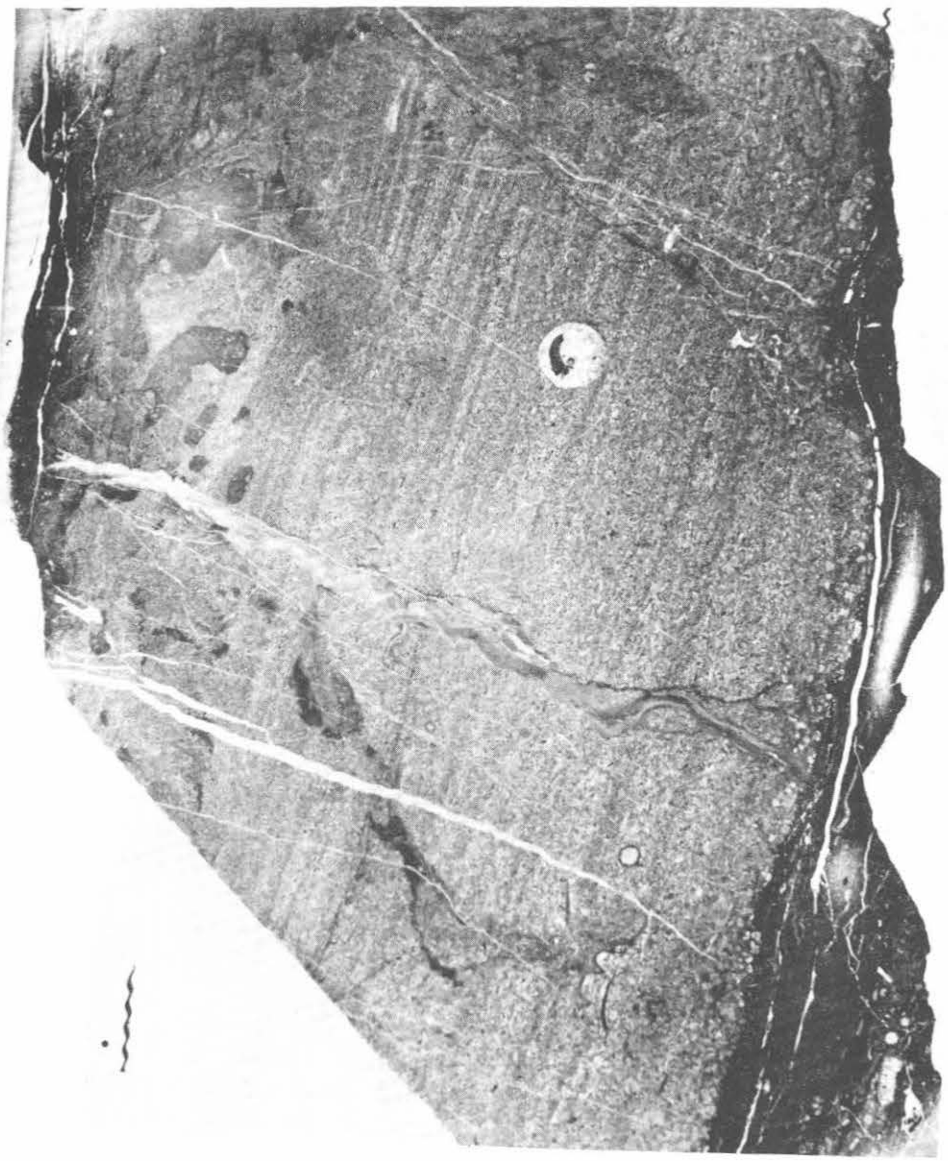

Figure 10. Another large thin section illustrating Tab Bouma Units.

turbidite from undifferentiated middle Trenton Group. This thin section, 9 x 10 cm in size, is bounded on top and bottom with calcareous shale units interpreted as interturbidite deposits, or Te units. The turbidite has at its base a 1 cm thick layer of disarticulated crinoid ossicles analogous to the traction carpet of coarse quartz sand of a clastic turbidite. This layer (Ta) gerades into a less fossiliferous laminated microspar (Tb). Close examination of Figure 8A suggests that the grading visible through the bulk of the thin section owes its origin to three processes, illustrated in Figure 10: (1) concentrations of ossicles at the base and a vertical diminuition in their size and abundance; (2) a vertical size change in highly fragmented silt-sized skeletal fragments; (3) vertical decrease in size of the interparticle spar between grains of the laminated micrite matrix. This size change in the spar is thought to reflect a vertical change in interparticle porosity (Choquette and Pray, 1970). The origin of the interparticle porosity grading and subsequent spar cement is thought to be related to the thickness of the current-laminated horizontal laminae, and is therefore primary in origin. The unit overlying the current-laminated Tb unit is a highly bioturbated barren micrite horizon (td'). Internal structures within the burrows confirms that they are descending from the overlying shale and thus reflects the recolonization of the emplaced sediment.

Figure 10 is a thin section of a 6 cm thick turbidite horizons from the Denley Limestone (Russia Member). This thin section contains a poorly sorted and poorly graded biospar at its base (Ta). Numerous fragmented, unidentifiable pelmatozoan debris is visible. This 2 cm thick basal unit grades up into the horizontally laminated unit (Tb) through an interval of 0.5cm. Within this interval the skeletal fragments decrease in the size and abundance vertically. There is an associated decrease inthe size of interparticle spar. The effects of shelter porosity are clearly visible beneath the trilobite fragment. The horizontally laminated Tb unit is 4 cm thick and composed of current-laminated micrite, terriginous silt and microspar cement.

Figure 11 is a thin sectio of a 9cm thick turbidite horizon from the Denley Limestone (Poland Member). This thin section exhibits a thin (1 cm) basal lag of poorly sorted, fragmented skeletal debris and spar cement (Ta). Grading is not well developed. This basal lag is immediately overlain, without any gradational interval, by 8 cm of barren micrite (Td'). Except for a few small worm burrows, this micrite horizon is structureless. This is a very characteristic turbidite occurence. Examination of Table 1 indicates that turbidites composed of a basal lag overlain by barren micrite (Tad')makes up 11% of the Trenton turbidites at Trenton Falls Gorge and 40% of the turbidites from undifferentiated Trenton near Beekmantown. This turbidite is very important because it is the most likely sequence to be overlooked or misidentified as being the result of resuspension and settling due to storms.

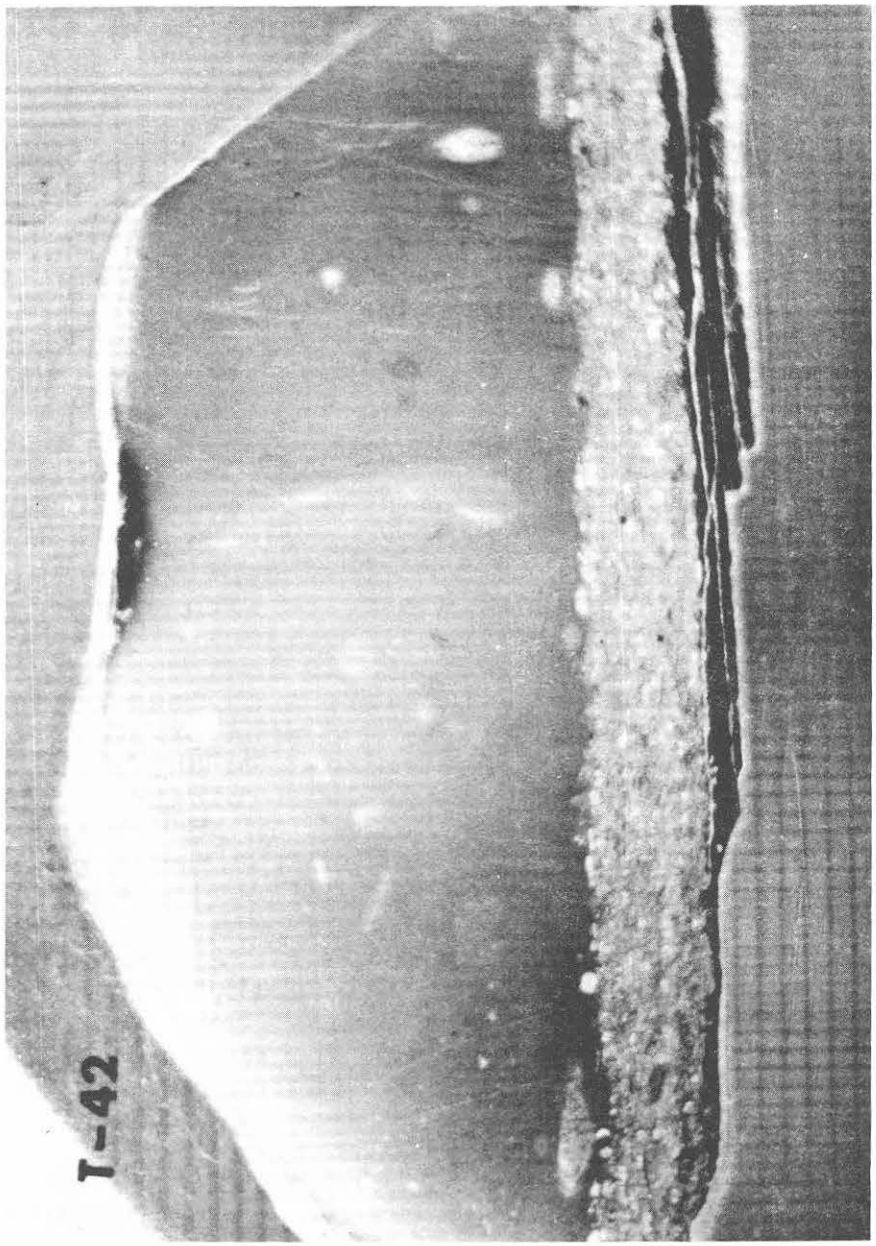

Figure 11. Another large thin section illustrating Tad ′ sequences, a common turbidite occurrence.

These two mechanisms can be differentiated by a simple statistical measure, a plot of grain size (measured by crinoid ossicle diameter) vs graded bed thickness (illustrated in Figure 12). The rationale and threoretical basis for this relationship was discussed by Mehrtens (in press). Basically, sediment that is simply resuspended and allowed to settle as a function of its settling velocity will produce a graded bed. The size of the grains within this graded layer will not however, be related to the hydraulics of flow, and will therefore not show size sorting. A correlation between bed thickness and grain size exists in turbidites, however, because the hydraulics of flow control not only the size of the particles moved, but the thickness of the bed. This relationship (bed thickness and grain size) has been described for modern bioclastic turbidites (Crevallo and Schlager, 1980).

Figure 13 illustrates the nature of the vertical changes visible within the Trenton trubidites. From bottom to top there is a progressive decrease in allochem abundance and associated interparticle cement. Accompanying this decrease is the expected increase in carbonate mud. The effects of varying source area on turbidite composition is seen by comparing the Bouma Units present in the Denley versus the Dolgeville Facies. The Dolgeville, as described earlier, consists of barren, horizontally laminated micrite with planar bases and thick shale interbeds (Figure 5). Horizons of coarse graded skeletal debris (Ta) and horizontally laminated skeletal fragments and carbonate mud (Tb) are much less common than in the Denley Limestone turbidites, and when present, contain an overall higher percentage of micrite. Tc units are more commonly preserved in the Dolgeville turbidites, probably because the lack of crushed allochems in Td limited the reworking which subsequently removes Tc. Based on these differences, the turbidites of the Dolgeville Facies could be interpreted as being more distal than those of the Denley. However, as Scholle (1971) and Welch (1978) demonstrated, the same features which distinguish "distality" (overall fine-grained size, Tcd) could represent deposits derived from a dominantly fine-grained source. The Dolgeville turbidites could be quite proximal in origin, having been derived from portions of the shelf which had already foundered sufficiently to have accumulated a supply of carbonate mud. Alternately, they could represent the distal portion of a Denley turbidite. If the latter were true we might expect to see a change in bed thickness from the proximal Denley turbidites to the more distal Dolgeville turbidites. In fact, the opposite is seen. The Denley turbidites average 7 cm in thickness while those of the Dolgeville average 8-10cm in thickness.

Mehrtens (in review) discussed the criteria for distinguishing turbidites from storm generated ebb currents. Many recent papers have described the internal structures generated by storm processes (see Nelson, 1982; Kriesa, 1981, Seilacher, 1982 and Aigner, 1982 for examples). The following

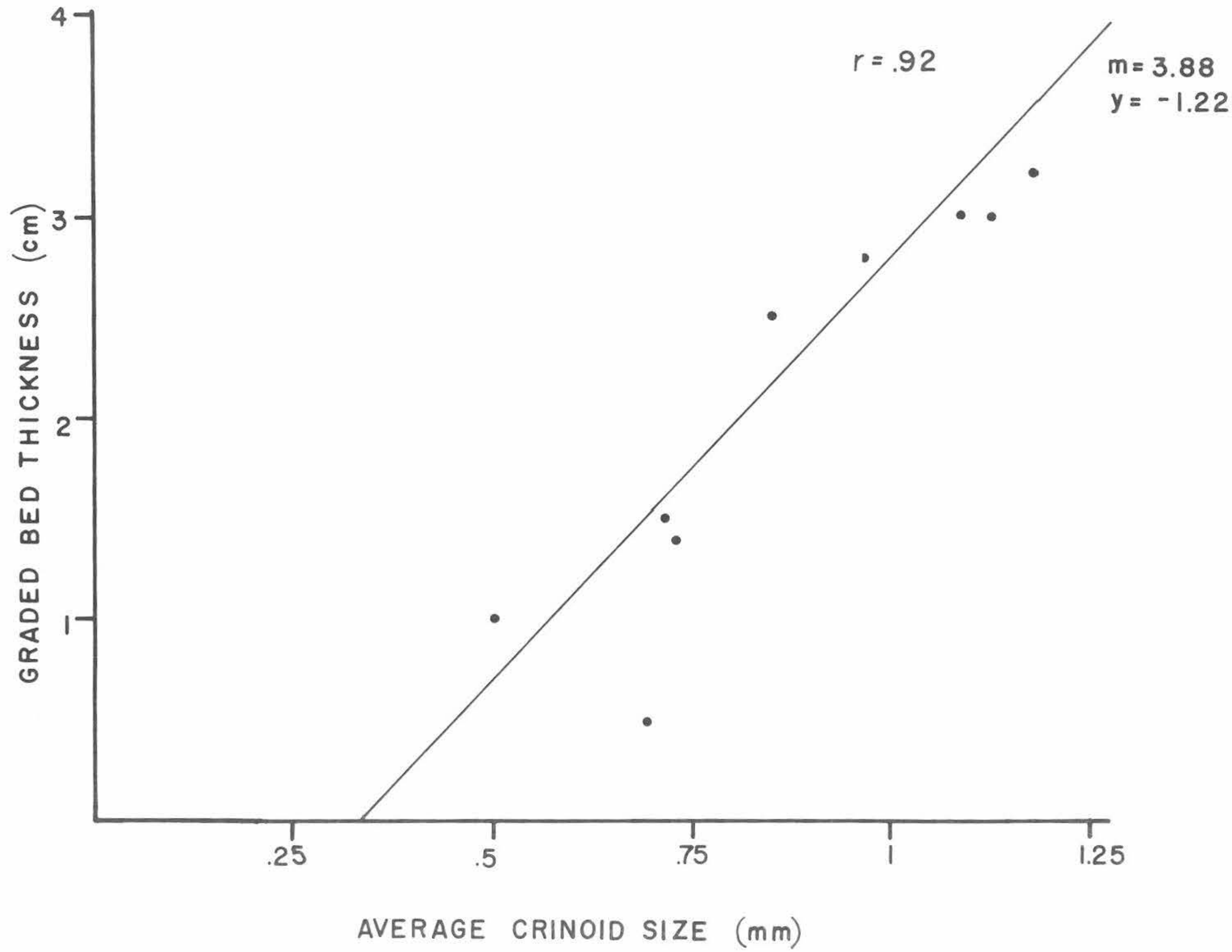

Figure 12. "Best fit" line and correlation coefficient (0.92) produced by linear regression analysis. This plot indicates a statistically valid positive relationship between the size of the fossils being deposited and the hydraulic conditions of the flow regime. Because there is a correlation between the size and the velocity of a turbidite, this observed relationship between grain size and bed thickness is characteristic of a turbidite.

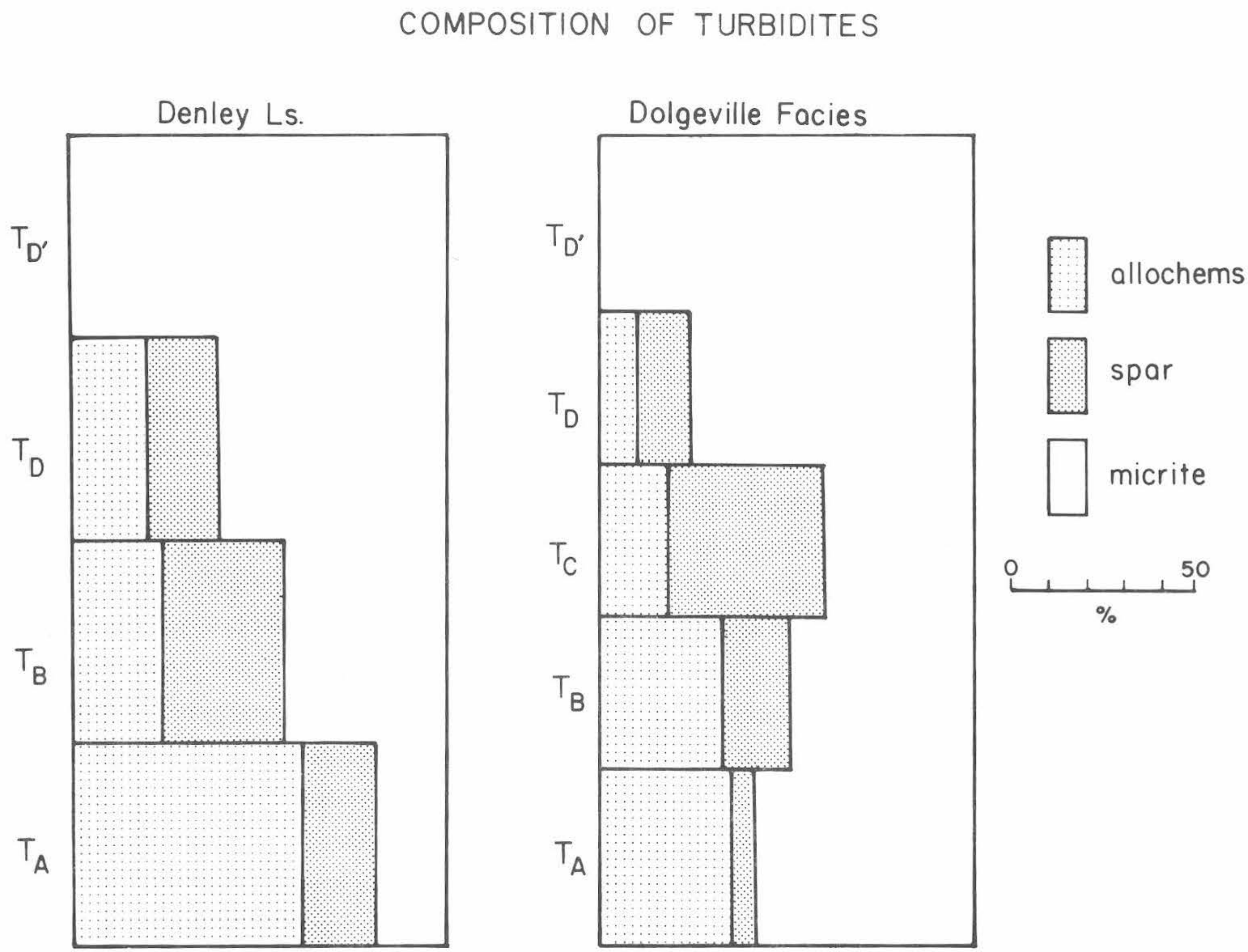

Figure 13. Graphic representation of the vertical changes which occur within the Denley and Dolgeville Turbidite beds.

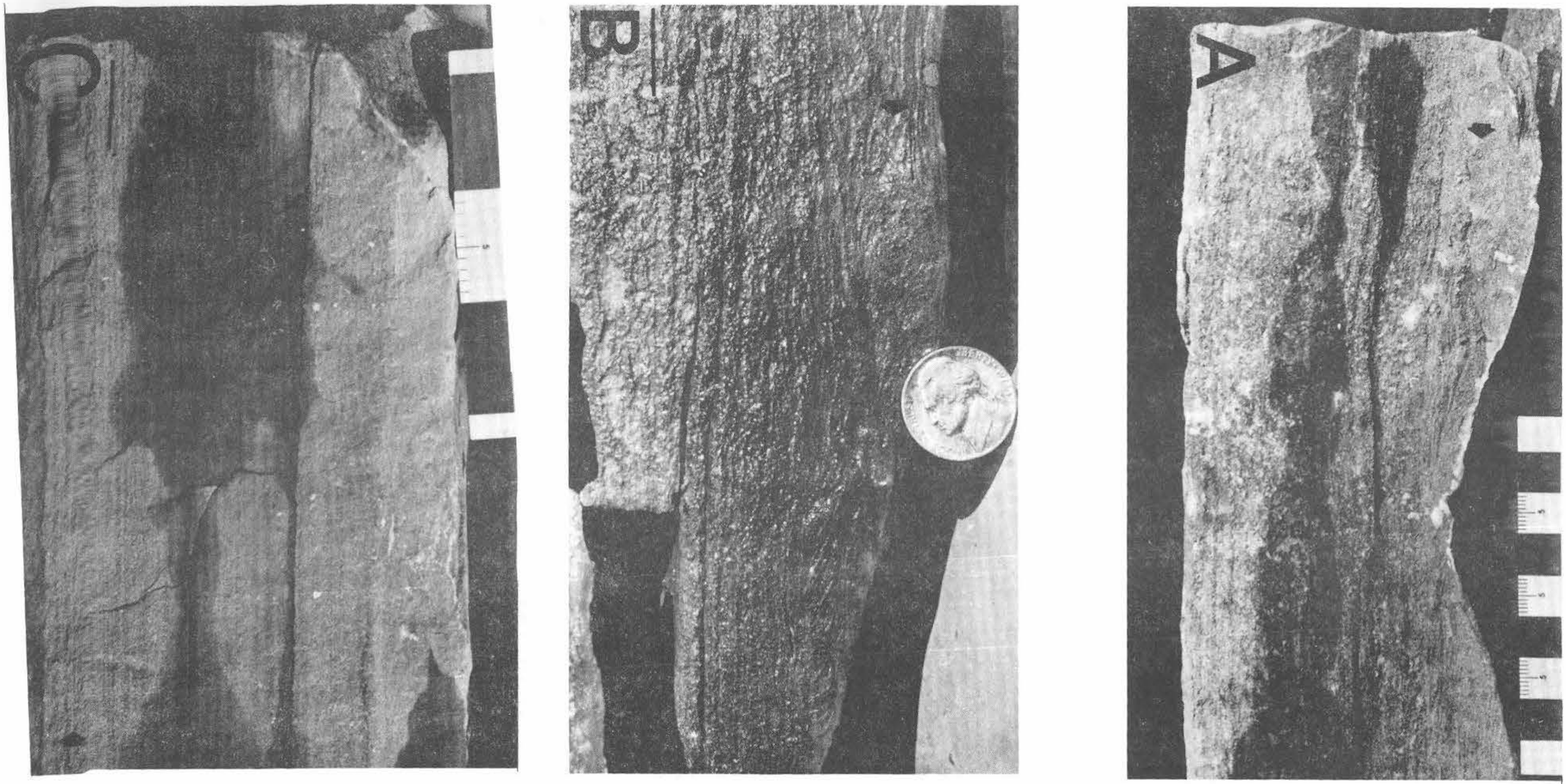

Figure 14. Photographs of the weathered surfaces of outcrops where cross laminations are most visible. **A.** Cross lamination of Tc is well developed over a horizontally laminated Tb. **B.** Laminations of Tc distorted into convolutions. **C.** Basal concentration of crinoid ossicles (lower arrow) of Ta is overlain by horizontally laminated Tb, and at upper arrow, cross laminated Tc.

characteristics for storm-generated deposits were frequently cited: (1)internal structures, although similar to Bouma Units, lack the repetitive consistency of turbidites; (2) association with Skolithos burrows; (3) lateral continuity on the order of 10's of meters; (4) strongly lenticular bedding is present; (4) multiple paleocurrents, one fairweather and another storm in origin; (5) association with trough cross lamination from wave oscillation; (6) association with hummocky cross stratification; (7) lateral facies grading into normal (fairweather or below wave base) marine shallow water litho- and biofacies; (8) escape burrows; (9) high degree of post-storm bioturbation.

Based on this list of criteria, the turbidite origin for the limestone/shale couplets is clear: (1) the internal structures occur with regular frequency throughout a stratigraphic sequence (see Figure 7); (2) Skolithos burrows are not associated with the Denley Limestone or Dolgeville Facies; (3) beds exhibit lateral continuity on the order of 100's meters; (4) the same paleocurrents are derived from measurement of ripple cross lamination (Figure 14) within the limestone beds as is obtained from graptolite and cephalopod orientations in the shale horizons (Cisne and Rabe, 1978); (5) trough cross-laminations associated with wave oscillation are absent in the Denley and Dolgeville; (6) hummocky cross stratification is also absent; (7) facies changes are abrupt and rarely gradtional; (8) escape burrows are not seen, rather all burrows are recolonization burrows, the result of repopulation of the resedimented limestone; (9) frequency of turbidite emplacement was sufficient to restrict post-emplacement bioturbation (a criteria cited by Dott, 1982, as a means of distinguishing storm and turbidite generated graded beds). Based on this evidence, association with slump fold horizons, statistical analysis and hydraulic interpretation of textural data, and consistency with the depo-tectonic model presented by Cisne, et al (1982), the limestone and shale couplets of the Denley Limestone and Dolgeville Facies can be interpreted as bioclastic turbidites.

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ROAD LOG

This Road Log starts at Thruway Exit 30, Herkimer, New York

Mileage	
0.0	From Thruway, follow signs for Rt. 28N, through Herkimer to Middleville and intersection with Rts. 29 and 167.
1.7	Make left and continue north on Rt. 28 for 1.7 miles to Castle Rd. Turn right.
2.0	Go 0.3 miles up Castle Rd. to road coming in sharply on left.
2.1	Take sharp left, descending hill, and passing lower portion of City Brook on right.
2.6	Continue 0.5 miles to T junction. Take sharp right onto White Creek Road.
3.1	Go 0.5 miles to small bridge over Buttermilk Creek. Park immediately beyond bridge. STOP 1.

STOP 1 Buttermilk Creek

Park immediately beyond the bridge and descend to Buttermilk Creek. The section downstream of the bridge is part of the Black River Group (Gull River), consisting of calcareous limestone. Underneath the bridge and a short distance upstream are ledges of coarsely crystalline limestone alternating with massive calcareous sandstone beds. These are the lagoonal sediments of the Napanee and Selby Limestones. These rocks have been described by Titus and Cameron (1976) and are characterized by their aphanitic texture, birdseye structures, massive bedding, and disseminated quartz sand. This facies continues in a series of benches going upstream. The contact of the Napanee and Selby with the Kings Falls occurs immediately above the pool below the first waterfall. The Kings Falls is differentiated from the underlying facies by the first appearance of fossiliferous calcarenites that are thinner bedded and exhibit lateral thickening and thinning of individual beds. The Kings Falls continues upstream to the larger waterfall, consisting of beds of calcarenite and bioturbated calcilutite. The lateral lensing of the beds, so characteristic of the Kings Falls is readily visible.

Mileage

STOP 1 (cont'd)

While at this stop you should: (1) examine the Napanee and Selby (lagoonal) facies as this is the only stop where they will be seen on this trip; (2) examine the Kings Falls, noting in particular the features which may suggest that it accumulated near wave base. Are there any internal structures within the limestone beds? How much shale, if any, is present? Are the beds laterally continuous? How frequent was bioturbation? Where are the fossils - within limestone beds or on bedding planes?

3.6 Return to cars and retrace steps on White Creek Rd., going 0.5 miles to road coming in on left.

4.2 Make sharp left. Go 0.6 miles (passing lower City Brook on left) to junction with Castle Rd.

4.6 Make sharp left onto Castle Rd., ascend hill 0.4 miles to Farrington Rd. (dead end).

4.7 Take left onto Farrington Rd. and immediately park near bridge crossing City Brook. STOP 2.

STOP 2 City Brook

After disembarking from the cars, descend to City Brook upstream of the bridge. The section downstream is one of the most beautiful exposures of the Black River, Selby, Napanee and Kings Falls Limestones in the Mohawk Valley but is off limits. Under the bridge, and continuing up section, are small benches of the Sugar River Limestone. These bedding planes are distinguished by their abundant Prasopora colonies, along with the typical Sugar River benthic fauna described in Titus and Cameron (1976). Compare the Sugar River Limestone to the Kings Falls Limestone seen at STOP 1. How does it differ? What is the evidence that it might have accumulated at slightly deeper depths? How laterally continuous are these beds? Do they exhibit the same degree of lateral lensing? Is there a change in the relative percentage of calcarenite versus calcilutite beds? The contact of the Sugar River and the Denley Limestones occurs in the small waterfall. The basal Denley seen here is similar to the basal Denley seen at the next stop (Shedd Brook) but is very different than the basal Denley at Trenton Falls Gorge.

Figure 15. An example of a compressed stratigraphic section, this one at Canajoharie Creek. The observer's feet are on the Lower Ordovician dolomites, which are unconformably overlain by limestones of the Trenton Group. The color change from light to dark marks the contact with the overlying Canajoharie Shale.

Mileage

STOP 2 (cont'd)

Here the Denley is much less fossiliferous and much finer-grained. Note the nodular limestone bedding. Bedding planes are not littered with fauna, except for an occasional trilobite and cephalopod. Continuing up the stepped benches note the abundance of calcilutite, mostly all barren. The dominant two lithologies of the Denley here are: (1) a barren calcilutite with planar bottoms and reworked (bioturbated) tops and (2) a nodular, more argillaceous calcilutite. These lithologies continue upstream until the waterfall is reached. Here, on the cliffs above the stream, the contact with the overlying Dolgeville Facies is seen. The Dolgeville is characterized by the interbedded planar calcilutite with thick interbeds of shale (see Figure 15). This City Brook section is interesting because it records a very rapid, non-gradational facies change in the Sugar River/Denley contact. It also is unusual in that the Denley here is such a deep water deposit. Examination of the limestone beds reveals a dominance of planar laminations interpreted as Td and structureless, bioturbated micrite interpreted as Td'. While at this stop you should ponder the nature of the facies change from the Sugar River to Denley. Do you think it's the result of lateral facies mosaicing accompanying gradual transgression, or, could it be the result of a rapid drop of the fault block, drowning the shelf?

5.2 Return to cars and retrace steps, descending Castle Rd. 0.5 miles to Rt. 28.

7.8 At intersection of Castle Rd. and Rt. 28, turn right towards Newport, going 2.6 miles to flashing light in town of Newport.

7.9 Turn left at flashing light, going 0.1 miles and crossing West Canada Creek.

8.0 At T intersection, go left.

9.2 Go 1.2 miles and ascend hill.

9.4 Park car over crest of hill. STOP 3.

Mileage

STOP 3 Shedd Brook

After parking at the top of the hill, cross the road and descend to the stream bed. Walk down to the base of the section. Exposed at the base of Shedd Brook is the upper Kings Falls Limestone. It should look similar to the upper part of the section at Buttermilk Creek, that is, coarse calcarenite to calcirudite beds which exhibit lensing. All of the Sugar River Limestone is exposed in this stream cut, and comparisons can be made between the Kings Falls and Sugar River. Study the Sugar River lithologies and bedding styles to draw some conclusion about environment of deposition relative to wave base. This is an important point because the Sugar River is interpreted to represent "normal" carbonate accumulation below wave base. If this is what "normal" below-wave base sedimentation looks like, what interpretations can be made about the overlying Denley Limestone? The contact of the Sugar River and the Denley Limestone occurs at the top of the waterfall. Examine the upper horizons of the Sugar River. Note the abundance of relatively unfossiliferous calcilutite beds with shale interbeds. How gradational are the Kings Falls to lower Sugar River to uppermost Sugar River facies? The top of the waterfall represents the last of the fossiliferous calcilutite and calcarenite beds in this section. The percentage of interbedded shale is also increasing. A bentonite horizon is present in the bushes on the left side of the stream. There is a long bench in the stream bed of the nodular barren calcilutite beds with thick shale interbeds of the Denley Limestone. This should look similar to the Denley seen at the previous stop. Up stream at the bend in the brook, Dolgeville beds of planar carcilutite and shale are seen. Passing under the road you continue to walk on nodular bedded, calcilutite horizons of the Denley Limestone. About 50 feet above the road on the right side of the stream, there is a small exposure of interbedded nodular calcilutite and planar calcilutite beds with bioturbated tops, also of the Denley Limestone. Two more bentonite horizons are located in this small cliff. The stratigraphic section seen along Shedd Brook is very similar to that in City Brook but it is worth walking this section to demonstrate:
(1) the gradational Kings Falls to Sugar River section;
(2) interbedded bentonite horizons.

10.8 Return to cars, turn around carefully on hill, descend hill to intersection.

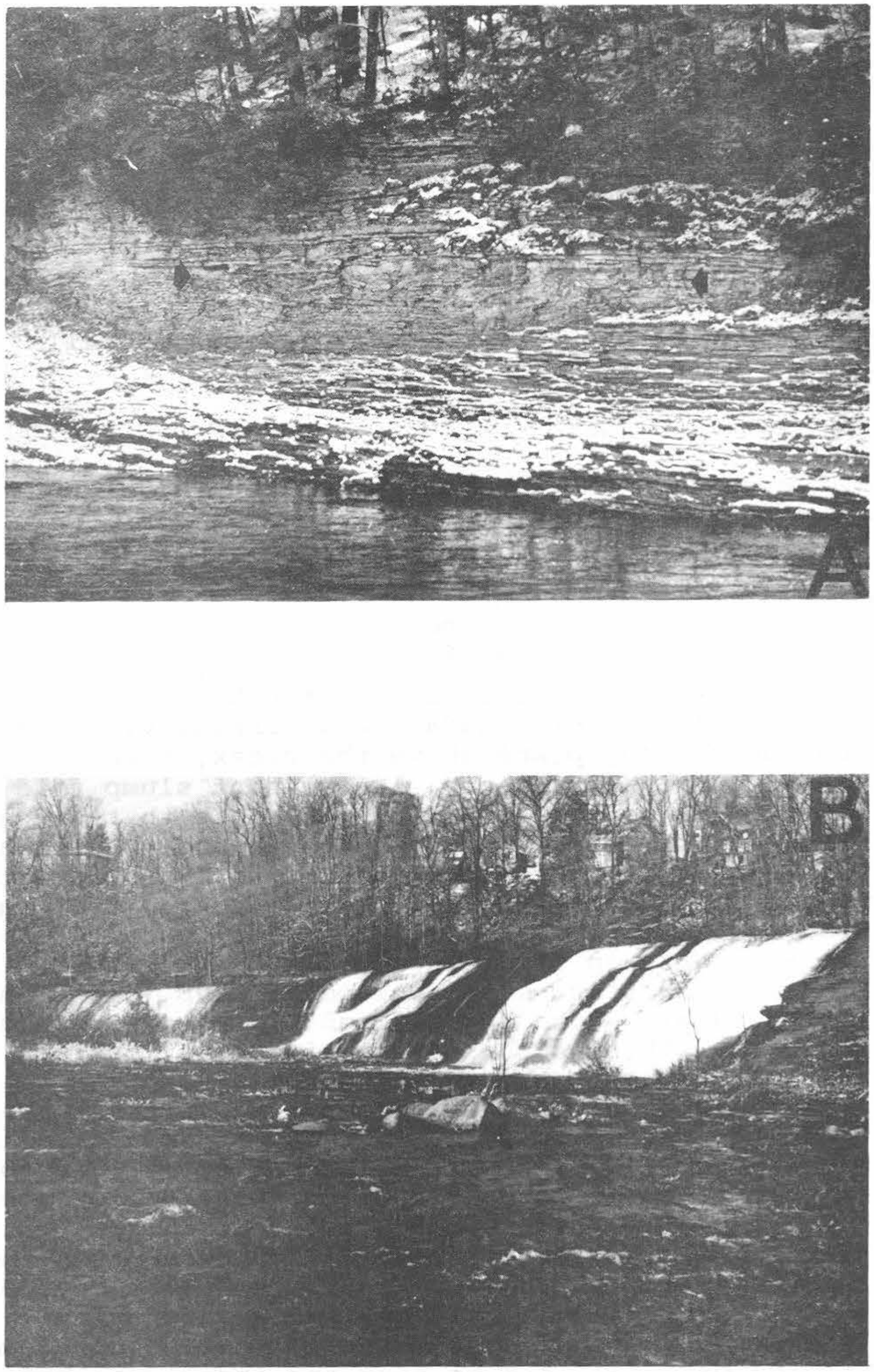

Figure 16. **A.** Exposure of a slump fold horizon (between arrows) in the Denley Limestone on West Canada Creek. **B.** Exposure of the Denley Limestone along West Canada Creek illustrating the large degree of lateral continuity of beds (houses above waterfall for scale).

Mileage

11.1 Turn left, recross West Canada Creek to flashing light.

14.8 Go left (north) on Rt. 28 to Poland.

21.5 Continue north on Rt. 28 to unmarked triangular intersection. Go straight off Rt. 28 to stop sign.

22.5 Make right and go 1 mile to intersection. West Canada Creek on right, access to Trenton Falls Gorge is ahead.

25.0 Go left, following paved road into Prospect.

25.2 At stop sign in village of Prospect, go right and descend hill to bridge.

25.3 Park on far side of bridge. STOP 4.

STOP 4 Prospect (needs special permission from Niagara Mohawk Power Corp.)

After parking the cars across the bridge, descend the cliff to West Canada Creek carefully. Standing on the bedding plane above the creek, turn around and examine the cliff face. A prominent slump fold horizon is visible (Figure 16). Note that the slump fold can be traced laterally to an undisrupted horizon of calcarenite. The nose of the fold is characterized by brecciated clasts of limestone so the sediment was fairly lithified at the time of emplacement. Paleoslope measurements on this and other slope indicators (aligned fossils) indicate a paleoslope of S80E (Cisne and Rabe, 1978). Note the sediment in which the slumping occured. This is the Denley Limestone, but a very different Denley than what has been seen at Stops 2 and 3. Note how fossiliferous some of the calcarenite and calcirudite beds are. The strophomenid Rafinesquina is especially abundant, as is Prasopora. Interbedded with this fossiliferous facies of the Denley are lithologies we have already seen: nodular, bioturbated calcilutite and calcilutite with planar bases and bioturbated tops (Td'). Also present are beds of graded skeletal debris (Ta), and horizontally laminated skeletal debris (Tb). Looking upstream a wide waterfall can be seen and individual beds can be traced across this exposure. Beds which exhibit this degree of lateral continuity, with negligible change in thickness could not have accumulated near wave base.

Mileage

25.6	Return to cars, turn around, return to intersection in village of Prospect.
28.2	Make left and return to intersection by West Canada Creek and Gorge access.
34.8	Go right and return to Rt. 28 south to Poland.
38.6	Continue on Rt. 28 to flashing light in Newport.
42.9	Continue south on Rt. 28 to Middleville and junction with Rts. 28, 29, and 167.
55.6	Turn left onto Rt. 29 and follow to Dolgeville.
55.7	At stop sign in Dolgeville, turn left (staying on Rt. 29) to bridge. Go straight across bridge and make immediate (poorly marked) right onto Dolgeville Extension Road.
56.1	Go along Dolgeville Extension Road, with East Canada Creek and the Daniel Green Factory on your right.
56.8	At T intersection, turn left onto Dolgeville Ave. and go 0.7 miles. Park on hill before guard rail. STOP 5.

STOP 5 Dolgeville

This stop along East Canada Creek in Dolgeville is made for examination of the basinal shales. Parking on the hill and descending to the stream below the dam, the calcareous turbidites and interbedded shale horizons of the Utica shale are exposed. The calcareous turbidites are identifiable by their buff yellow weathering. Tc and Td horizons are the only Bouma Units present. The thick shale interbeds contain numerous good graptolites. Based on the Bouma Units present within the turbidite horizons and the thickness of the interturbidite deposits, it is possible to interpret this section as representing a distal turbidite facies. The source of the calcareous horizons is the foundered shelf (Dolgeville Facies) intertonguing with the basinal terriginous muds. Continuing down to the pool at the base of the waterfall, drag folds associated with the Little Falls Fault are visible.

57.3	Return to cars. Ascend hill to turn around.

Mileage

58.5 Descend hill into Dolgeville.

58.9 Turn right onto Dolgeville Extension Rd. and return to bridge.

66.1 Turn left onto Rt. 167 and follow south. Lunch stop at roadside park. Stay on Rt. 167S through Little Falls until Rts. 5 and 167 divide.

68.3 Stay on Rt. 167S, crossing Barge Canal, to junction with Rt. 5S. Go east (left) on Rt. 5S.

69.2 Passing Kings Falls, Sugar River, and Utica lithologies on right and left of road (abandoned quarry on left).

83.8 Continue east on Rt. 5S to Fort Plain.

83.9 At junction of 5S and 80, take a dog-leg left, staying on 5S.

88.0 Continue into Canajoharie and junction of Rts. 5S and 10.

88.1 Go right into triangular square and traffic light. Go right at traffic light and make immediate right onto Moyer St.

88.5 Go up Moyer St. hill and make right onto Floral Ave.

88.7 Proceed to end of Floral St. and park. STOP 6.

STOP 6 Canajoharie Creek

Descend to the creek from the parking lot. The creek bed is Chuctanunda Creek Dolostone of Lower Ordovician age. The lower cliff face contains the unconformity between this dolostone and the interbedded limestone and shales of the Kings Falls Limestone. Looking up the cliff the Kings Falls is overlain by the Canajoharie Shale (Figure 15). In other words, all of the Lower Trenton is missing, with the exception of a thin section (15 feet) of Kings Falls, and all of the upper Trenton is missing. Cameron (1971) described this section of limestone as Sugar River, but noted that it represents a shallower water environment than the Sugar River seen to the northwest (Stops 2 and 3). This unit can be better described not by the fossil assemblage it contains,

Mileage

STOP 6 (cont'd)

but by the lithofacies present. The bedding style is identical to that of the Kings Falls seen earlier; laterally discontinuous beds exhibiting lensing and cross laminations. Looking at this section you can ask yourself: How can wave-reworked limestones and basinal shales be juxtaposed with no intervening deeper water limestones? Why are there no lower Trenton units between the Kings Falls and the Lower Ordovician dolomites? Do these abrupt facies changes represent lateral mosaicing produced by sea level changes or does it seem more probable that movement on a syn-depositional block fault could generate this type of sequence?

88.9 Return to cars. Drive out Floral St. to intersection with Moyer St. Turn left.

89.0 Descend Moyer St. hill. Make left into triangular square and traffic light. Make right onto Rt. 10.

89.3 Go to intersection of Rts. 10 and 5.

94.6 Go left (west) on Rt. 5 to a sign for the Palatine Church.

94.7 Take a right onto Old Mill Rd., cross Caroga Creek and part. STOP 7.

STOP 7 Caroga Creek

After parking, descend to the creek below the bridge under Route 5. The section below the bridge and downstream are undifferentiated Black River deposits. Depending on the height of the stream, variable amounts of rock are exposed here. At low stage, aphanitic dolostones with birdseye structures are visible. The section between the two bridges has many covered intervals but ledges of coarsely crystalline limestone outcrop and are interpreted as Kings Falls. Under the second bridge ledges of laminated calcarenites outcrop, exhibiting planar tops and bottoms and lacking fossils. Upstream of these are beds of the bioturbated, barren calcilutites. These last two lithologies are characteristic of the Denley Limestone as seen on City and Shedd Brooks. A covered interval occurs above this and below the cliff with Utica Shale. This section then, is another

Mileage

STOP 7 (cont'd)

thinned section of Trenton. Both the lower Trenton and upper Trenton units are missing or greatly reduced in thickness. Because the quality of this exposure is not equal to that at Canajoharie Creek, it is not as spectacular, but the significance of the facies changes is equally important. Note that as one progressed east for the last two outcrops, the magnitude of the variation in stratigraphic sections became more significant. This is in keeping with our moving further into the foreland basin from the stable shelf represented in the earlier outcrops.

95.0 Return to cars and continue down Old Mill Rd., past Palatine Church, to junction with Rt. 5.

108.0 Turn right (west) on Rt. 5 to Little Falls.

116.0 Continue west on Rt. 5 through Herkimer.

116.5 Take left on Rt. 28 south to Thruway.

ANIMAL-SEDIMENT RELATIONSHIPS IN MIDDLE ORDOVICIAN HABITATS

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INTRODUCTION

In recent years a number of workers have presented arguments to suggest that utilization of infaunal life strategies by marine organisms has varied through the Paleozoic. Evidence for a progressive "infaunalization" of Paleozoic communities comes from functional analysis of trace fossils (Seilacher, 1977), patterns of taxonomic richness of skeletonized bioturbating organisms (Thayer 1979, 1983), and the sedimentary record of the effects of burrowing organisms (Garrett, 1970; Sepkoski, 1982; Larson and Rhoads, 1983). Between the Middle Ordovician and Early Devonian, it seems that the depth of biogenic reworking of sediment in marine shelf settings increased from essentially zero to over 5 cm. Exploitation of buried food resources by deposit-feeders propels this infaunal invasion (Larson and Rhoads, 1983). Miller and Byers (1984) present an opposing view, that infauna are abundant and diverse throughout the early Paleozoic. At present, there is no consensus as to the style and degree of biogenic reworking in the early Paleozoic.

This field trip is meant to provide an opportunity to examine some of the evidence that has convinced me that biogenic reworking in Ordovician sedimentary environments was much reduced in comparison to recent counterparts. We will visit a series of four outcrops of Middle Ordovician carbonate rocks in the Black River Valley of northwestern New York (Figure 1). Our trip will take us on a transect of facies from the intertidal carbonates of the Black River Group through a range of shallow to deep shelf and basinal environments in the Trenton Group. The goal is to examine Ordovician animal-sediment relationships and their sedimentologic and paleoecologic implications.

BIOTURBATION AND DEPOSIT-FEEDING

By definition, all benthic organisms have some interaction with the sediments of the seafloor. Since physical and chemical properties of the bottom sediments are an important ecologic factor in the distribution of both epifauna and infauna, activities of an organism or group of organisms that alter these properties may also influence the ecologic structure of the entire community. The style and extent of biologic modification of the substrate will be preserved as biogenic sedimentary structures.

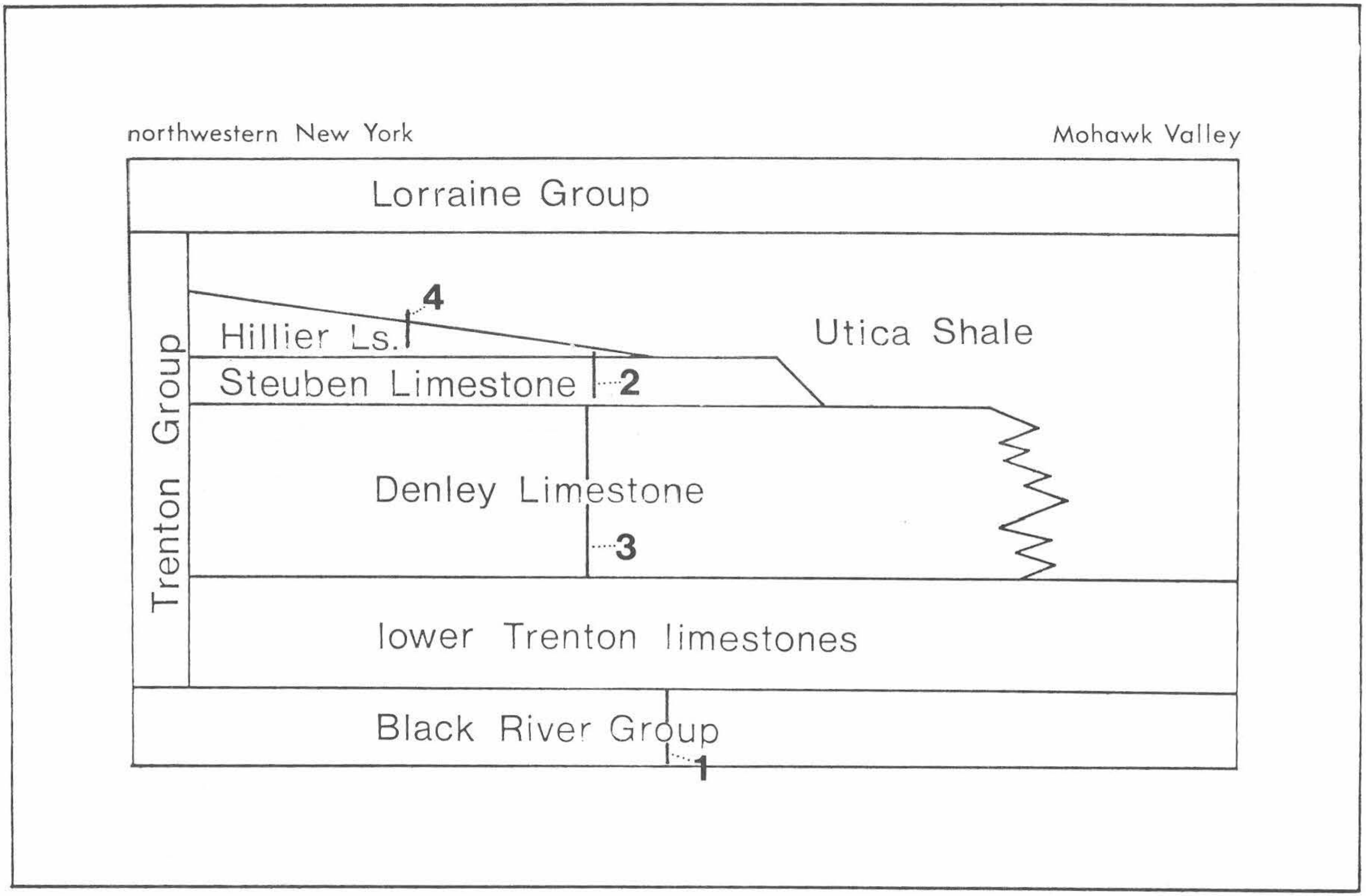

Figure 1. Stratigraphic cross-section of Ordovician rocks in the Black River Valley. Stops for this field trip are indicated by a vertical line and appropriate numeral. Modified from Fisher (1977).

Biogenic structures are produced by a wide variety of activities that include construction of dwellings, crawling and grazing, escape activity, or deposit-feeding (see Osgood, 1970; Schafer, 1972). Crawling and grazing trails, except in regions of very low sedimentation rate, are unlikely to rework large volumes of sediment. Likewise, escape and dwelling traces are discrete structures that ordinarily are not responsible for large-scale reworking of the substrate. Deposit-feeding organisms do extensively rework the sediments on the bottom. In the process of extracting their food, many deposit-feeders produce a layer of pelletized sediment with reduced shear strength and high water-content (review in Rhoads and Boyer, 1982). This sediment is readily resuspended by water motion and may contaminate the feeding and respiratory structures of hapless suspension-feeders. Sediment destabilization by deposit-feeders can also disorient and bury stationary taxa as well as prevent recruitment of juveniles. Such functional-group amensalism (Rhoads and Young, 1970; Brenchley, 1982) is responsible for excluding suspension-feeding taxa from substrates inhabited by errant deposit-feeders. Importantly, we may be able to identify such thoroughly bioturbated sediments on the basis of diffuse, poorly defined burrows produced in high water-content sediments (Rhoads and Young, 1970; Rhoads and Boyer, 1982).

How can we evaluate the degree of biogenic reworking that has affected a sedimentary rock? In a comparison of bioturbation in Ordovician and Devonian rocks, Larson and Rhoads (1983) used the following criteria:

1. Morphology of individual traces
2. Sedimentary fabric of the rock
3. Thickness of preserved bedding units

All three points will be useful in the examination of bioturbation in Ordovician sedimentary environments. The importance and significance of trace fossil morphology needs little elaboration. The size and orientation, that is parallel, sub-parallel, or perpendicular to beddin , of a biogenic structure are important in determining the extent of sedi ment reworking.

The use of the sedimentary fabric of a rock as a guide to bioturbation depends on distinguishing between fabric elements due to physical sedimentary processes and those imposed biogenically. Lamination, grading of grain size, and orientation of elongate grains parallel to bedding are all due to physical processes; these may be disrupted by burrowing. Oddly however, some of the most conspicuous burrows occur in rocks that are not extensively reworked. For example, Figure 2 is a view perpendicular to bedding of a well laminated rock with a prominent series of vertical tubes. Less than 20% of the rock volume has been reworked by organisms. Similarly, burrow networks that appear on bedding planes may actually cause little reworking of the sediment.

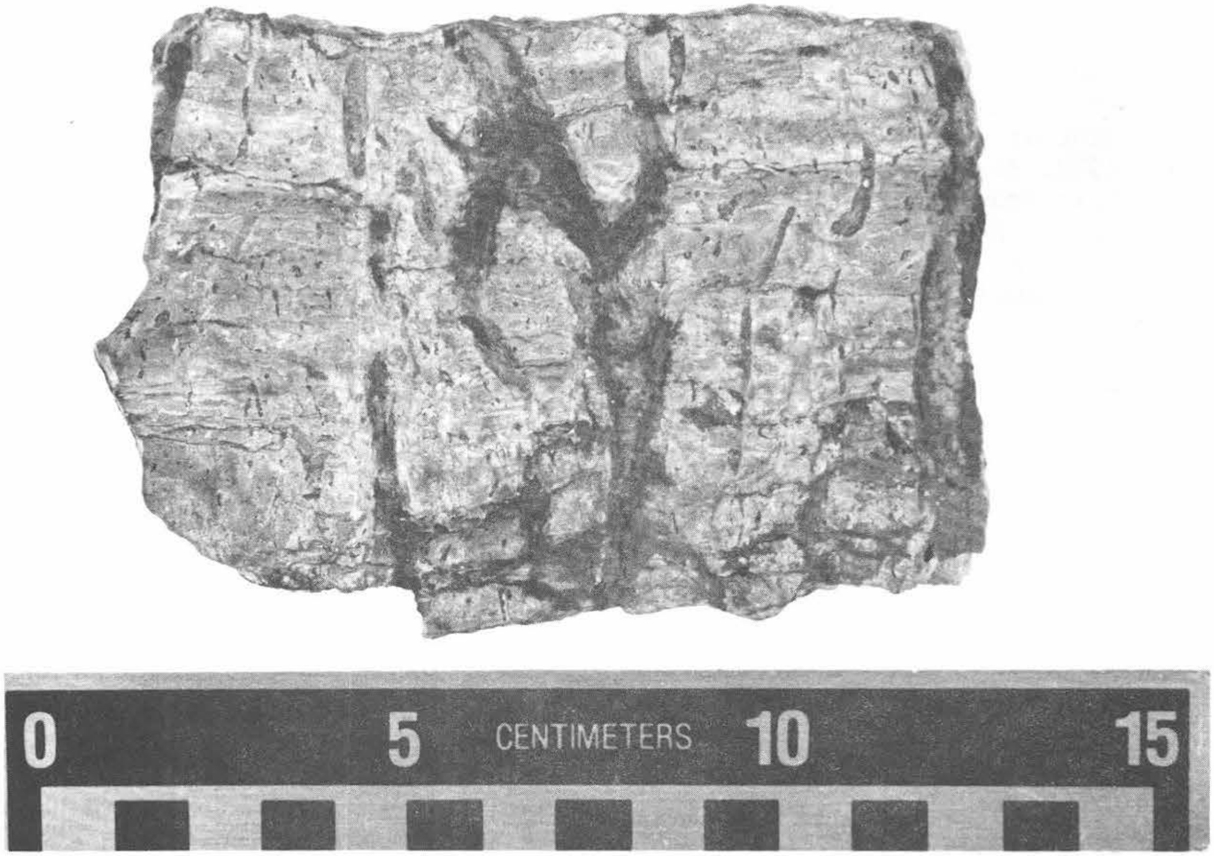

Figure 2. Vertical burrow in a finely laminated carbonate mudstone. Lowville Formation, Black River Group, from Ingham Mills, New York.

Thickness of preserved bedding units may be useful in determining the depth of bioturbation. Sedimentary units thinner than the average depth of reworking are unlikely to be preserved as distinct layers in the sedimentary column. Thus, the absence of beds thinner than 5 cm thick may indicate a depth of reworking of 5 cm. Comparison of bedding unit thickness is best done on a within-habitat basis to ensure that differences in sedimentary regime are not responsible for major differences in bed thickness. Since our traverse today takes us across habitat boundaries, we should use bedding thickness only as a guide to degree of bioturbation.

Sedimentary Fabrics and Environments

Mobile infauna greatly influence the sedimentary record of Recent shelf environments. In nearshore settings, physical structures dominate under the influence of high sedimentation rates and wave and current processes. Offshore in less turbulent waters, physical structures are replaced by burrows. Sediments lying in water below the depth of storm wave base are thoroughly bioturbated (Moore and Scruton, 1957; Howard and Reineck, 1972; 1981). It is important to note that it is the rapidity of physical reworking in the nearshore and burrowing in the offshore that produces the dominant sedimentary fabric. The increase in bioturbation in the offshore direction is a feature of Recent shelves dominated by detrital clastics (see Howard and Reineck, 1972; 1981) as well as carbonates (Ginsburg and James, 1974; James and Ginsburg, 1979). Figure 4 includes a general representation of the distribution of physical and biogenic sedimentary structures in Recent shelf environments.

How does sediment reworking in Ordovician habitats compare to the Recent? Figure 4 also includes my interpretation of the relative importance of biogenic reworking of sediment in the Ordovician carbonates of the Black River Valley; we will be examining the field evidence for this interpretation. Unlike Recent shelves, maximum reworking occurs nearshore in wave-influenced waters. There is no trend of increasing bioturbation in an offshore direction--sediments at and below storm wave base do not show evidence of extensive reworking. What is the style of substrate utilization in this habitat? The Denley Limestone at Stop 3 is an excellent locale to consider this point.

The Denley Limestone contains graded packstones interbedded with carbonate mudstones and shaly partings. (Figure 4). Following the criteria of Kreisa (1981), I have interpreted these as storm deposits. The alternation of turbulent and quiet water builds many fine scale bedding units into the sedimentary column. As seen in Figure 4, the effect of burrowing in reworking these deposits has been minor. The burrows that are present are either restricted to upper bedding surfaces, penetrate less than a few centimeters, or fail to rework large volumes of sediment. Unlike Recent sediments accumulating in similar conditions, the Ordovician material is not reworked by deposit-feeders and retains its physical sedimentary fabric.

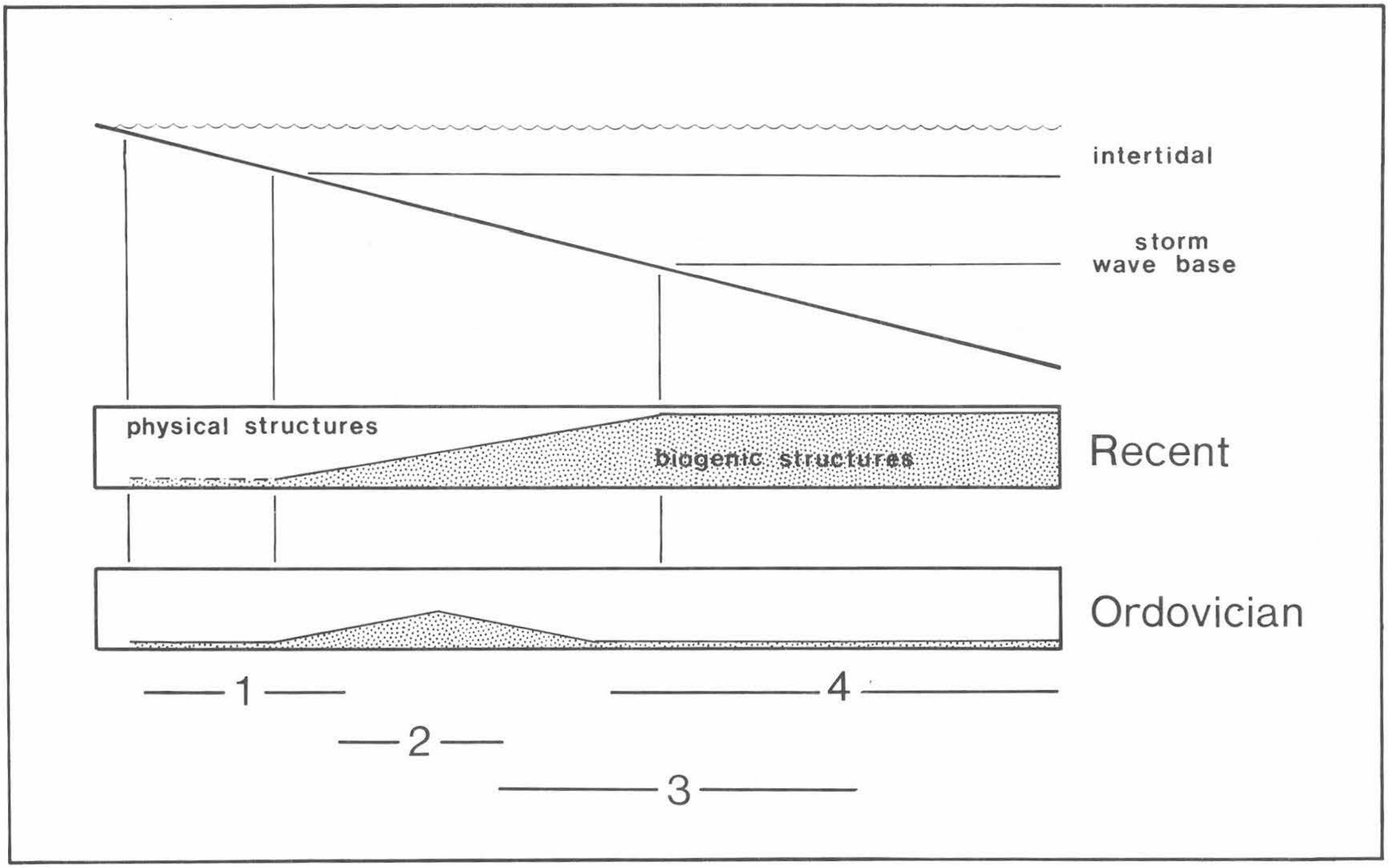

Figure 3. Comparison of the effects of burrowing along an onshore-offshore gradient. Recent near-shore settings are underlain by sediments bearing physical sedimentary structures. Biogenic structures increase in importance in the offshore direction. Below storm wave base very few physical structures are preserved. Ordovician carbonates of the Black River Valley show a different pattern. Shallow subtidal environments show the greatest degree of biogenic reworking. Deeper shelf settings show only minor bioturbation. The lines and numerals beneath the Ordovician bar indicate the environmental range at the field trip stops.

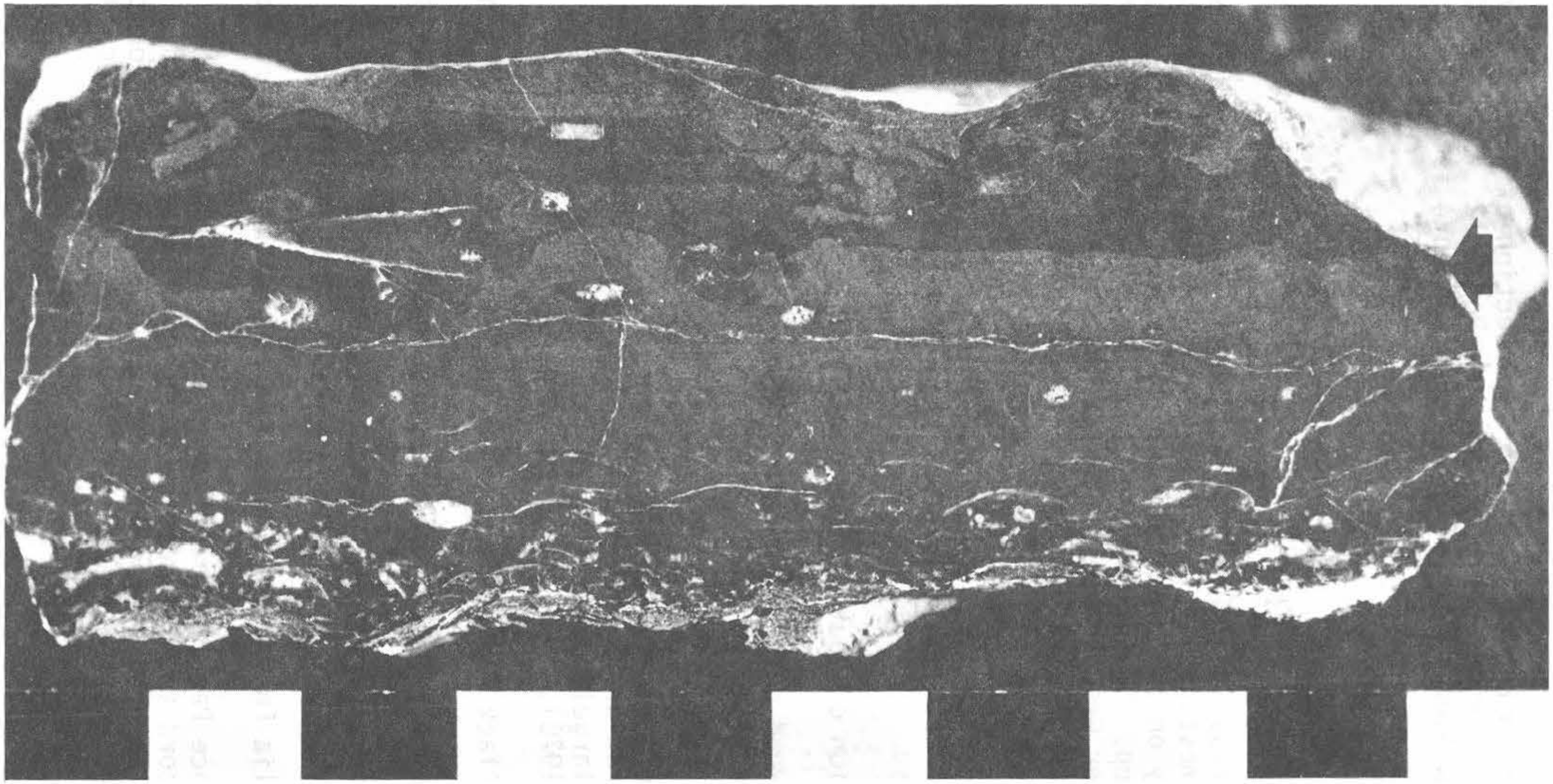

Figure 4. Sample of the Denley Limestone from exposure along Roaring Brook, Stop 3. Burrowing at the contact (arrows) occurred before deposition of the upper unit. This surface marks a former sediment-water interface at which burrows penetrated less than 1 cm into substrate. The lower unit is a storm deposit that grades from a packstone with fragmented, imbrivated fossils to a laminated mudstone. Centimeter scale.

ROAD LOG AND STOP DESCRIPTIONS

Road Log begins in Boonville at the intersection of Routes 12 and 12D. This point is 32 miles north of Utica on Route 12. Stops 1, 2, and 3 are located on the Glenfield 7.5' quadrangle. Stop 4 is on the Rodman 7.5' quadrangle.

Mileage

0.0 Intersection of Routes 12 and 12D. Proceed north on Route 12.

3.3 Bridge across Sugar River. Upstream is an excellent exposure of the limestones in the lower portion of the Trenton Group. The quarry on the right side of the highway is in the Black River Group. Downstream, solution cavities and channels in the Black River Group cause subsurface drainage of Sugar River. At low discharge all of Sugar River disappears into the streambed.

Our route north parallels the channel of the Black River and runs at or near the contact of the Black River Group on the Precambrian.

11.0 Intersection of Route 12 and Turin Road. Turn left.

11.1 Stop 1. Road cuts in the Black River Group on north and south side of Turin Road.

The Black River Group is a well documented example of tidally influenced carbonate deposition. This discussion of the stratigraphy and sedimentology of the Black River Group is largely drawn from Walker (1973) and Walker and Laporte (1970).

Here the Black River Group is divided into three formations:

Chaumont Formation: burrowed, fossiliferous wackestone.

Lowville Formation: fenestral, laminated mudstone, wackestone, and packstone.

Pamelia Formation: dolostone and dolomitic sandstone.

The sequence Pamelia/Lowville/Chaumont is interpreted by Walker (1973) to record the progressive transgression of the Middle Ordovician sea onto an eroded Grenville terrane. From base to top the Black River Group records the transition from supratidal to intertidal to shallow subtidal conditions.

The Lowville Formation includes a number of sedimentologic features indicative of an intertidal origin: mudcracks, fenestral fabric, and algal laminations. In addition, oolites, intraformational conglomerates, and fragmented mounds of *Tetradium* indicate vigorous stirring of the bottom.

The Chaumont Formation contains fewer sedimentologic criteria on which to base an environmental interpretation. Walker (1973) bases his assignment of the Chaumont to the shallow subtidal on the presence of brachiopods and bryozoa in the Chaumont and the interbedding of Lowville and Chaumont lithologies.

A striking feature of the outcrop is the contrast in sedimentary fabrics between the Lowville and Chaumont Formations. Burrows are rare in the Lowville. The sedimentary structures that permit such a straight-forward facies assignment are barely altered by biogenic reworking. Chaumont sedimentary fabrics on the other hand are dominated by burrows. In many cases the outlines of individual burrows are distinct; in others, the burrow outlines are diffuse. Importantly, other than laterally discontinuous bedding units, the biogenic structures have very nearly obliterated the depositional features. This style of bioturbation is characteristic of deposit-feeding communities (Rhoads and Young, 1970; also Rhoads and Boyer, 1982).

Here then we are able to see that for these Ordovician habitats, like their recent counterparts, the effects of bioturbation increase in an offshore direction. A deposit-feeding community was clearly active in this *shallow* subtidal setting.

Mileage

	Continue uphill on Turin Road.
12.0	Intersection with East Road. Bear right onto East Road.
12.3	South Lewis High School on right.
14.2	T-intersection, continue on East Road.
16.1	T-intersection, continue on East Road.
16.5	T-intersection. Turn right onto Houseville Road.
16.8	Stop 2. Abandoned railroad cut through the Steuben Limestone of the Trenton Group. From this vantage point you can see the Tug Hill Plateau, underlain by the Utica Shale and Lorraine Group, to the west. To the east across the Black River Valley are Grenville rocks of the Adirondacks.

At this exposure we will examine a ten meter section of the Steuben Limestone (Figure 6). About 40 cm of Hillier Limestone is exposed at the top of this cut. We will see a much thicker section of the Hillier at Stop 4.

The Steuben Limestone is generally a thickly bedded fossiliferous packstone with some important variations in lithology. Near the base of the exposure, fossiliferous mudstones and wackestones are interbedded with centimeter thick argillaceous mudstones. Both grainsize and bed thickness increases upward through the Steuben. Midway through the section are cross-laminated grainstones and packstones, some with mega-rippled upper bedding surfaces. There are shaly partings near the top of the Steuben, but no interbedded lime or argillaceous mudstones. The upper portions of the Steuben accumulated in more turbulent, more frequently agitated waters than the sediments at the base of the exposure. On the onshore-offshore gradient, I assign the Steuben to an open marine shelf subject to occasional stirring by currents or waves (Figure 3). Certainly the Steuben was deposited in a more exposed environment than the Chaumont of Stop 1.

Trace fossils are prominent throughout the Steuben. Near the base of the exposure are examples of Palaeophycus, Planolites, and Chondrites. These burrows are confined to bedding surfaces or generally penetrate the sediment only several cm. Monocraterion, a vertical sediment filled tube one cm in diameter and up to 8 cm long, is common in the crinoidal packstones near the top of the Steuben.

The style of bioturbation also varies vertically in the Steuben. Although nowhere is the Steuben completely burrow reworked, it seems that the greatest degree of biogenic alteration of the sedimentary fabric occurs in the upper Steuben with Monocraterion. This trend runs counter what we would expect from the modern: rather than finding the greatest degree of biogenic reworking in fine-grained sediments at depth, here the coarser-grained, shallow-water environments are more reworked. Consequently, at this point Figure 3 shows the divergence in bioturbation trends for the Recent and Ordovician.

Mileage	Continue on Houseville Road, heading downhill.
18.1	Intersection with Duncan Road. Turn left.
19.5	Intersection with Lee Road. Turn right.
19.8	Intersection with Glendale Road. Turn left.

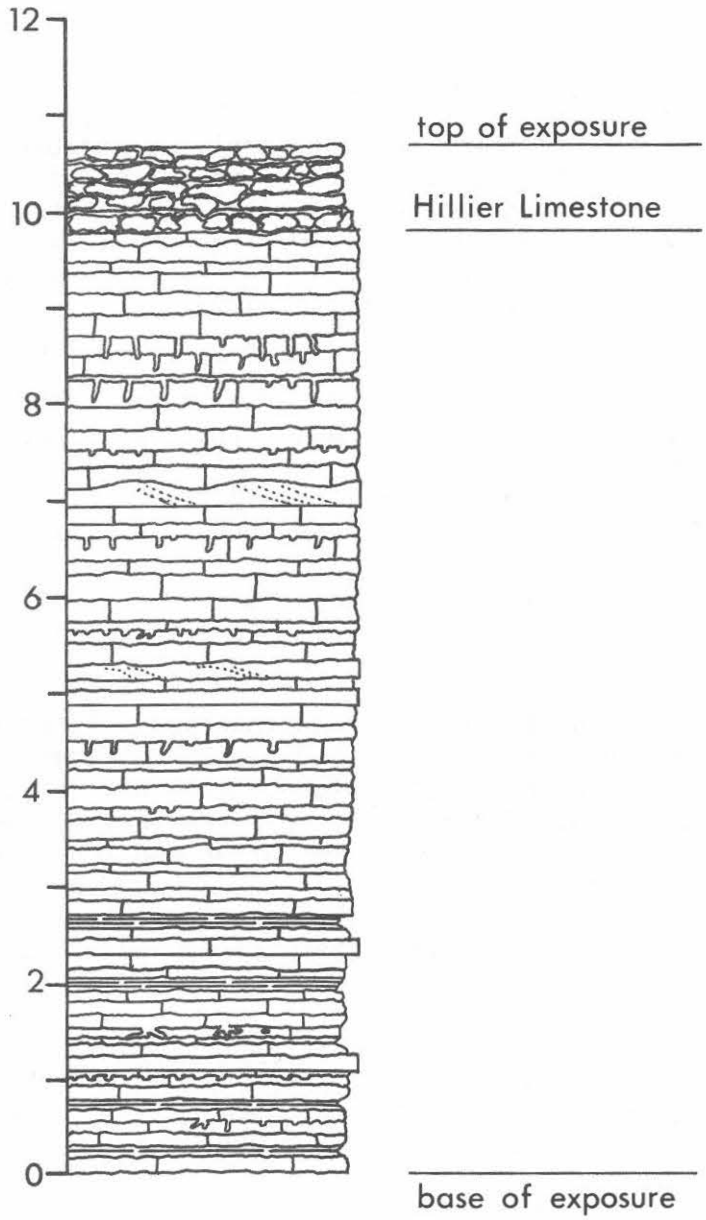

Figure 6. Stratigraphic column of Steuben Limestone at Stop 2. Scale on left side is in meters.

Mileage

20.7 Bridge over Roaring Brook. Denley Limestone of Trenton Group in creek bed.

20.9 Stop. 3. Entrance to Whittaker Falls Park. Turn right.

Nearly all of the Trenton and Black River Groups are exposed along Roaring Brook. In a combination of rapids, falls, and level stretches, the rocks are exposed in both vertical section and on bedding planes. A long, nearly continuous exposure of the Denley Limestone is the focus of our attention.

On the basis of smaller grain size and fewer indications of turbulent conditions, I have assigned the Denley to a position further offshore than the overlying Steuben Limestone. The upper 50 m of the Denley contains mega-rippled and cross-stratified grainstones interbedded with finer-grained lithologies. This portion of the Denley was deposited within the reach of storm wave base. Because the basal 9 m of the Denley lack these grainstones, this portion of the formation appears to have accumulated below storm wave base.

The Denley Limestone contains a diverse fauna including brachiopods, bryozoa, crinoids, and trilobites. Many of the body fossils apparent on outcrop here are fragmented, abraded, and occur in grainstone or packstone units. Overturned heads of the bryozoan Prasopora attest to disturbance and relocation of many of the fossils. Figure 7 is an illustration of the role of turbulent events, major storms, in producing both the sedimentary structures and the fossil assemblages in the Denley.

Not all of the fossils are reworked, however. Adhering to the tops of limestone beds, or entombed within centimeter thick shaly partings are some fossil assemblages that indicate in place accumulation. Evidence for in place accumulation includes lack of abrasion and fragmentation and the co-occurrence of fossils ranging in size from .1 to 1.5 cm. In addition, several specimens of juvenile crinoids, complete with holdfast, suggest in place burial rather than transportation before burial.

Trace fossils are abundant here. Palaeophycus and Chondrites are the most conspicuous. Again, despite the presence of burrows, the sedimentary fabrics of the Denley retain their original features. Maximum depth of burrowing is about 3 cm.

Return to park entrance and turn right onto Glendale Road.

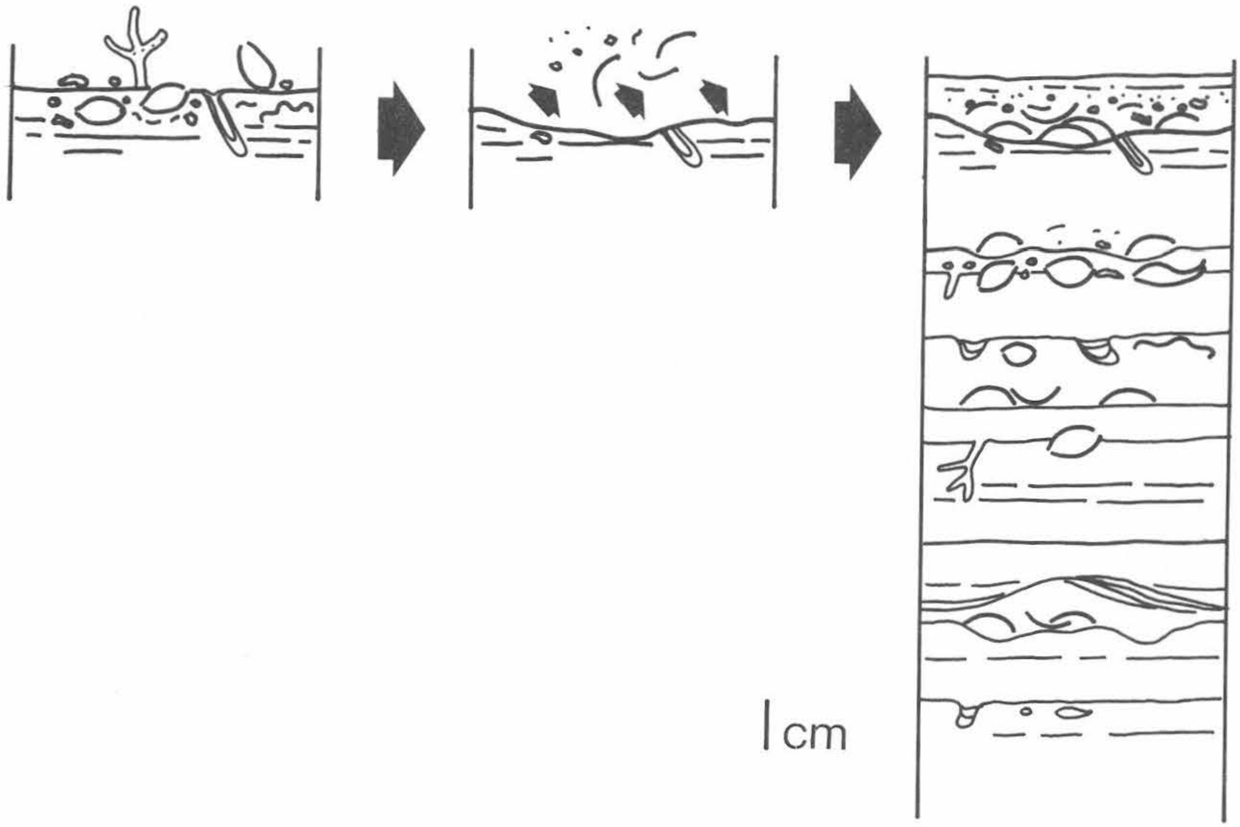

Figure 7. Forming a storm deposit in the Denley Limestone. Turbulence of a storm suspends sediment and disarticulates fossils. Commonly, a graded fossiliferous packstone is formed as sediments settle from suspension, illustrated in Figure 4. Depending on conditions, a storm may winnow sufficient sediment to produce a mega-rippled grainstone or may bury a fossil assemblage in place.

Mileage

30.0	Y-intersection, bear left.
30.6	Intersection with Route 26. Turn right into village of Martinsburg.
31.4	Crummy roadcut on left is Steuben Limestone.
33.9	Entering Lowville. Routes 12 and 26 join. Continue straight ahead on 12 and 26.
34.5	Downtown Lowville. Turn left on Route 12. Time and temperature on bank on right side of road.
35.2	Bridge over Mill Creek. Excellent exposures of Trenton Group.
37.2	West Lowville. Junction with Route 177. Bear left onto 177. We are climbing to the top of the Tug Hill Plateau.
48.0	Crossing Deer River at New Boston. Continue straight.
51.8	Barnes Corners. Continue straight.
59.0	Village of Rodman. Turn right.
59.2	Right turn onto Creek Road.
59.6	Bridge over Gulf Stream.
59.7	Stop. 4. Park off of road on left-hand side. Cut along Gulf Stream where we can examine the contact between the Utica Shale and the underlying Hillier Limestone of the Trenton Group.

Watch out! Poison ivy is abundant and lush here, especially on the Utica Shale.

This is a 13 m section that records the transition from wave influenced shelf to deep, anaerobic basin (Figure 8). Illustrating this trend through the Hillier Limestone is a decrease in grainsize, loss of mega-rippled and cross-laminated beds, and an increase in number and thickness of shaly partings in the limestone.

Here the transition to deeper water is marked by an increase in burrowing. Palaeophycus, Planolites, and Chondrites are present, but evidence of a bioturbating, deposit-feeding

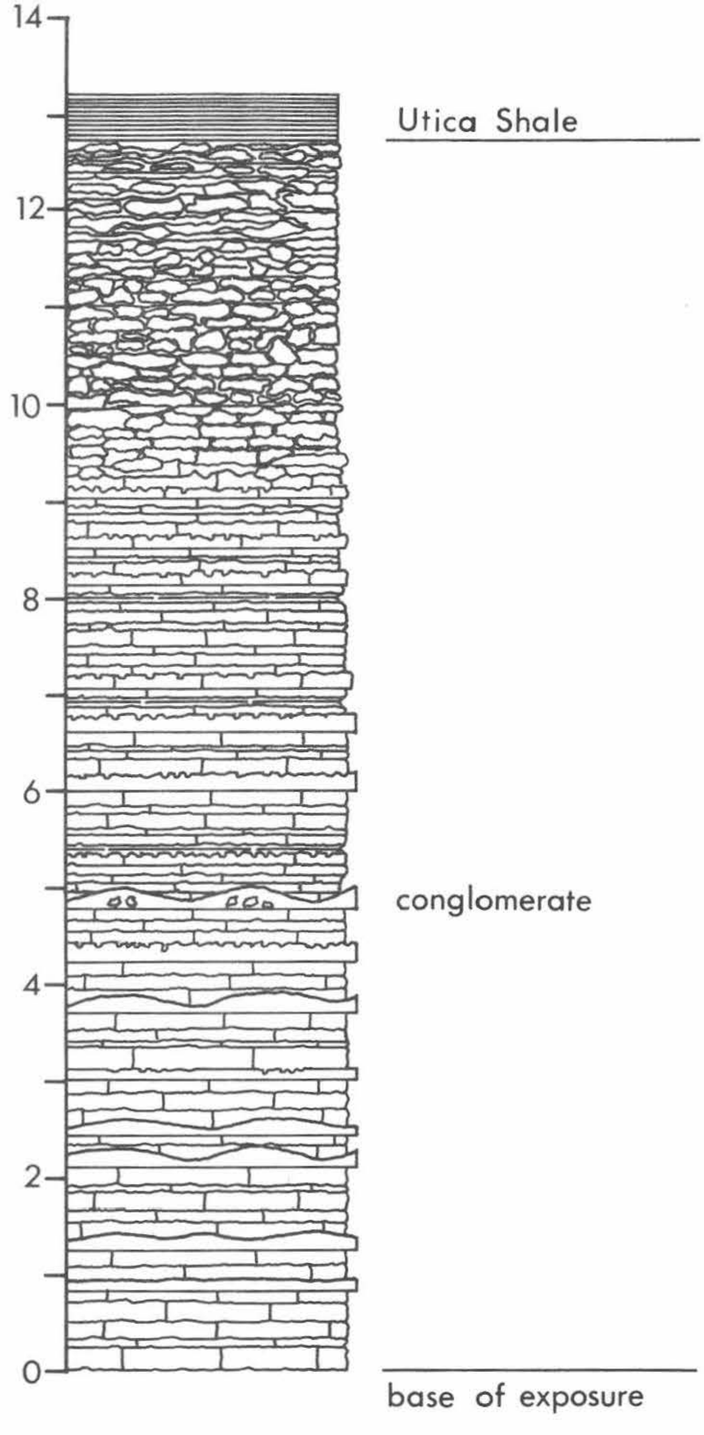

Figure 8. Stratigraphic column of Hillier Limestone at Gulf Stream, Stop 4. Scale on left is in meters.

community is absent. Here, too, the conspicuous body fossils are brachiopods, bryozoa, and crinoids although the gastropod Liospira is locally abundant.

The uppermost Hillier is an interesting lithology with lumpy nodular bedding. The fauna here includes mostly phosphatic forms: trilobites, conularids, and lingulids with setae preserved around the margin of the valves.

The Utica Shale is a black, fissile, argillaceous mudstone. Careful collecting can turn up cephalopods, graptolites, and the trilobite Triarthrus.

End of trip. Reverse direction to return to Clinton. Follow 177 to junction with Route 12. Follow Route 12 to Utica and New York State Thruway. Follow 12B from Utica to Clinton.

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Sedimentary Structures and Paleoenvironmental Analysis of the Bertie Formation (Upper Silurian, Cayugan Series) of Central New York State

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INTRODUCTION

The Bertie Formation (Upper Silurian, Cayugan Series) of New York State, is world renowned for its spectacular eurypterid fossils. However, details of the total fauna (not just the eurypterids) and its significance, along with an understanding of the depositional environment, lags behind most other rock units in New York State. An understanding of the depositional environment of the Bertie is further complicated by a lack of published studies on the sedimentary structures that occur in the formation. Almost without exception, the fossils and the sedimentary structures have been treated as separate and totally unrelated aspects. Only minor attention had been paid to the structures (e.g., mud-cracks, and the windrow accumulations).

The writers are fully aware that the fossils are an important aspect in the overall interpretation of the depositional environment of the Bertie, especially when they are considered as primary sedimentary structures. Matters of identification and errors of stratigraphic relations of the fossils in the Bertie strata severely limit their usefulness and delay understanding of their significance (see further comments under Paleontology).

With regard to the sedimentary structures, we know of no published studies that list, describe, and discuss all of them. Buchwald (1963) did study some of them. The main purpose of this paper, then, is analyses of all of the known sedimentary structures of the Bertie, with emphasis on their use in the interpretation of the depositional environment.

Although only three stops will be made in the field, all are key outcrops. Stop 1 is at Forge Hollow, New York, where a complete sequence of the Bertie with both lower and upper contacts with the Camillus and Cobleskill Formations, respectively, are exposed. Stop 2 is at Jerusalem Hill, where the upper Fiddlers Green and the lower Scajaquada Members are exposed. Stop 3 is at Passage Gulf, where the lower contact with the Camillus is exposed.

ACKNOWLEDGMENTS

We thank Mr. William Parker of the Utica College Audio-Visual Department for taking the photographs. The senior writer thanks his wife, Mary, for her patience and understanding.

HISTORY

The Bertie Formation was named by Chapman (1864, p. 190-191) for approximately 50 feet of "...thin-bedded grayish dolomites, interstratified towards the base with a few brownish shales, and with a brecciated bed composed chiefly of dolomitic fragments" that were exposed near the Township of Bertie, Ontario, Canada. The name was first used in New York State by Schuchert (1903, p. 171-172) in his study of the Manlius Formation. Subdivision of the Bertie in its historical perspective is shown in Table 1. A more complete historical perspective is found in Rickard (1953; 1962). The terminology used in this paper follows that of Rickard (1975); Correlation with the terminology of Ciurca (1973) is included in the discussions.

Stratigraphically, the Bertie is the uppermost formation of the Salina Group (Cayugan Series). The New York State Geological Survey (1970) has mapped the Bertie either as undifferentiated from the Salina Group (eastern map sheets), or with the Camillus and Cobleskill Formations (western map sheets). The only continuous cross-section of the Bertie for New York State is found in Rickard's Masters thesis (1953).

Historically, the Bertie Formation has received sporadic attention, mainly because of its prolific and spectacular eurypterid fauna. One of the main purposed of many older studies had been to reconcile the habitat of the eurypterids with the occurrence of other fossils, and the presence and/or absence of some of the sedimentary structures (e.g., mudcracks) into a coherent environmental interpretation.

According to Alling (1928, p. 42, 54), O'Connell (1916) interpreted the Bertie as a deposit of clastic origins. She interpreted the Bertie as having been either 1) a flood-plain deposit; 2) a deltaic deposit; or 3) a playa-lake deposit. Her conclusions on the habitat of the eurypterids were that they lived in rivers, because few other fossil groups were rarely, if ever, found in the same beds as the eurypterids. Also, most of the eurypterids found were disarticulated exoskeletons.

Ruedemann (1916c; 1925) disputed O'Connell's work. He concluded eurypterids were truly marine organisms, in that the entire fauna was of truly marine origins. The disarticulated nature of the eurypterids was, according to Ruedemann (1916c, p. 114) a natural phenemnon for chelicerate arthropods in a near-shore setting. Ruedemann's (1925, p. 12-13) conclusions as to the environment of the Bertie, were that it was a lagoon deposit, formed behind (north of) coral reefs (located further south). According to Leutze (1959, p. 99) this interpretation by Ruedemann (that the Bertie formed behind coral reefs) was based upon a misinterpretation of the stratigraphy in the Syracuse, N.Y. area. With the exception of the depositional strike, Ruedemann's views are currently the most widely accepted (Alling and Briggs, 1961; Rickard, 1962; Treesch, 1972).

A small number of workers are currently re-examining the Bertie. Ciurca (1973, 1978, 1982) has contributed much toward an understanding of the stratigraphy with regard to the eurypterids. An up-dated review of the stratigraphy and sedimentology of the entire Salina Group was done by Treesch (1972).

TABLE 1. Historical review of the terminology of the Bertie Formation.

Vanuxem (1842)	Hall (1843)	Chapman (1864)	Schuchert (1903)	Hartnagel (1903)	Hopkins (1914)
Fourth or Magnesian, and Third or Gypseous Deposits of the Onondaga Salt Group	Fourth or Magnesian or Waterlime or Hydraulic Deposit of the Onondaga Salt Group	Bertie Formation	Bertie Formation	Eurypterus-bearing Salina beds	Bertie Formation
					Camillus Shale
					Fiddlers Green Member, of the Camillus Shale

Chadwick (1917)	Rickard (1962) Western N.Y.	Rickard (1962) Central N.Y.	Rickard (1975)	Ciurca (1978) Rochester, N.Y.
Buffalo Member, Bertie Formation	Williamsville Member, Bertie Formation	Oxbow Member, Bertie Formation	Williamsville Member, Bertie Formation	Williamsville Formation, Bertie Group
Scajaquada Member, Bertie Formation	Scajaquada Member, Bertie Formation	Forge Hollow Member, Bertie Formation	Scajaquada Member, Bertie Formation	Scajaquada Formation, Bertie Group
Falkirk Member, Bertie Formation	Falkirk Member, Bertie Formation	Fiddlers Green Member, Bertie Formation	Fiddlers Green Member, Bertie Formation	Phelps Mbr. Victor Mbr. Morganville Mbr. all of Fiddlers Green Fm., Bertie Group
Oatka Member, Bertie Formation	Oatka (?) Member, Bertie Formation	Not recognized		
				Oatka Formation

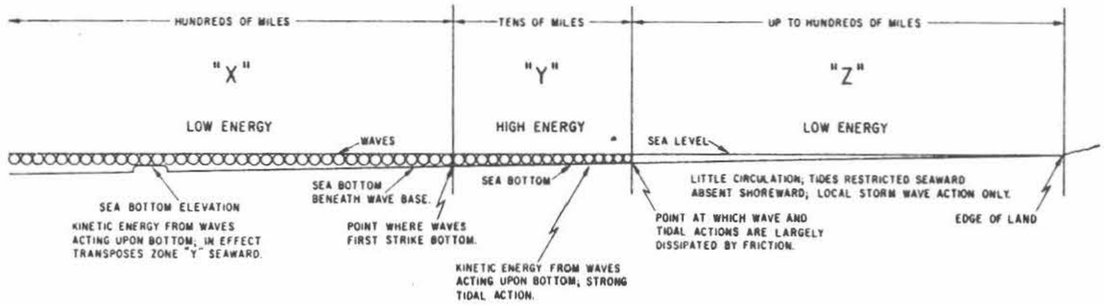

Figure 1. Generalized section showing the energy zones in epeiric seas. Not to scale. From: Irwin (1965, fig. 3, p. 450).

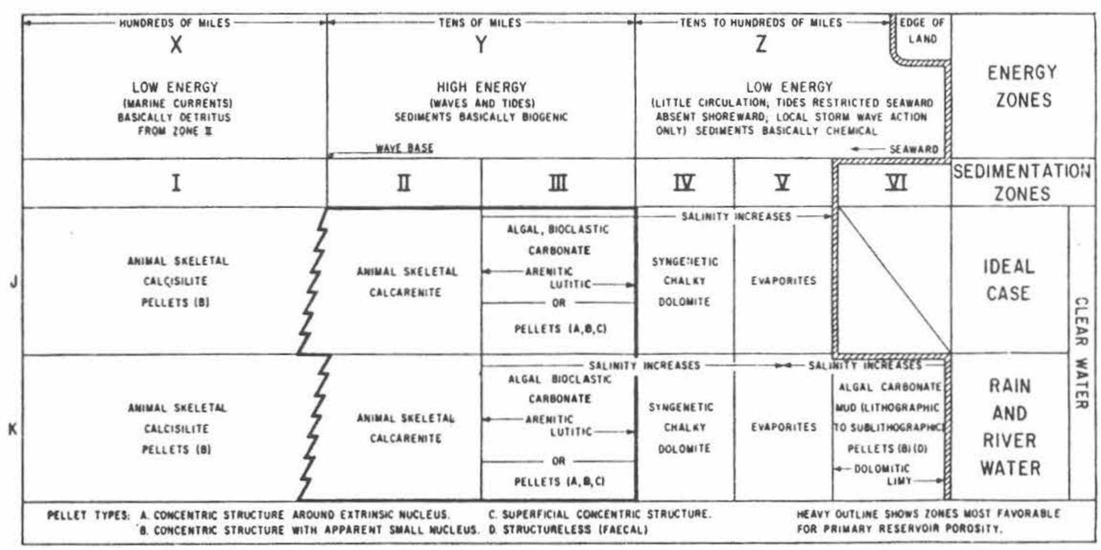

Figure 2. Energy and sedimentation zones of epeiric seas. From: Irwin (1965, fig. 5, p. 452).

Recently, attempts have been made to apply the epeiric sea model of Irwin (1965) (Figs. 1 and 2) to the Bertie (Belak, 1980; Hamell, 1981; Tollerton, 1983). Additional field work is necessary before this model is fully applicable to the Bertie. In this respect, the present paper adds to what is currently known about the depositional environment of the Bertie. Much remains to be learned. Problems that remain are: 1)Has the Bertie been subjected to compaction? If so, how much, and what kind of compaction mechanism produced the features observed? 2)How were the shrinkage cracks formed? Several different types of shrinkage cracks are reported in the literature, but precise descriptions for their recognition are lacking. This problem pertains not only to the Bertie, but also to other sedimentary deposits that contain shrinkage cracks (e.g., the Green River Formation, a lake deposit. See Smoot, 1983, for a discussion on shrinkage cracks). 3)Are the reported isolated occurrences of fossils and sedimentary structures real or apparent? Have we been too enamored with the eurypterids to the point of indifference? Have we misidentified many fossils; given them new names based on size and stratigraphic position (e.g., comments by Leutze, 1959); or collected the fossils from the wrong stratigraphic unit (e.g., Syracuse Fm. and Cobleskill Fm.) referring them to the Bertie based on lithologic similarity? 4)How saline was the environment? Can the water chemistry be determined or inferred? 5)How deep was the water? 6)Do the fossils really represent a life assemblage throughout, or only for certain beds? 7)Was dolomitization primary or secondary? Has dolomitization obscured the details of the depositional environment - structures as well as the fauna? Some of the possibilities, along with preliminary results of study and suggestions for further study, are presented throughout this paper.

The results of our work in assigning Irwin's (1965) epeiric sea model to the Bertie are summarized in Table 6. In the tabulation, the previously proposed models and the epeiric sea model are compared with respect to the expected occurrences of the fossil groups, sedimentary structures, and the interpretations furnished from their study.

PALEONTOLOGY

With the exception of the eurypterids, the fossils found in the Bertie Formation are generally poorly preserved, and in need of extensive study. A preliminary, provisional, revised faunal list is given in Table 2. An extensive discussion of the revision is beyond the scope of the present paper, as are comparisions with the lists in Ruedemann (1925) and Monahan (1931). However, a few remarks are warranted with regard to their paleoecological implications.

Several writers (Clarke and Ruedemann, 1912; Ruedemann, 1925; Alling, 1928; Leutze, 1959, 1964; Alling and Briggs, 1961; Treesch, 1972; Ciurca, 1973, 1978, 1982) have remarked that most of the fossils reported as occurring in the Bertie are restricted to single outcrops, with only a small number of species occurring at more than one outcrop. The question then, is why would the organisms be so different and so restricted throughout the Bertie? This restriction may be apparent, and due in part to three factors. First, the past practices of naming species based upon differences in size and stratigraphic position (comments by Leutze, 1959,

TABLE 2. Provisionally revised faunal list of the Bertie Formation. Sources given in parentheses.

Coelentrates
- Conularids (Ruedemann, 1925; Monahan, 1931)
  - *Metaconularia perglabra* (Ruedemann)
- Corals (Ruedemann, 1925)
  - **Aulocystis* sp.
  - *Stromatopora* sp.

Bryozoans (Ruedemann, 1925; Monahan, 1931)
- *Hederella* cf. *canadensis* (Nicholson)
- *H.* sp.
- **Stigmatella* sp.

Brachiopods (Ruedemann, 1916a, 1925; Monahan, 1931; Leutze, 1959; Berdan, 1972; Ciurca, 1982)
- *Eccentricosta jerseyensis* (Weller)
- *Howellella eriensis* (Grabau)
- *Lingula semina* Ruedemann
- *L. subtrigona* Ruedemann
- *Morinorhynchus ? interstriatus* (Hall)
- *Orbiculoidea* cf. *numulus* Hall and Clarke
- *Protathryris sulcata* (Vanuxem)

Mollusca
- Gastropods (Ruedemann, 1916a, 1925; Leutze, 1959)
  - **Loxonema bertiensis* Ruedemann
  - **Murchisonia* (*Hormotoma*) *gregaria* Ruedemann
  - *Platyceras* (*Platyostoma*) sp.
- Cephalopods (Ruedemann, 1916a, 1925; Monahan, 1931; Flower, 1948; Leutze, 1959)
  - *Dawsonoceras oconnellae* Ruedemann
  - *Mitroceras gebhardi* (Hall)
  - *"*Orthoceras*" *vicinus* Ruedemann
  - *Pristeroceras timidum* Ruedemann
- Pelecypods (Ruedemann, 1925; Monahan, 1931; Leutze, 1959)
  - *Goniophora* (?) sp.
  - *Hercynella buffaloensis* O'Connell
  - *Nuculites salinensis* (Ruedemann)

Worms (Ruedemann, 1925)
- *Ruedemannella obesa* (Ruedemann)
- *Spirorbis* sp.

Arthropods
- Scorpions (Ruedemann, 1925; Kjellesvig-Waering, 1966)
  - *Archaeophonus eurypteroides* Kjellesvig-Waering
  - *Proscorpius osborni* (Whitfield)
- Eurypterids (Clarke and Ruedemann, 1912; Ruedemann, 1925; Kjellesvig-Waering, 1958, 1963, 1964; Ciurca, 1982)
  - +*Acanthoeurypterus dekayi* (Hall)
  - +*A. wellsi* Kjellesvig-Waering
  - +*Alloeurypterus linsleyi* Kjellesvig-Waering
  - *Buffalopterus pustulosus* (Hall)
  - *Clarkeipterus testudineus* (Clarke and Ruedemann)
  - *Dolichopterus herkimerensis* Caster and Kjellesvig-Waering
  - *D. Jewetti* Caster and Kjellesvig-Waering
  - *D. macrocheirus* Hall
  - *D. siluriceps* Clarke and Ruedemann

TABLE 2. (Continued)

Arthropods
- Eurypterids (Continued)
  - *Erettopterus (Erettopterus) grandis* (pohlman)
  - *Eurypterus remipes lacustris* Harlan
  - *E. remipes remipes* DeKay
  - *Paracarcinosoma scorpionis* (Grote and Pitt)
  - *Pterygotus (Acutiramus) macrophthalmus cummingsi* (Grote and Pitt)
  - *P. (A.) macrophthalmus macrophthalmus* (Hall)
  - *P. (Pterygotus) cobbi* Hall
  - *P. (P.) juvenis* Clarke and Ruedemann
- Phyllocarids (Ruedemann, 1925)
  - *Ceratiocaris acuminata* Hall
  - *C. aculeata* Hall
  - *C. maccoyana* Hall
  - *C. minuta* (Ruedemann)
- Xiphosurans (Ruedemann, 1925; Ciurca, 1982)
  - *Bunaia woodwardia* Clarke
  - *Limuloides (?) eriensis* (clarke)
  - *Pseudoniscus clarkei* (Ruedemann)
- Ostracods (Ruedemann, 1925; Monahan, 1931)
  - *Eukloedenella umbilicata* Ulrich and Bassler
  - *E.* sp.
  - *Leperditia alta* (Conrad)
  - *L. scalaris* (Jones)
  - *Zygobeyrichia* cf. *regina* Ulrich and Bassler

Echinodermata
- Edrioasteroids (Ruedemann)
  - *Pyrgocystis batheri* Ruedemann
- Machaeridians (Ruedemann, 1925)
  - *Lepidocoleus reinhardi* Ruedemann

Graptolites (Ruedemann, 1925)
- *Climacograptus ultimus* Ruedemann
- *Inocaulis lesquereuxi* (Grote and Pitt)
- *Medusaegraptus gramminiformis* (Pohlman)
- *Orthograptus (?)* sp.
- *Palaeodictyota buffaloensis* Ruedemann

Conodonts (Barnett, 1972)
- *Hindeodella* sp.
- *Ligonodina* sp.
- *Lonchodina greilingi* Walliser
- *L. walliseri* Ziegler
- *Neoprioniodus bicurvatus* (Branson and Mehl)
- *Ozarkodina denckmanni* Ziegler
- *O. media* Walliser
- *Plectospathodus alternatus* Walliser
- *Spathognathodus remscheidensis* Ziegler
- *Trichonodella excavata* (Branson and Mehl)

Algae (Ruedemann, 1925)
- *Callithmnopsis silurica* Ruedemann
- *Morania (?) bertiensis* Ruedemann
- *Sphenophycus (?)* sp.
- LLH Stromatolites

TABLE 2 (Continued)

Vascular (?) Plants (Ruedemann, 1925; Banks, 1972, 1973)
- *Cooksonia hemispherica* Lang
- *C. pertoni* Lang
- *Hostimella silurica* Goldring

Trace Fossils (Ruedemann, 1925)
- Algal wrinkles
- Chrondrites
- Worm (?) trails (?)
- Worm (?) excrementa (?)

* Genus and/or species possibly misidentified. Previous terminology provisionally accepted until original collections can be studied.

\+ Original descriptions not published (?). To date, these names only appear in Ciurca (1982).

p. 92-99, 114, 116-126, 142), in which case several species would then be synonymous. Second, mis-identification of the Bertie at various localities (see Leutze, 1959), in which case, several species reported as occurring in the Bertie would no longer be included in the faunal lists. Third, the poor state of preservation of the fossils, which upon further examination, may be found to be mis-identified, in which case several species would be synonymous.

The question of geographic restriction, however, may not be totally the result of man. Some of the reported isolated and/or patchy occurrences may be related to the harsh and variable hypersaline environment with extremely low-angle slopes. Such occurrences as the graptolite *Medusaegraptus* and the plant *Cooksonia*, both occuring at Passage Gulf, would then be related to such natural factors as storms, persistent yet local tide pools, or transgressive/regressive sea levels. An attempt to develop this idea further would be the tabulation of species and numbers of individuals with geographic and stratigraphic position. Such a study is currently being planned. Although the work is greatly hampered by the lack of adequately preserved fossils and by the abundance of mid-identified (?) fossils, preliminary results indicate that the geographic and stratigraphic restrictions are not as extensive as previously supposed.

Published discussions on the paleoecology of the Bertie fauna have been essentially restricted to the eurypterids, although minor comments have been made in the older papers under remarks in the descriptions of new fossils (e.g., Ruedemann, 1916a, 1916b, 1925; Monahan, 1931). Generalizations have been made on the paleoecology of the entire fauna of the Salina Group (Alling, 1928; Leutze, 1959, 1964; Treesch, 1972), but detailed discussions relative to the Bertie fossils are lacking.

The following comments on the Bertie fossils are only generalizations, in that they have been studied from the aspect of being primary sedimentary structures. Specific details of the fauna are not only beyond the scope of the present paper, but merit more careful attention to detail than we have been able to give them so far. For example, are the eurypterids the only fossils that tend to show a preferred orientation (Ciurca, 1978) ? What is the extent of abrasion and disarticulation of the fossils? Do some of the fossils indicate ecological competition with its resulting restrictions?

The conularids would seem to indicate shallow, marine conditions for the Bertie (Tasch, 1973, p. 140). Too few specimens are known to speculate on anything else.

Poorly described corals reported by Ruedemann (1925) may have been mis-identified and they may be bryozoans. Until the original collections are re-studied, their true categorization and paleoecological significance is unknown.

The presence of the bryozoans would seem to indicate warm, shallow, marine conditions, as the forms reported are small and delicate branching types. The material collected by the senior writer has not been studied in detail at this writing.

Brachiopods of the Bertie would seem to represent the *Lingula* and possibly the *Eocoelia* Communities of Ziegler (1965). However, the complications of identification and stratigraphic misplacement severely limit this idea.

The gastropods, both high- and low-spired forms, indicate a marine environment (Knight, et al, 1960, p. I171). They appear to be more common in the lower parts of the Fiddlers Green and Williamsville Members. A few very small slabs collected from Passage Gulf show a preferred orientation for both forms. Whether this is a relect of the small specimen size, or an actual preferred orientation, is not known at present.

Cephalopods are orthocone and brevicone types. Although their geographic distribution is unknown at present, we believe that they indicate marine conditions for the Fiddlers Green and the Williamsville Members. That they have been washed in is considered unlikely as the material seen thus far does not show signs of either preferred orientation or of abrasion.

Pelecypods are less abundant than the brachiopods, and may have competed with them in the shallow marine environment. However, there are no indications, as yet, that they co-existed during the frequent oscillations of environmental conditions. To date, no specimens have been found by the writers with both pelecypods and brachiopods preserved together in undoubted Bertie strata.

The scorpions from the Bertie may or may not have been washed into the environment. Published information on the morphological characteristics of these animals are inconclusive as to whether thay were marine, brackish-water, or terrestrial invertebrates (Kjellesvig-Waering, 1961, p. 360). However, the total assemblage of fossils occurring with the scorpions and the eurypterids, strongly suggests a quiet, shallow, marine environment.

The Xiphosurans of the Bertie, *Limuloides*, *Bunaia*, and *Pseudoniscus*, are believed to have been benthonic, brackish-water organisms (Størmer, 1955, p. P16). A more detailed paleoecological account is not feasible at present because too few specimens are known, and their relationships to other organisms were not indicated when they were described.

The phyllocarids represent shallow marine, quiet-water conditions (Rolfe, 1969, p. R308). Further comments cannot be made until their relationships to other aspects of the Bertie are known.

The eurypterids fall into the Eurypteridae Phase (Kjellesvig-Waering, 1961, p. 793) of ecological environments, indicating an intermediate zone between normal marine and nonmarine environments. Additional study is necessary to determine if all eurypterid zones in the Bertie display a preferred orientation, or if they represent windrow accumulations, as suggested by Ciurca (1978). We agree with Ruedemann's views (1916c, p. 114) on the disarticulated nature of the eurypterids as being normal for chelicerate arthropods in a shallow, near-shore environment.

The ostracods, especially *Leperditia*, represents shallow, marine, tidal-flat environments (Berdan, 1968). The Leperditids are ubiquitous throughout the Fiddlers Green and the Williamsville Members. Some are found on the exoskeletons of some of the eurypterids, suggesting that they may have been scavengers.

The edrioasteroid, *Pyrgocystis batheri* Ruedemann, is cited by Regnell (1966, p. U157-U158) as having been tolerant of various substrate lithologies, living in soft muds or oozes, and indicating generally quiet, shallow, marine conditions.

Graptolites collected at Passage Gulf are only the carbonaceous thecae of *Medusaegraptus gramminiformis* Ruedemann, not the entire organism. These have either been washed in or disarticulated in situ by currents.

The conodonts indicate a shallow, marine environment (Müller, 1962, p. W87). Their stratigraphic distribution given in Barnett (1972, p. 903) shows that portions of the Fiddlers Green and Williamsville Members were shallow marine. However, only one locality was sampled. Additional collections are required before further comments can be made relative to the environmental conditions of conodonts (Barnett, 1971) for the entire geographic extent of the Bertie, as some of the conodonts may have been washed in.

With the exception of the stromatolites, the algae are poorly preserved. Thin, non-descript films occur on the bedding planes, and do not allow for accurate interpretations other than that they were probably marine. The stromatolites are discussed in more detail under sedimentary strucutres.

The plants *Cooksonia hemispherica* Lang and *C. pertoni* Lang have been found in the Bertie at Passage Gulf, and possibly in the Buffalo, New York area (Banks (Cornell University), 1983, oral communication). It is not known with certainty if they were washed in, or if they were native to the environment. In addition, the vascular nature of both species of *Cooksonia* and of *Hostimella silurica* Goldring has not been proven for the Bertie material (Banks, 1972, p. 367, 373). The idea that vascular plants may occur in the Bertie, however, is not new (Banks, 1972, p. 374). Regardless, their significance lies in the constraints placed upon any hypothesis of depositional environment for the Bertie.

The remainder of the Bertie fossils are too poorly known to comment on relative to their paleoecology. The same can be said with regard to the trace fossils, although a study of them is in progress. For example, infrequent coarser-grained-filled, relatively horizontal burrows are suggestive of either a burrowing crustacean (?) or of the worm *Ruedemannella obesa* (Ruedemann).

SEDIMENTARY STRUCTURES

Studies of sedimentary structures within the Bertie Formation, without exception, qualitatively use the structures to identify the depositional environment. One of the main purposes of this paper is to relate all of the sedimentary structures to a depositional environment, its subsequent history, and to discuss the significance of each structure in terms of epeiric sea sedimentation. Also, structural relationships as well as stratigraphic occurrences will be examined, and

TABLE 3. Sedimentary structures and bedding types occurring in the Bertie Formation. Forms marked with an asterick are those that have been observed at the field stops for this trip. Sedimentary structures and bedding types listed in approximate order of frequency.

Sedimentary Structures	Bedding Types
Shrinkage Cracks*	Tidal Bedding*
Gypsum Crystal Molds*	Massive Bedding*
Reticulate Ridge Halite Casts*	LLH Stromatolites*
Salt Hopper Casts	Nodular Bedding*
Mottled Coloration*	Cross-bedding
Micro-faults*	Intraformational Breccia
Mud Volcanos*	
"Diastemic Surfaces"*	
Trace Fossils*	
Windrows*	
Channels*	
Ripple Marks*	
Impact Prints*	
Oolites	

TABLE 4. Sedimentary structures and bedding types. Preliminary summary relative to placement within Irwin's (1965) energy and sedimentation zones. Structures marked with an asterick are not environmentally determinable at present.

Sedimentary Structures and Bedding Types	Epeiric Sea Model						
	X	Y		Z			
	1,	2	3	4	5	6	Land
Shrinkage Cracks						x	x
Gypsum Crystal Molds					x		
Reticulate Ridge Halite					x	x	
Salt Hoppers					x		
*Mottled Coloration							
*Micro-faults							
Mud Volcanos						x	
"Diastemic Surfaces"						x	x
Trace Fossils				x	x	x	
Windrow Accumulations				x	x		
Channels			x	x			
Ripple Marks					x	x	
Impact Prints						x	x
Oolites			x	x			
Tidal Bedding			x	x	x		
Massive Bedding	x	x	x				
LLH Stromatolites					x		
Nodular Bedding						x	x
Cross-bedding		x	x	x			
Intraformational Breccia	x	x	x				

where necessary or appropriate, to the fossils.

Table 3 lists all the currently known sedimentary structures occurring within the Bertie. Tabulation of the geographic and stratigraphic distribution of the structures and bedding types is not possible at this time, being the writers have not visited all known sites of the Bertie. Such a tabulation would greatly enhance the inferences regarding the depositional environment. Nevertheless, all the known sedimentary structures can be placed, without too much controversy, in the different energy and sedimentation zones of Irwin (1965) (Table 4).

A previous study by Tollerton (1983) outlined a general stratigraphic sequence of both the sedimentary structures and the fossils. The present paper presents more detailed work on the sequences. The sequences may be incomplete due to the effects of the depositional consequences of epeiric sea sedimentation (low-angle of depositional slope; compaction; frequent subaerial exposure; and reflux dolomitization).

Although the Bertie contains many sedimentary structures, most are either ignored, or not recognized as sedimentary structures. This condition exists for two reasons. First, most investigators are overly preoccupied with the eurypterids, to the exclusion of everything else. Second, the degree of preservation and the degree of weathering tend to obscure most of the structures. Many of them cannot be seen until the specimen has been cut and polished.

Structures that have been widely interpreted as mud-cracks (shrinkage cracks of desiccation origin) are common in the Bertie (Fig. 8). However, there is little direct evidence for their being interpreted as mud-cracks. Few of the diagnostic features (Table 5) that would substantiate the condition of subaerial exposure and subsequent desiccation are found directly associated with the Bertie shrinkage cracks. The only known sedimentary structures found associated with the shrinkage cracks are: 1)reticulate ridge halite casts; 2)small-scale mud volcanos; 3)micro-faults; and 4)LLH Stromatolites. With the possible exception of the reticulate ridge halite casts, the other structures are not exclusively indicative of subaerial exposure (Reineck and Singh, 1975). Preliminary studies show an apparent stratigraphic restriction of association. Thus far, the approximately top one foot of the upper-most beds of the Fiddlers Green Member are directly associated with the reticulate ridge halite casts, and the small-scale mud volcanos. Below this is another layer of shrinkage cracks associated only with the micro-faults. Another foot down is a third layer of shrinkage cracks associated only with the LLH Stromatolotes. Studies on the geographic continuity of this sequence are in progress.

Designation of a desiccation origin to shrinkage cracks is frequently based solely on their external appearance. Several blocks and slabs of the shrinkage cracks collected from the three layers mentioned above were cut and polished. The internal pattern found from all the shrinkage cracks (Fig. 3) was not expected for mud cracks solely of desiccation origin. Clearly, these shrinkage cracks are not simply due to subaerial exposure and desiccation. Their origins must account for both the internal pattern resembling a form of fluid escape and the surface polygonal pattern.

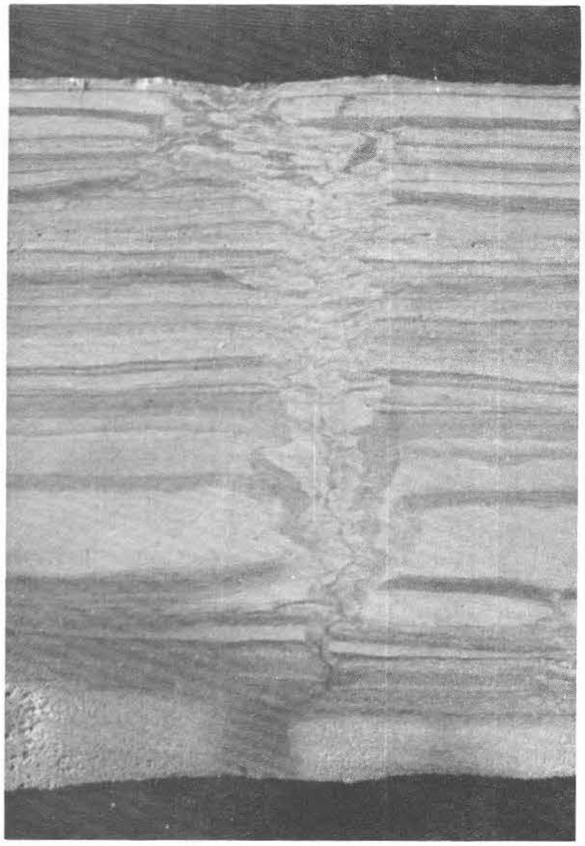

Figure 3. Desiccation crack in cross-section. Top of specimen is at top of photo. Approximately natural size.

Table 5. Summary of shrinkage cracks, origins, and characteristics. From: Plummer and Gostin (1981, Table 1, p. 1153).

Type of Shrinkage Crack and Plan Shape			Cross-Sectional Shape	Derivation of Infill	Number of Generations of Cracks	Preferred Orientation	Associated Features (Diagnostic)	Frequency Encountered in Geologic Record	Origin and Depositional Environment
DESICCATION	Polygonal	rectangular	V or U shaped	from above	can be multiple	?	raindrop imprints hailstone imprints bubble tracks foam impressions gypsum and/or salt casts flat-topped ripples vertebrate tracks etc.	uncommon	Algal or mud shrinkage by atmospheric drying. Exposed.
		hexagonal				none		rare	
		irregular						common	
	non-Polygonal	'incomplete'						uncommon	
		radial				?		rare	
		ribbon				parallel to water retreat			
SYNAERESIS		polygonal		from above or below	generally one	none	none	rare	Surface or substratal dewatering of submerged muds. Submerged.
		spindle				can be aligned	often with load phenomena	common	
		sinuous				along ripple troughs	ripple marks		

Oomkens (1966), in a study of the environmental significance of sand dikes, showed material from the Ubari Basin in Southwestern Libya (a playa setting), and from the Lower Permian of West Germany, as examples of sand dikes with polygonal surface patterns. His illustrations are remarkably similar to the Bertie material. His explaination of this structure applies to that observed in the Bertie, and increases our confidence in the environmental interpretation. Although Oomken's material is from a Recent playa setting, the basic processes are presumably the same, and are considered viable for epeiric sea environments. The sequence of events following Oomken's (1966, p. 146) interpretation is: 1)subaerial exposure; 2)shrinkage due to desiccation; 3)cracking; and 4)expulsion of the wet sediments from below the hard, impermeable surface layers. Evidence for such an interpretation for the Bertie is as follows: 1)generally V-shaped cross-section of the cracks; 2)"mobilized" internal pattern of the cracks; 3)the surface polygonal pattern; and 4)association of mud volcanos and mocro-faults with the cracks.

Additional support for this generic interpretation is found in the similarity of depositional regimes between coastal playa and sabkha settings, as well as lacustrine settings (Kinsman, 1969; Picard and High, 1972; Reineck and Singh, 1975; Shearman, 1978; Shinn, 1983b, and Smoot, 1983) with the depositional processes postulated for epeiric sea environments (Irwin, 1965).

For the present, a reconstruction of the paleosalinity based upon the shrinkage cracks (Baria, 1977) has not yet been attempted because of the many variables. However, as suggested by Shinn (1983b, p. 175), mudcracks preserved in carbonate rocks are almost invariably restricted to supratidal and upper intertidal zones.

The sedimentary structures shown in fig. 4 are interpreted as gypsum crystal molds. This interpretation is based on a comparision of similar material from Ohio, described by Summerson (1966). According to Summerson (1966, p. 223) the occurrence of gypsum crystal molds, in association with stromatolites and a restricted fauna, indicates conditions of a penesaline environment. Similar gypsum crystal patterns are observed in sabkhas and tidal flats of the Persian Gulf area (Kinsman, 1969; Reineck and Singh, 1975; and Shearman, 1979).

The sedimentary structures from the Bertie shown in fig. 5 have been interpreted by Tollerton (1983) as reticulate ridge halite casts. This form of skeletal halite was first described by Southgate (1982, p. 395) from silicified specimens from the Middle Cambrian of northern Australia, as "...a network of mutually perpendicular ridges that forms by the preferential precipitation of halite on the edges of cubes." According to Southgate (1982, p. 405), the reticulate ridge halite is indicative of very shallow brine pools that evaporated to dryness. Also the brines exhibited rapid variations in brine concentration. If the structures from the Bertie are indeed reticulate ridge halite, then they are strong evidence that at least portions of the Bertie were aubaerially exposed and desiccated, especially because of the associated desiccation cracks (as opposed to layers without desiccation cracks as reported in the Jurassic of Massachusetts by Parnell (1983, p. 711)). Exactly how shallow the brine pools were is still speculative. However, the experimental data of Southgate (1982, p. 404) suggests that reticulate

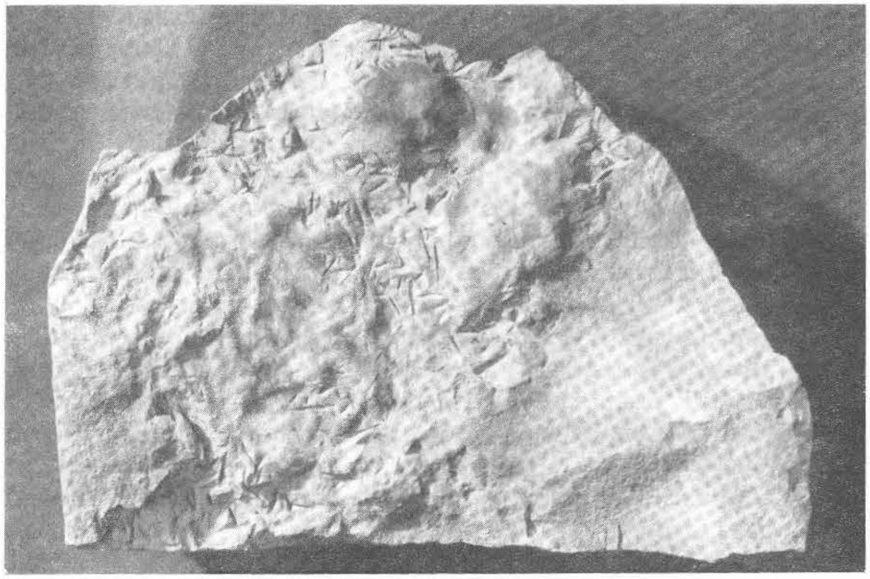

Figure 4. Gypsum crystal molds. Specimen collected from talus. Approximately one-half natural size.

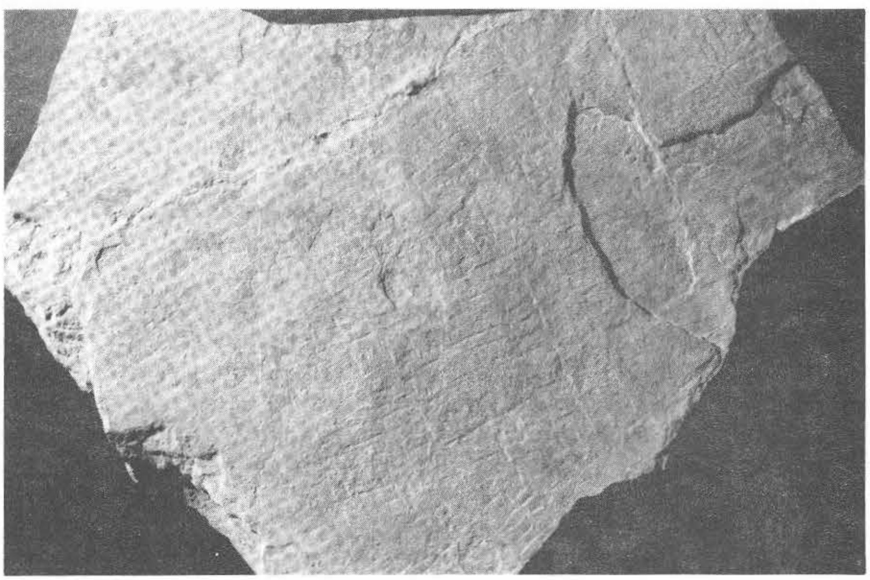

Figure 5. Reticulate ridge halite casts. Under-surface shown. Top shown in fig. 8. Approximately one-third natural size.

ridge halite, such as from the Bertie, may have precipitated in brine pools with depths on the order of only a few centimeters.

Although none have been observed by the writers, one example of salt hopper casts is reported by Ciurca (1978, p. 230) as occurring in the Fiddlers Green Member (his Phelps Member) at Passage Gulf. Ciurca also reports (1978, p. 230) that salt hoppers are characteristic of the upp ermost Fiddlers Green Member (his Phelps Member) at localities further to the west. This supports the extremely shallow-water origins postulated for the reticulate ridge halite casts. Experimental evidence from Southgate (1982, p. 404) suggests that with deeper water, the salt hoppers will form, while in shallow water, the reticulate ridge halite will form. The question remains, however, how shallow is shallow and how deep is deep. Published studies relating the distribution of skeletal halite morphologies with water depth seem to be inconclusive, at least quantitatively. Qualitatively, detailed field investigation of the size and abundance per square meter of the skeletal halite morphologoes may prove rewarding.

Some specimens from the Bertie that show a mottled coloration have been tentatively interpreted either as the result of bioturbation (fig. 6) or the result of compaction (fig. 7). Those specimens labelled as bioturbation (fig. 6) indicate an active infauna, even in extremely adverse ecological conditions (e.g., hypersalinity), as indicated by the association of salt (?) crystal molds. That the fauna was indigenous is indicated by the presence of geopetal brachiopod molds. Both of these are shown in fig. 6.

Those specimens identified as resulting from compaction (fig. 7) are virtually indistinguishable from the published figures of Shinn and Robbin (1983, fig. 15B, p. 607), and Shinn (1983a, fig. 1B, p. 621), both of which show the results of laboratory produced compaction on Recent shallow-water and lagoonal carbonate sediments. Beds showing the effects of compaction are apparently restricted to the lower portions of the Fiddlers Green Member. What process and mechanism produced the compaction? If the compacted beds are restricted to the lower portions of the Fiddlers Green Member, why don't the overlying beds also show effects of compaction?

The answers may lie in the nature of the stratigraphic record for the Bertie; a discontinuous nature that has not been recognized before. If the depositional record is discontinuous as suggested by the effects of compaction, then the Bertie Formation may have been deposited over a much longer period of time than previously supposed. On the other hand, the mottled beds may be burrowed subtidal beds, as the specimens closely resemble the published figures of Shinn(1983b, fig. 30, p. 190). Petrographic studies are in progress in an attempt to resolve the proper origin for these mottled beds.

The micro-faults seem to occur only with the shrinkage cracks, as well as with other structures found with the shrinkage cracks. The sense of movement is always that of a normal fault, and occur either en echelon or as micro-grabens. These structures are distinctly different from the off-set and "dropped" polygons of the larger shrinkage cracks. Whether these micro-faults are the result of compaction (Pettijohn and Potter,

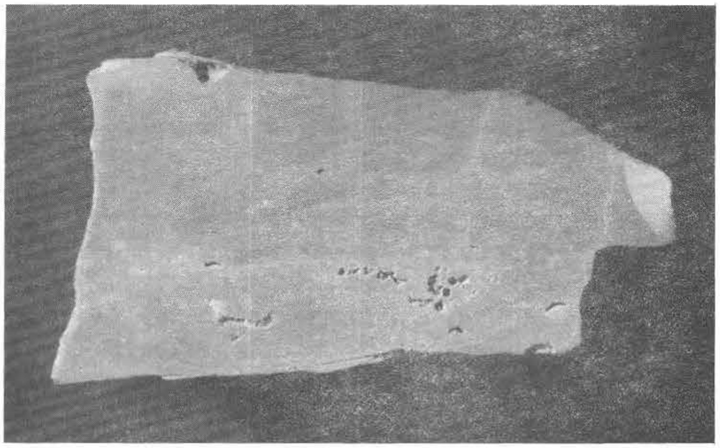

Figure 6. Mottled coloration due to bioturbation? Top of specimen at top of photo. Fenestrate (birdseye) structure or salt molds near bottom. Geopetal brachiopod in lower right. Approximately natural size.

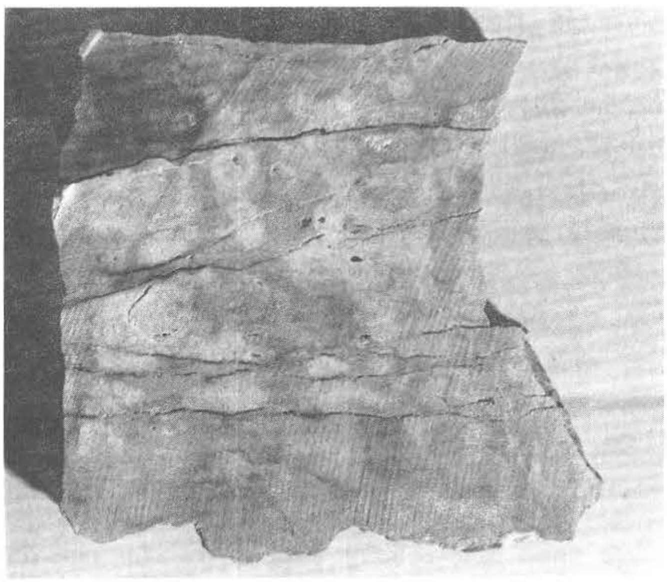

Figure 7. Mottled coloration due to compaction(?) or normal subtidal deposition (?). Top of specimen at top of photo. Approximately one-half natural size.

1964, Plate 111B), flowage (Wells, Prior, and Coleman, 1980), or earthquake shock (Sims, 1973), is not known. Any of these hypotheses are equally valid. One suggestion for study would be detailed outcrop maps of their in situ occurrences and orientations.

Only one specimen of mud volcanos exists in the senior writer's collection (figs. 8 and 9). As such, the specimen has neither been slabbed, nor have thin-sections been made. However, other sedimentary structures are in direct association with the specimen; shrinkage cracks, and reticulate ridge halite casts. This assemblage of sedimentary structures suggests that the mud volcano is the result of water expulsion due to either desiccation or compaction, or both.

So far as is known, the mud volcanos are only found in the uppermost zone of shrinkage cracks within the Fiddlers Green Member, about three inches below the contact with the overlying Scajaquada Member.

The term "diastemic surface" is used to designate bedding-plane surfaces that, when associated with features indicating subaerial exposure, also indicate extremely short periods of non-deposition. The actual duration of non-deposition is not yet determinable.

Because the depositional regime of epeiric sea sedimentation operates on an extremely shallow slope, such "diastemic surfaces" are believed to be a logical consequence of such a depositional regime. These "diastemic surfaces" are common throughout the Bertie, with at least nine such surfaces identified in the Fiddlers Green Member at Passage Gulf. It may be that these "diastemic surfaces" are PAC boundaries, but this is open to discussion, and requires additional field study.

Only a few trace fossils are known from the Bertie; those tentatively identified as the result of bioturbation (see mottled coloration), and those found with the ripple marks discussed later. Another form of trace fossil are the stromatolites. Forms of stromatolites that are not recognized as such are discussed later. They strongly resemble the intensely shriveled algal mats figured by Ginsburg (1957, fig. 15, p. 95). Some algal structures may have been erroneously identified as "worm burrows".

Several writers (e.g., Ciurca, 1978; and Hamell, 1981) have noted the occurrence of oriented fossils in parallel zones that are about one to two feet in width. These parallel zones may be either windrow accumulations, or due to the effects of compaction (most of the microfaults occur in the immediately overlying beds). Excavation for the eurypterids will generally expose the windrows. However, specimens of complete windrows are not known to exist, a condition which greatly hampers detailed study of their origins. Windrow accumulations in the Bertie are concentrations of generally undeformed and unbroken fossils, and may indicate deposition in quiet, shallow-water areas of tidal flats (Reineck and Singh, 1975). Such accumulations as have been seen in the field, suggest a biocenose, because 1)gross size-measurements of whole eurypterids follow a normal distribution, and 2)high-spired gastropods, ostracods, and brevicone cephalopods have been found on the same bedding planes as the eurypterids.

Figure 8. Small-scale mud volcano with shrinkage cracks. View is top of bed. Bottom shown in fig. 5. Approximately one-third natural size.

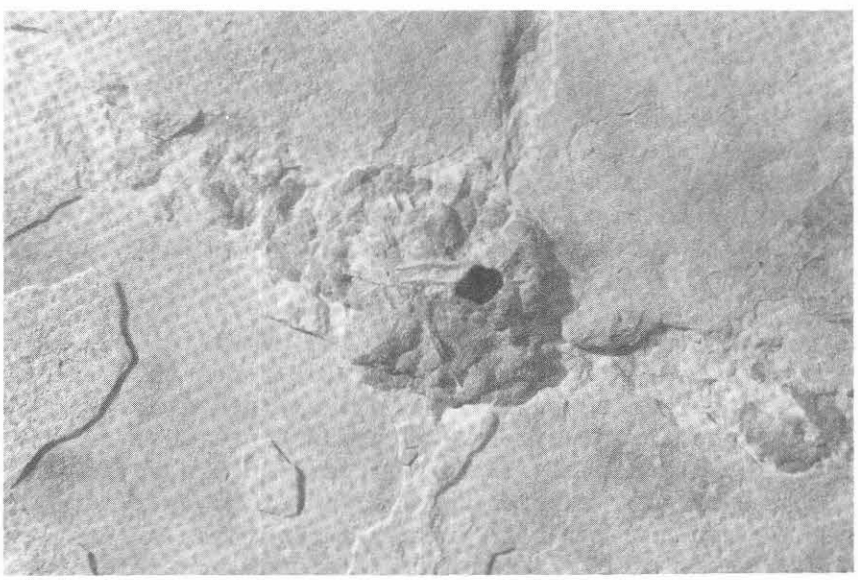

Figure 9. Close-up of fig. 8; small-scale mud volcano. Current flow toward the left. Current modified fecal mound? Approximately twice natural size.

A few examples of probable small-scale channels are known from the Bertie. However, the specimens are from small loose pieces, which severely limits interpretation. They may be cross-beds or cross-laminations, or may be due to the effects of compaction. Until specimens are located in situ, their value for interpretation is limited.

Only one specimen of ripple marks is known to the writers. It was collected at stop 2, on Jerusalem Hill Road. The material is a fine-grained, gray, dolomicrite, with a ripple index of 9. The ripples are slightly asymmetric, and probably formed by wind generated small waves. Depth of water was probably extremely shallow.

Features similar to worm burrows and worm trails with these ripple marks, supports a shallow-water interpretation, in that none of the burrows and trails show any signs of either rapid escape or burial.

No specimens of raindrop imprints from the Bertie have been reported in the literature: they may yet be found. Tracing the known horizons from the Brayman Formation containing such raindrop imprints may not prove fruitful, as such horizons are probably very local.

Using the technique of Metz (1982), structures shown in fig. 10 have been positively identified as hailstone imprints. To date only one horizon of hailstone imprints has been located in the Bertie.

Strata containing oolites have been observed only at outcrops of the Bertie in Canada. Although exposures in western New York State have not been examined by the writers, oolites have not been reported from any Bertie outcrop in New York State. It may be that the oolitic beds seen in the Canadian outcrops were deposited in "deeper" water and subjected to subtidal and tidal current action, yet still within the overall realm of shallow epeiric sea sedimentation (sedimentation zone III in energy zone Y; see figs. 1 and 2). Another explaination, is that dolomitization by reflux action has obscured the oolites in the "shallower" New York State sections.

Included in tidal bedding are flaser, lenticular, wavy, and laminated bedding. Although no detailed work has been completed at this writing, preliminary study of slabbed specimens suggests that they are of tidal origin.

Massive bedding appears to be volumetrically dominant, and devoid of sedimentary structures. Additional detailed petrographic studies are in progress at this writing.

Those structures from the Bertie shown in figs. 11 and 12 are interpreted as LLH Stromatolites of Logan, Rezak, and Ginsburg (1964). Portions of tidal bedding (e.g., laminated bedding) may also be stromatolitic in origin; additional thin-section study is needed before further comments can be made.

The significance of the LLH Stromatolites in the Bertie is that they indicate very shallow water that was subject to low wave and/or current action (Logan, Rezak, and Ginsburg, 1964, p. 77-79).

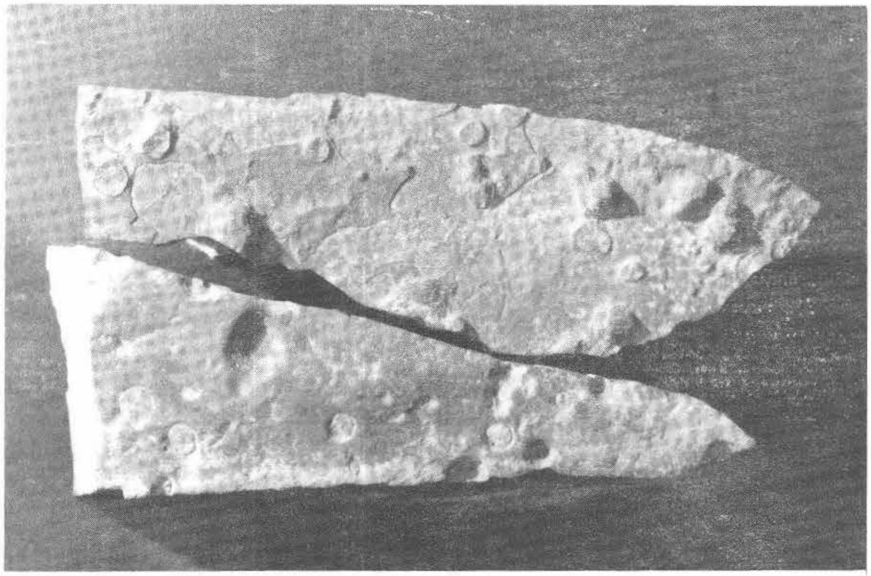

Figure 10. Hailstone imprints. Cast on top specimen is bed bottom. Approximately three-quarters natural size.

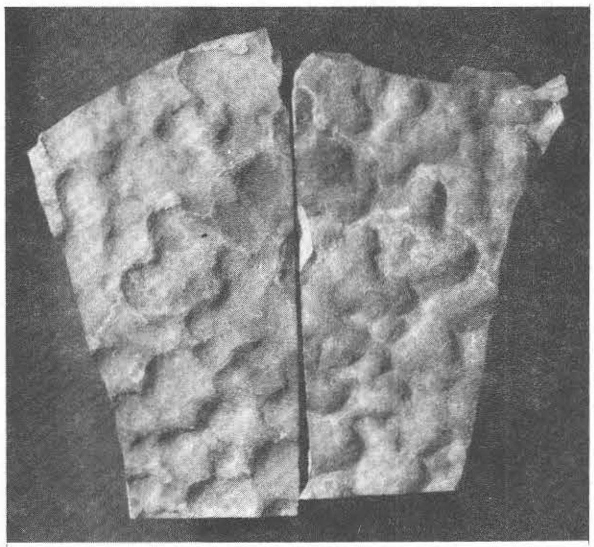

Figure 11. LLH Stromatolites. Cast on right is top of bed; mold on left is bottom of bed. Approximately one-quarter natural size.

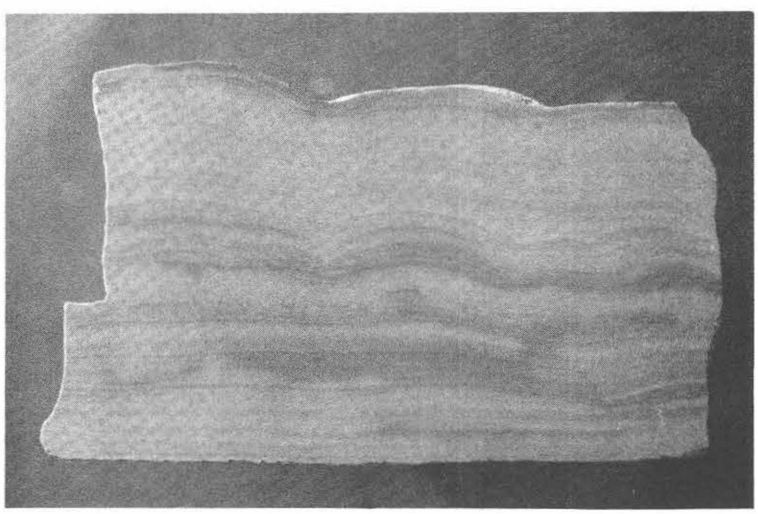

Figure 12. Corss-section of LLH Stromatolites. Top of specimen is top of photo. Approximately natural size.

Nodular bedding is observed in the Scajaquada Member wherever the member is exposed. The nodules appear to be calcite pseudomorphs after gypsum, and are indicators of supratidal deposition (Kinsman, 1969, and Shearman, 1978).

Cross-bedding has been reported from the Bertie (Ciurca, 1978), but none has been observed by the writers in the field. It is believed that the cross-bedding that may occur would be of small scale and of local stratigraphic importance because of the extremely low slopes involved in epeiric sea sedimentation.

An intraformational breccia in the Bertie probably represents a storm deposit. Only one such bed has been identified, which is the Ellicott Creek Member of Ciurca (1982, p. 103). Its absence east of Phelps, New York, may be evidence of regressive conditions within the Bertie. Whether the breccia indicates 1)that wave and/or current action was insufficient to form the breccia in the "shallower" eastern portions of the epeiric sea, or 2)that the probability of preservation was greater in Irwin's energy zones Y and X, are not clear.

CONCLUSIONS

When the total assemblage of fossils and sedimentary structures are examined and studied in detail, the epeiric sea model of Irwin (1965) is preferred over previously proposed environmental models. The epeiric sea model is the only model that apparently accounts for all the fossil groups, sedimentary structures, and interpretations proposed in the present paper (Table 6).

Although several problems still remain, it is hoped that this paper will stimulate further research, not only on the Bertie Formation, but also with regard to the formation of many of the sedimentary structures whose origins are unknown or are highly problematic. Such research is presently being conducted by the writers.

TABLE 6. Environmental relationships of the Bertie Fm. for proposed models

Structures/Bedding Types	River	Delta	Playa Lake	Lagoon	Epeiric Sea
Shrinkage Cracks			x	x	x
Gypsum Crystal Molds			x		x
Reticulate Ridge Halite Casts			x		x
Salt Hopper Casts			x		x
Mottled Coloration	x	x	x	x	x
Micro-faults	x	x	x	x	x
Mud Volcanos			x		x
"Diastemic Surfaces"		x	x		x
Trace Fossils		x		x	x
Windrow Accumulations		x		x	x
Channels	x	x		x	x
Ripple Marks	x	x		x	x
Impact Prints		x	x	x	x
Oolites				x	x
Tidal Bedding		x		x	x
Massive Bedding			x	x	x
LLH Stromatolites				x	x
Nodular Bedding			x	x	x
Cross-bedding	x	x		x	x
Intraformational Breccia		x		x	x
Fossils					
Conularids				x	x
Corals				x	x
Bryozoans		x		x	x
Brachiopods		x		x	x
Gastropods	x	x		x	x
Cephalopods		x		x	x
Pelecypods	x	x		x	x
Worms				x	x
Scorpions		x		x	x
Eurypterids	x	x		x	x
Phyllocarids		x		x	x
Xiphosurans				x	x
Ostracods		x		x	x
Edrioasteroids				x	x
Graptolites				x	x
Conodonts		x		x	x
Algae				x	x
Vascular (?) Plants	x	x		x	x
Trace Fossils		x		x	x
Interpretations/Observations					
Extremely shallow water	x	x	x	x	x
Extremely low-angle slope		x	x		x
Frequent periods of exposure			x	x	x
Reflux dolomitization			x	x	x
Geographic restriction of fossils	x	?			x
Geographic restriction of structures	x	?	x		x
Stratigraphic restriction of fossils	x	?			x
Stratigraphic restriction of structures	x	?	x		x
Hypersaline conditions			x		x
Little or no wave action	x		x		x
Little or no current action			x	x	x
Biocenoses		x		x	x

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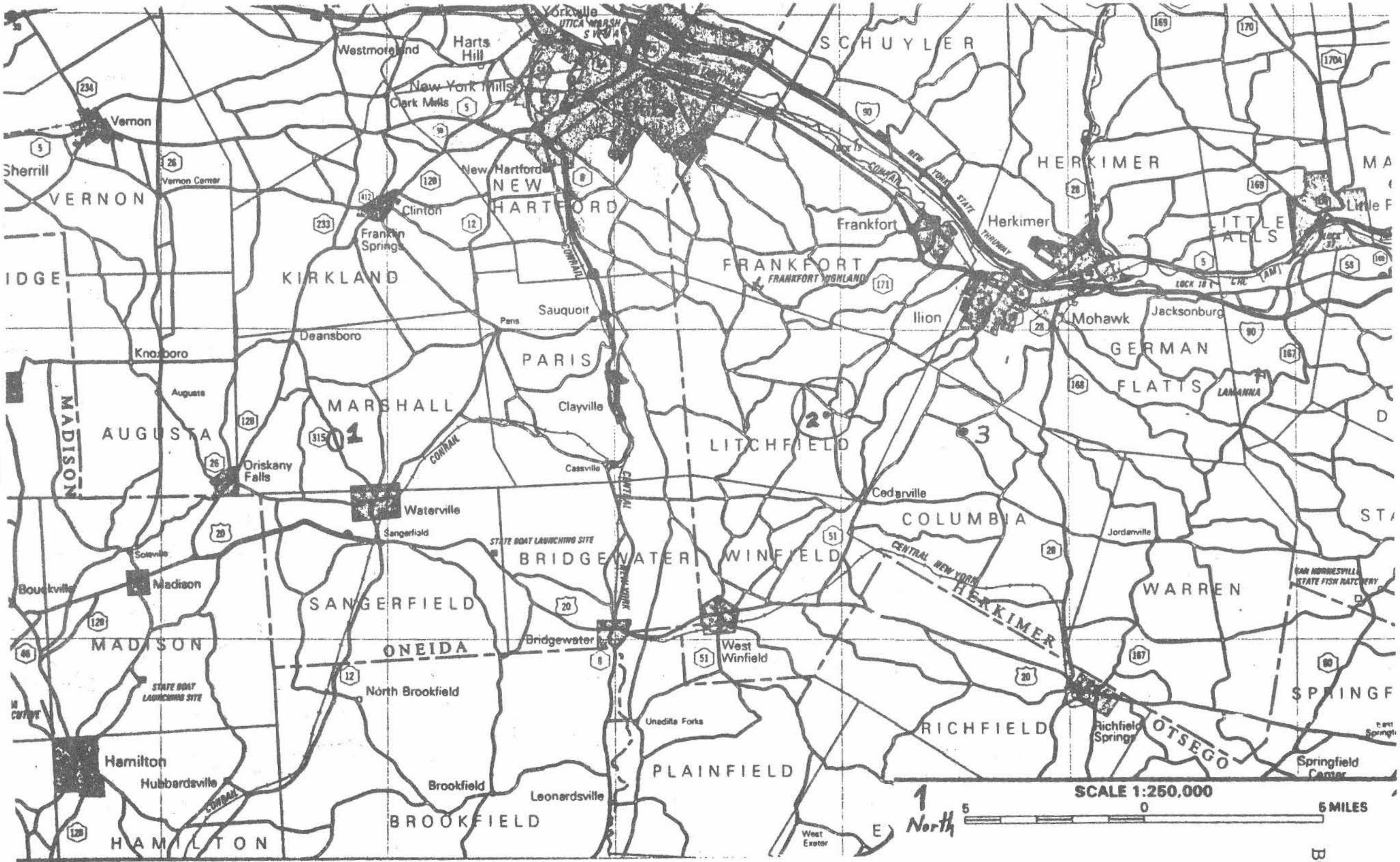
Yorkville
UTICA MARSH
Westmoreland
Harts Hill
New York Mills
Clark Mills
SCHUYLER
Vernon
Sherrill
VERNON
Vernon Center
New Hartford
NEW HARTFORD
Clinton
Franklin Springs
KIRKLAND
HERKIMER
Frankfort
Herkimer
LITTLE FALLS
FRANKFORT
FRANKFORT HIGHLAND
Ilion
Mohawk
Jacksonburg
GERMAN FLATTS
LAMANNA
IDGE
Sauquoit
Pens
Deansboro
Knoxboro
PARIS
MARSHALL
Augusta
MADISON
AUGUSTA
Clayville
LITCHFIELD
2
3
1
CONRAIL
Oriskany Falls
Cassville
CENTRAL
Cedarville
COLUMBIA
Waterville
Sangerfield
STATE BOAT LAUNCHING SITE
BRIDGEWATER
WINFIELD
Jordanville
Solsville
Madison
Bouckville
CENTRAL NEW YORK
HERKIMER
WARREN
VAN HORNESVILLE STATE FISH HATCHERY
SANGERFIELD
MADISON
ONEIDA
Bridgewater
West Winfield
North Brookfield
STATE BOAT LAUNCHING SITE
Unadilla Forks
RICHFIELD
Richfield Springs
OTSEGO
SPRINGF
Hamilton
Hubbardsville
PLAINFIELD
Springfield Center
Brookfield
Leonardsville
SCALE 1:250,000
5 0 5 MILES
North
BROOKFIELD
HAMILTON
West Exeter
E

Figure 13.


ORISKANY FALLS QUADRANGLE
NEW YORK
7.5 MINUTE SERIES (TOPOGRAPHIC)
NW/4 SANGERFIELD 15' QUADRANGLE
UTICA (P.O.) 14 MI.
CLINTON 4.6 MI.
Creek
Deansboro
Small Corners
Gibbons Corners
Grant Hill
Oriskany
Big
M A R S H A L L
Hanover
Forge Hollow
1
Creek
SCALE
CONTOUR INTERVAL 20 FEET
DATUM IS MEAN SEA LEVEL

Figure 14.

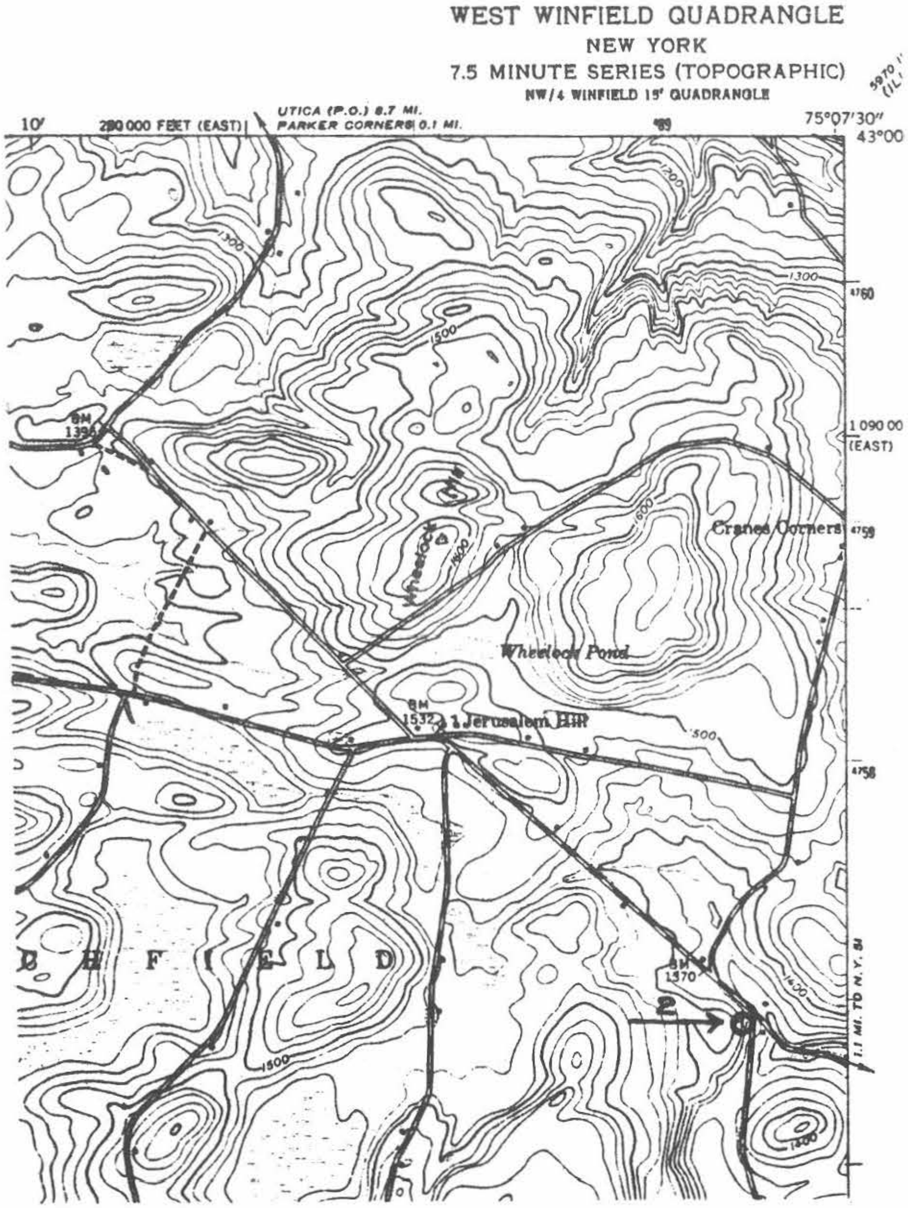

Figure 15a

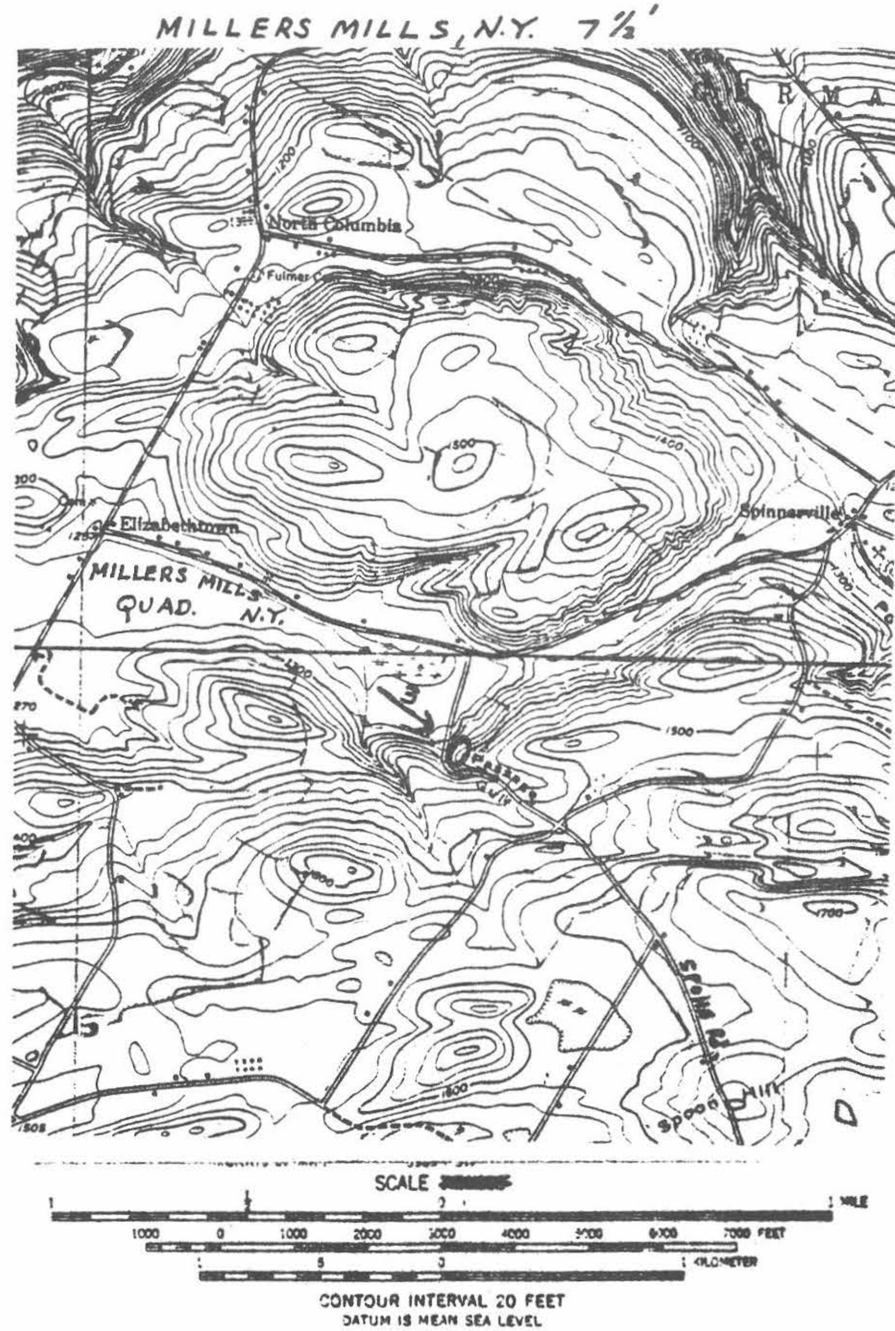

Figure 15b

BERTIE FORMATION ROAD LOG

Miles

0.0 Start at flashing light intersection of Rte. 233 and College Hill Rd. (412); bottom of Hamilton College Hill. Proceed south on Rte. 233 (Harding Rd.).

1.2 "T inters." with Rte. 12B, turn right.

3.7 Cemetery on left.

4.2 Village of Deansboro. Turn left onto Rte. 315 east.

4.5 Bridge over Oriskany Creek.

7.0 STOP 1 Outcrop on right. Big Creek on left. Parking on right.
Forge Hollow, New York, on Route 315 (fig. 14)
Oriskany Falls, N.Y., 7 1/2 minute topographic quadrangle
Rickard (1953) locality No. 43.

This locality is believed by Leutze (1959, p. 101-102) to be the Waterville outcrop cited by Vanuxem (1982) and Hall (1859). Rickard (1953, 1962) designated this site as the type locality for his Forge Hollow Member (= Scajaquada Member of present terminology).

This is the only complete sequence of the Bertie Formation in east-central New York. Both the lower and upper contacts with the Camillus and Cobleskill Formations, respectively are exposed. The lower contact is at the southern end of the exposure.

Fossils collecting is highly discouraged during this field trip because of the dangerous conditions at this outcrop (e.g., blind curve, narrow road, steep outcrop). KEEP OFF THE ROAD.

Nearly all faunal and sedimentary structural elements discussed in this paper have been found by the writers or reported by others as occurring here. The purpose of this stop, then, is examination of the fossils and sedimentary structures, as well as their stratigraphic sequence.

Continue south on Rte. 315.

9.4 Intersection of Buell (315) and Main St. W (Rte. 12) in Village of Waterville. Turn right onto Main St. W (Rte. 12).

Miles	
9.5	Continue on Rte. 12 (Sanger Ave.), left fork.
10.7	Rte. 20 intersection, turn left, east.
18.3	Intersection with Rte. 8, Bridgewater. Continue east on Rte. 20. Outwash plain.
21.4	Intersection with Rte. 51, West Winfield. Turn left onto Rte. 51 (North St.).
22.6	Proceed to stop sign at intersection of N. Winfield Rd. and Stone Rd. Continue on N. Winfield Rd.
23.7	Proceed on right fork at "Y inters.", Brace Rd.
25.4	Intersection of Brace Rd. and Babcock Hill Rd. Continue on Brace Rd. (now Berberick Rd.).
28.8	"T inters." with Jerusalem Hill Rd. Turn right onto Jerusalem Hill Rd.
29.8	STOP 2 Jerusalem Hill Rd. outcrop. Parking in lot of Town of Litchfield Town Hall, across from outcrop.

Jerusalem Hill, Litchfield, New York (fig. 15a)
West Winfield, N.Y. 7 1/2 minute topographic quadrangle Lautze (1959) locality No. 207

This locality is the Litchfield outcrop cited by Vanuxem (1842) Hall (1859), Clarke and Ruedemann (1912), and Ruedemann (1925). The wheelock Hill locality is located about 1 1/2 miles to the NW.

Although many eurypterids have been collected here in the past, NO collecting is authorized by the town board.

The purpose of this stop is to examine the geographic and stratigraphic continuity of the fossils and sedimentary structures with those seen at stop 1, located about 16 1/2 miles straight-line distance to the west of this stop.

Continue east on Jerusalem Hill Rd.

30.3	"Y inters.", continue on left fork (paved road).
30.4	Quarry on right. Camillus Formation, 80 feet exposed. STOP: Depends on time. (Ref. Treesh, 1972. Stop III).

Miles

Beckus Gulf. CAUTION, narrow road. Syracuse Formation, nearly complete section, (Treesh, 1972, Stop II).

31.3 "T inters." Rte. 51. Turn right, south. Vernon Shale on right, (Treesh, 1972, Stop I).

33.8 Cedarville. Intersection with Jordanville Rd. Turn left, east.

33.9 LUNCH. Ward A. Wheelock, Jr. Community Park, behind firehouse. Grocery, short walk up road.

Continue east on Jordanville Rd., past Rte. 51 south, to Columbia Center.

37.9 Intersection at Columbia Center. McKoons Rd. on right. Turn left onto Columbia Center Rd.

38.1 Spohn Rd., turn left.

After Brennan Rd. SLOW-CAUTION.

40.2 STOP 3 Passage Gulf (Spohn Rd.).

Stop 3 Passage Gulf, on Spohn Road (fig. 15b).
Millers Mills, N.Y. 7 1/4 minute topographic gradrangle
Rickard (1953) locality No. 39; Leutze (1959) locality No. 211.

This locality exposes an areally small outcrop, that is remarkably productive with regard to fossils. The upper-most beds of the Fiddlers Green Member are often excavated, giving the appearance of a battlefield. The farmer who owns the land at this locality has recently bulldozed off the top layers of a portion of this site.

The entire Fiddlers Green Member is easily accessible for detailed examination of the sedimentary structures and their stratigraphic sequence.

The purpose of this stop is to examine the continued geographic and stratigraphic continuity of fossils and sedimentary structures with those seen at stop 1, located about 20 miles straight-line distance to the west of this stop. Also, discussion is encouraged at this last stop relative to epeiric sea environments and the Bertie Formation.

For those intending to reach N.Y.S. Thruway. Continue down Spohn Rd. to "T inters." with Brewer Rd. Turn right to Hamlet of Spinnerville intersection. Turn right on Polly Miller Rd. to Rte. 28 intersection at Getman Corners. Turn left (north) onto Rte. 28 to Main St., Village of Mohawk. Turn right, watch for Thruway signs. About 9 miles from STOP 3.

Return to Hamilton College, Clinton, N.Y. and Utica. Backtrack south on Spohn Rd.

42.3 Columbia Center Rd. Turn right, south.

42.5 Jordanville Rd. intersection. Turn right, west.

46.0 Rte. 51 south, intersection, turn left, south.

49.2 Rte. 20 intersection, turn right onto Rte. 20, west. Flea market on right. Might be open, some bargains.

52.2 Village of West Winfield.

59.8 Village of Bridgewater. Rte. 8 intersection. Turn right onto Rte. 8, north to Utica and UTICA COLLEGE of Syracuse University.

63.6 Intersection with Rte. 12. Turn right, north onto Rte. 12 to Waterville.

64.9 Village of Waterville, center, turn left onto Buell Ave. (Rte. 315) at light.

Proceed on Rte. 315 to Deansboro.

70.2 "T inters." with Rte. 12B, Village of Deansboro. Turn right.

73.2 Rte. 233, turn left.

74.4 Intersection with College Hill Rd. stop sign and light. Turn left, up hill to Hamilton College. Turn right to Clinton, N.Y. Straight ahead to Rte. 5 intersection. Straight to Thruway. Left to Syracuse. Right to Utica <u>and</u> Utica College of Syracuse University.

END OF TRIP.

The Sterling Mine, Antwerp, New York
A New Look at an Old Locality

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Institute for Sensory Research, Syracuse University, Syracuse, NY 13210

INTRODUCTION

The Sterling mine was the first American locality for the nickel sulfide, millerite. In the last hundred years it has produced many fine specimens of that mineral, including some that many regard to the the finest examples of the species ever found. In the past several years George Robinson at the National Museums of Canada and I have reexamined the geology, mineralogy, and origin of this landmark locality (Robinson and Chamberlain, 1984). Our interest was stimulated by our finding that much of the crystallized hematite in the ore had been reduced to magnetite and that some of the millerite had altered to the rather rare nickel silicate, pecoraite. The following pages summarize the history, geology, mineralogy, and paragenesis of this locality.

LOCATION

The Sterling Mine is in Jefferson County, New York, along US Route 11 between Watertown and Gouverneur (Fig. 1). The water-filled open pit (Fig. 2) and surrounding dumps may be reached by foot via a farm lane which runs east from US Route 11 at a point 3.75 miles north of the intersections of Routes 11 and 26 in the village of Antwerp. The mine appears on contemporary topographic maps of the Antwerp quadrangle as a small pond surrounded by low hills between Route 11 and Hawkins Creek. The property is owned by Mr. Raymond Villeneuve whose farm is 0.6 miles south of the entry lane on the east side of Route 11. Permission to collect must be secured before visiting the mine.

The Sterling Mine lies within a band of hematite deposits (Fig. 3) known as the Antwerp-Keene belt. According to Smock (1889) the occurrences from southwest to northeast were the Colburn, Ward, Dickson, White, and Old Sterling Mines owned by the Jefferson Iron Company; The Keene, Caledonia, and Kearney Mines owned by the Rossie Iron Works; and the Clark and Pike Mines owned by the Gouverneur Iron Ore Company. The Sterling and Caledonia mines were the largest of the group, were the only ones studied in any detail by the geologists of the time, and were the only producers of any quantity of mineral specimens.

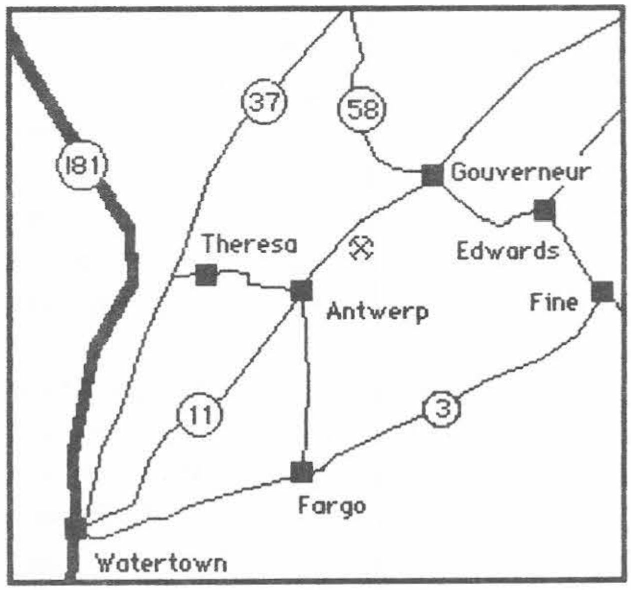


Figure 1. Location of the Sterling Mine.

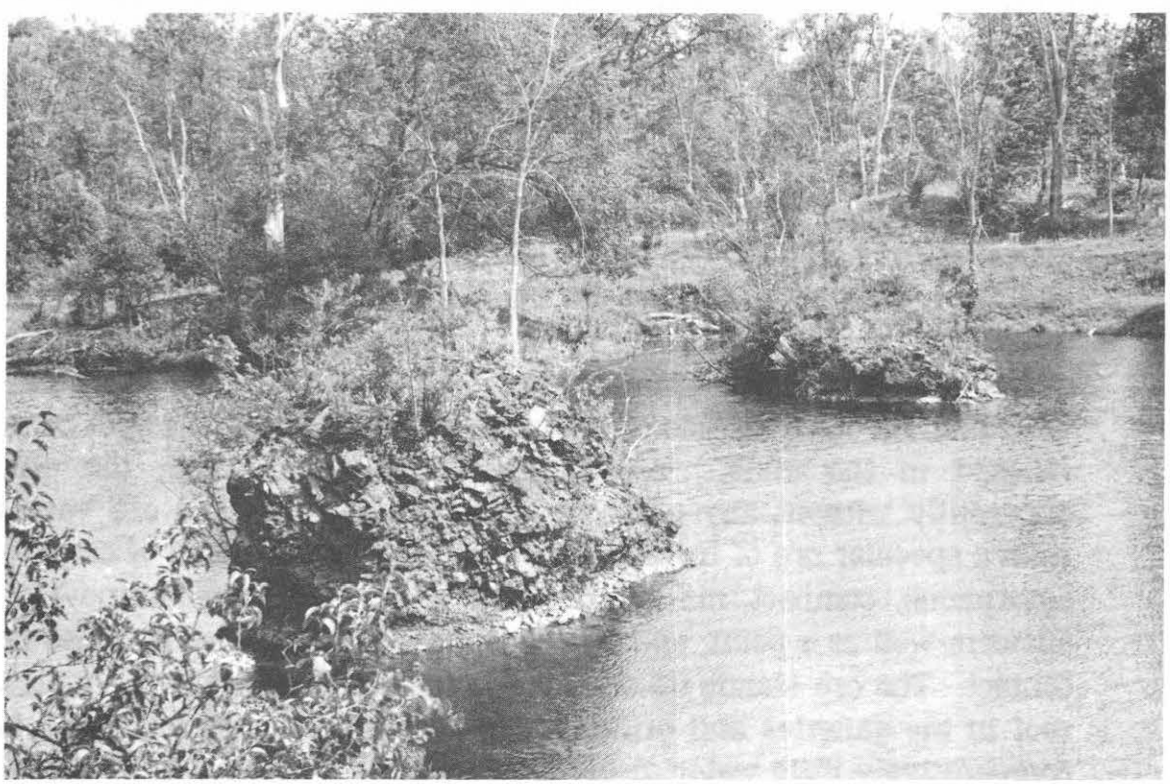


Figure 2. Looking east at the northern end of the Sterling Mine open pit in September, 1983. The two islands are topped by remnants of the cap of Potsdam sandstone. Photo by S. Chamberlain.

HISTORY

The Caledonia Mine, located about a mile southeast of Somerville in St. Lawrence County, was the first working mine in the district (Hough, 1853). Operations began in 1812, and by 1815 the Caledonia was furnishing ore to the Parish Iron Works in Rossie (Durant and Pierce, 1878). Between 1825 and 1835 the other deposits of the belt were discovered. One such deposit lay on an island surrounded by upland swamp on the farm of Hopestill Foster. In 1836, David Parish sold this property to James Sterling for $200, marking the beginning of the Sterling Mine (Hough, 1854; Haddock, 1895).

James Sterling, the "Iron King of Northern New York," formed the Sterling Iron Company in 1837. Ore was first hauled to his furnace at Sterlingville and later to Sterlingburgh. Pig iron was wrought using a charcoal, cold-air blast at the rate of about 12 tons per week. Horses were used to pump the water and raise the ore from the Sterling Mine pit. Later a hot-air blast was introduced and the Philadelphia Iron Company was organized. The Sterling Mine was operated during the late 1850's by Samuel G. Sterling (James' brother) and from 1859 onward by James' son A. P. Sterling. In 1869 the mine was sold to the Jefferson Iron Company of Antwerp (Emerson, 1898). Smock (1889, page 45) gives the following account of the operations under the Jefferson Iron Company:

> The open pit at the north-east is 115 feet deep, and approximately, 500 by 175 feet. The underground workings are south and south-west of it, and the ore has been followed for a distance of 900 feet, and to a depth of 185 feet. This deposit lies between the gneissic rocks on the south-east, 400 feet distant, and the sandstone (Potsdam) on the west side of the mine, but no walls have as yet been reached in the mine. A serpentine rock occurs with the ore, apparently without any order in its relations to it. The ore varies from a specular ore of metallic lustre and steel-gray shade of color to amorphous, compact masses of deep red. The crushed powder answers well as a point, and stains deeply all with which it comes in contact. The ore stands up well, and, by leaving pillars, with arched roof in the galleries and drifts, no timbering is necessary. There is comparatively little water in the mine. The serpentine is not so firm as the ore, and is full of slickensided surfaces. Small mine cars are used on the narrow gauge tramways in the mine drifts. A skip track

runs to the bottom of the open pit. A branch railroad three miles long connects this mine and the Dickson with the main line of the R.W. & O. R.R., near Antwerp...

By the 1880's the local furnaces had been largely abandoned and most of the ore was shipped to Pennsylvania for smelting. Competition from the Mesabi Range in Minnesota and a depressed economy gradually forced the closure of the mines in the belt. The Sterling Mine was last worked between 1904 and 1911 and was the last operating mine in the district (Newland, 1921). Buddington (1934) estimates the production of the Antwerp-Keene belt as 2.5 million tons, of which the Sterling Mine and adjacent Dickson Mine accounted for more than 750,000 tons (Smock, 1889).

The Rossie Iron Ore Company launched a brief, small-scale venture for making iron oxide pigments in 1942. In 1948, the Republic Steel Corporation conducted a diamond drilling program on the Caledonia and Sterling Mine properties, but concluded that further development was unwarranted.

GEOLOGY

The Sterling Mine ore body is bounded by Grenville marble to the northwest and gneissic granite to the southeast. It is locally overlain unconformably by an outlier of the Potsdam sandstone (Fig. 3). The rock immediately adjacent to the ore is heavily chloritized and slickensided and forms an irregular and intimate contact with it (Emmons, 1842; Hough, 1853; Smyth, 1894a,b). The ore itself contains many open cavities with marked evidence of stalactitic and botryoidal deposition of iron oxyhydroxides with subsequent crystallization or recrystallization of quartz and iron oxides, carbonates, and silicates.

Figure 4 shows a schematic longitudinal section through the ore body. The ore lies within a distinct unit of granitic gneiss locally rich in phlogopite, pyrite, and pyrrhotite. The unit is bounded on the hanging wall by crystalline marble and on the foot wall, by a garnetiferous, biotitic, granitic gneiss.

Although Emmons (1842) originally described the dark green rock associated with the ore as serpentine, and Shepard (1851) proposed it to be a new mineral species, dysyntribite, the best modern evidence suggests that it is a chloritized granite gneiss. Smyth (1894a,b) reported that in the operating mine, it was possible to observe areas in the footwall where the green rock graded gradually and completely into unaltered granite gneiss.

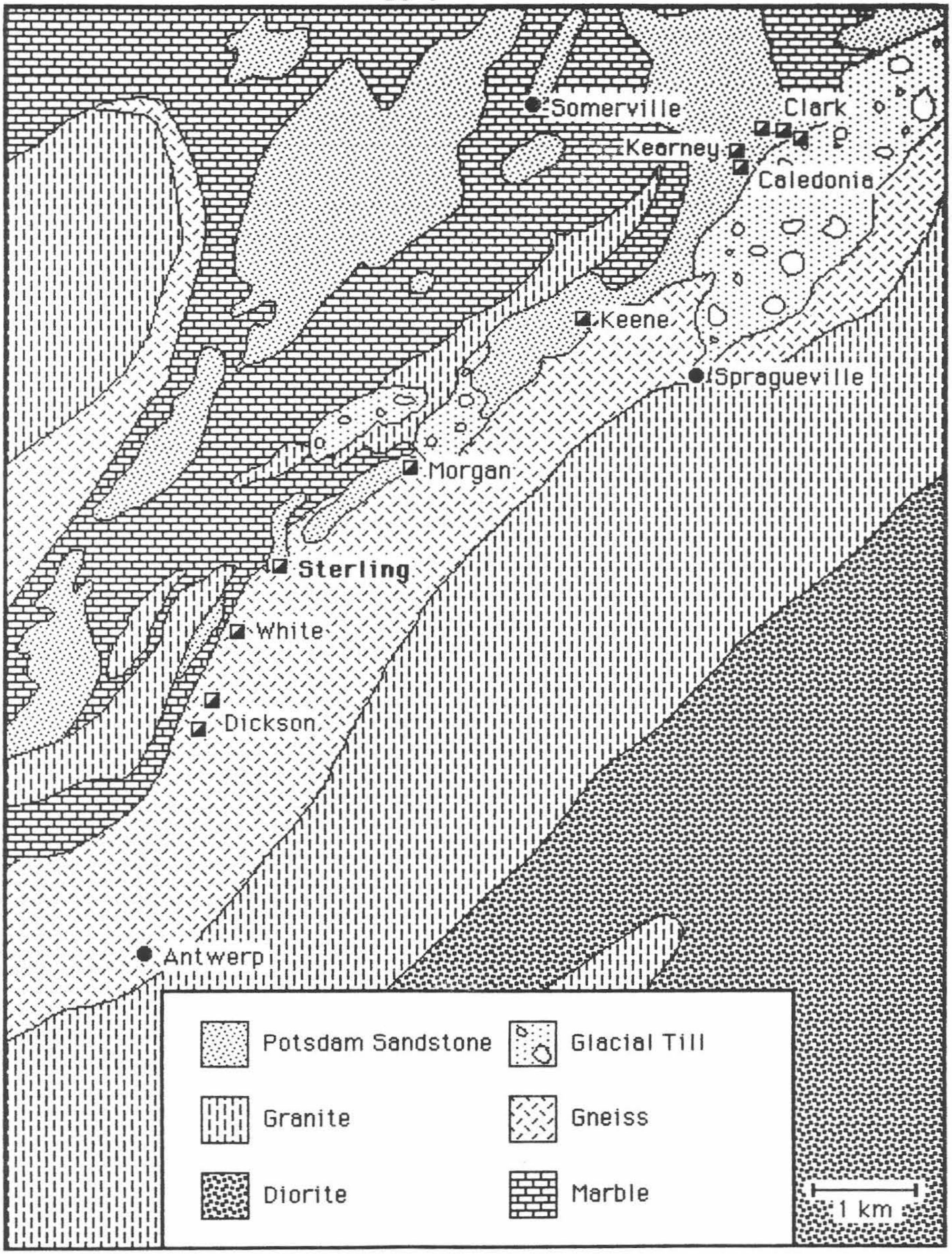


Figure 3. Geological map of the Antwerp-Keene belt of hematite deposits showing the location of the Sterling and other mines operated to exploit specular hematite (adapted from Buddington, 1934 and Newland, 1921).

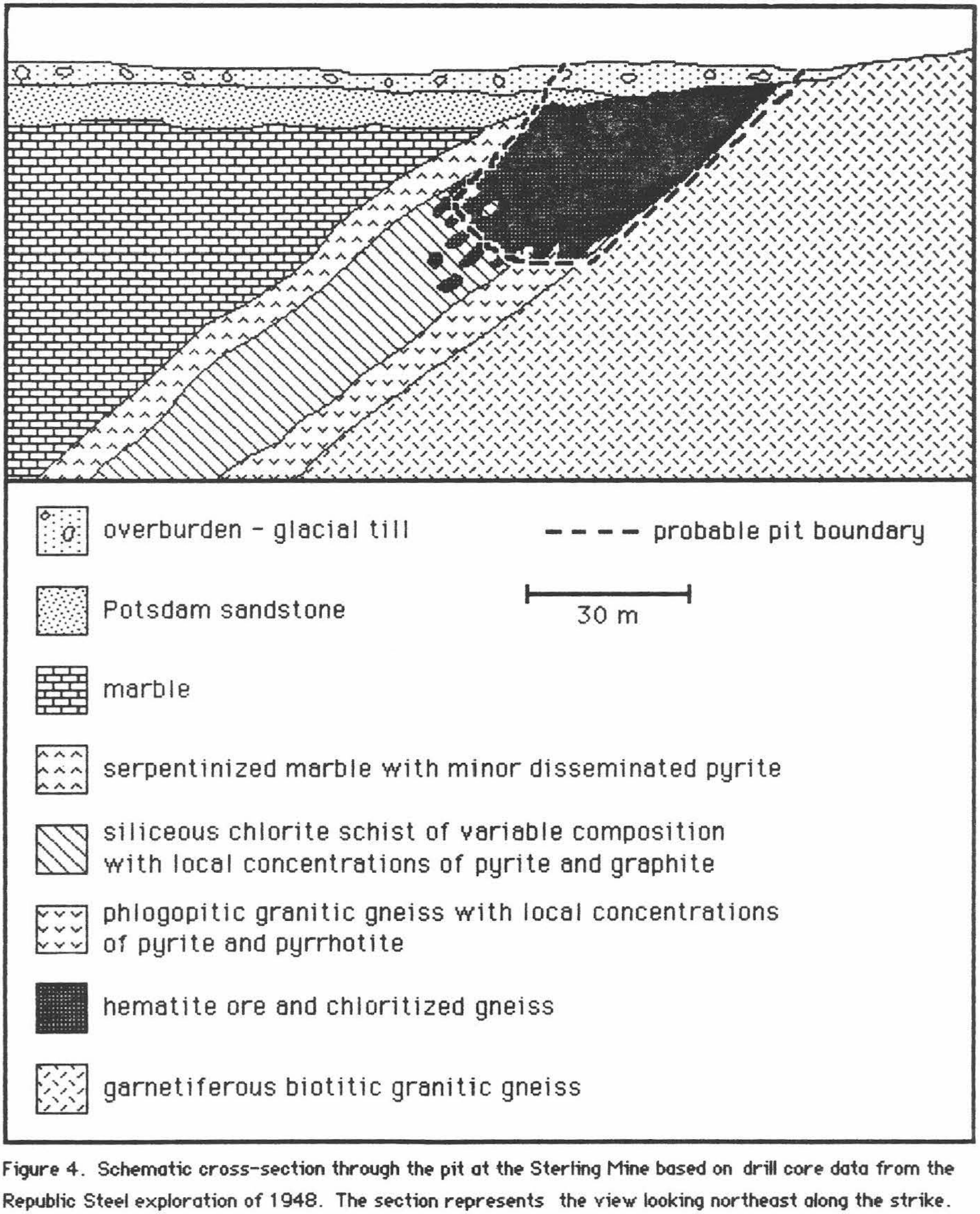


Figure 4. Schematic cross-section through the pit at the Sterling Mine based on drill core data from the Republic Steel exploration of 1948. The section represents the view looking northeast along the strike.

Buddington (1934) studied Smyth's thin sections and concluded that "the evidence all confirms Smyth's conclusions as to the origin of the chlorite and sericitic schists through alteration and replacement of gneiss or granite." Reexamination of the core logs and geologists' reports by W. M. Sirola and E. F. Fitzhugh, Jr. from the Republic Steel's diamond drilling program revealed clear evidence of complete and gradual chloritization of granite (Robinson and Chamberlain, 1984). Thus the ore body lies in a chloritized granite gneiss at its contact with the Grenville marble. The origin of the deposit will be discussed in greater detail under paragenesis below.

MINERALOGY

The majority of the minerals found at the Sterling Mine contain iron. Those of greatest interest to collectors occur in cavities in the hematite/quartz ore. These cavities range in size from a few millimeters to over 25 cm across. Hough (1853, page 683) gives the following description:

> **These red ores impart their color to whatever comes in contact with them, giving a characteristic tinge to every person and object about the premises. They are never crystallized, but occur in every variety of lamellar, slaty, botryoidal, and pulverulent forms, and in some cases, cavities are found lined with beautiful and peculiar crystallizations of carbonate of lime, spathic iron, heavy spar, aragonite, quartz, iron pyrites, and more rarely cacoxene or chalcodite, and millerite, the latter being the rarest and most beautiful of its associates. It occurs in brilliant needle shaped crystals, radiating from a centre like the fibres of a thistledown, having the color and brilliancy of gold. Groups of crystalline specimens of these minerals, often form objects of great beauty...**

No doubt the finest specimens were collected more than a century ago. Today, however, there are extensive dumps which have not been heavily dug and it is still possible to collect many of the species listed below, at least as microscopic crystals. Millerite and pecoraite are rather difficult to find, although hundreds of specimens have been found since 1979. Siderite, dolomite, quartz, stilpnomelane, and magnetite pseudomorphs after hematite are readily available.

The mineral descriptions which follow include only those species found in the crystallized cavities in the hematite ore. Species found in the marble, granitic gneiss and chloritic rock such as graphite, phlogopite, chlorites, micas, etc, will not be further discussed. These descriptions are based on the examination of several hundred hand specimens[1] and extensive field work at the locality, but in general, only the more common habits and associations are described.

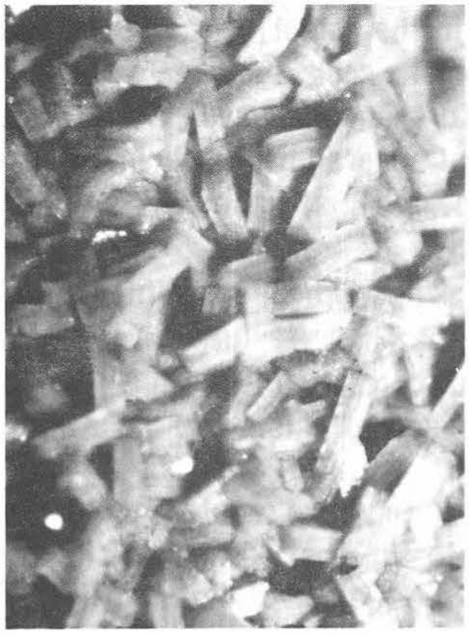
Apatite: A member of the apatite group, almost certainly carbonate apatite, occurs sparingly as microscopic tan, pink, or brown crystals associated with quartz and magnetite pseudomorphs after hematite (Fig. 5). The crystal forms observed are the hexagonal prism and basal pinacoid, rarely modified by the hexagonal dipyramid.

Aragonite: Aragonite occurs as clear to white crystals up to 1 millimeter in length. The crystals appear late in the paragenetic sequence. SEM photomicrographs show them to be relatively simple combinations of orthorhombic pyramids, pinacoids, and prisms (Fig. 6).

Calcite: At least three generations of crystallized calcite can be observed in cavities in the ore. The earliest is preserved as encrustation pseudomorphs of stilpnomelane after calcite scalenohedra, and is rarely encountered. The second generation consists of translucent, milky crystals of rhombohedral habit, described in detail by Whitlock (1910) (Fig. 7). The third generation is typically of the "nailhead" habit (Fig. 7). In addition, some cavities are partially or completely filled with a milky white, coarsely crystalline calcite. Whether this represents a continuation of one of these three generations, or an additional one is unknown.

Chalcopyrite: Chalcopyrite is uncommon at the Sterling Mine. Crystals up to a centimeter across are known, but most are only a few millimeters. The crystal forms observed include combinations of positive and negative disphenoids, {hhl} and {hhl}; two sets of tetragonal scalenohedra, {hkl}; and the tetragonal prism, {110}. Most specimens have a bright, metallic luster with occasional irridescence. A specimen in the Burridge collection at Harvard University, shows millerite needles protruding from a mass of chalcopyrite crystals with darkened surfaces and covered with minute sprays of acicular malachite. Similar specimens of malachite-covered chalcopyrite have recently been collected from the dumps.

[1]Specimens were examined from the following collections: Mineralogical Museum, Harvard University; Department of Geology, Hamilton College; National Display Collection, National Museums of Canada; Department of Geology, Cornell University; Geological Museum, Rutgers University; Department of Geology, Vassar College; New York State Museum; Ron Waddell, Syracuse, NY; Steve Chamberlain, Syracuse, NY; George Robinson, Ottawa, Canada; and Georgia and Everett Shaw, Cortland, NY.



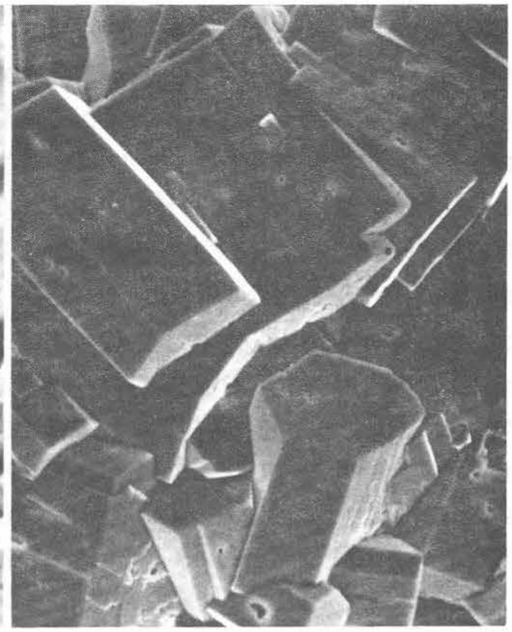


Figure 5a. Pink tabular carbonate apatite crystals. The field of view of about 1mm. Collection of Ron Waddell, Syracuse, NY. Photo by S. Chamberlain.

Figure 5b. Carbonate apatite crystals. The magnification of the SEM photomicrograph is about 300x. Photo by G. Robinson.

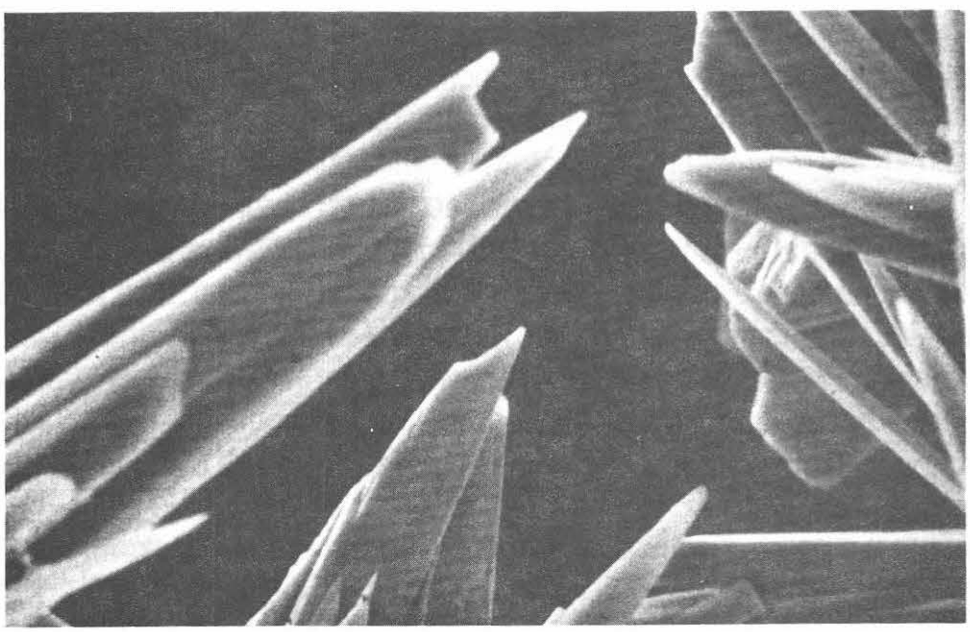


Figure 6. Aragonite crystals. The magnification of the SEM photomicrograph is about 400x. Photo by G. Robinson.

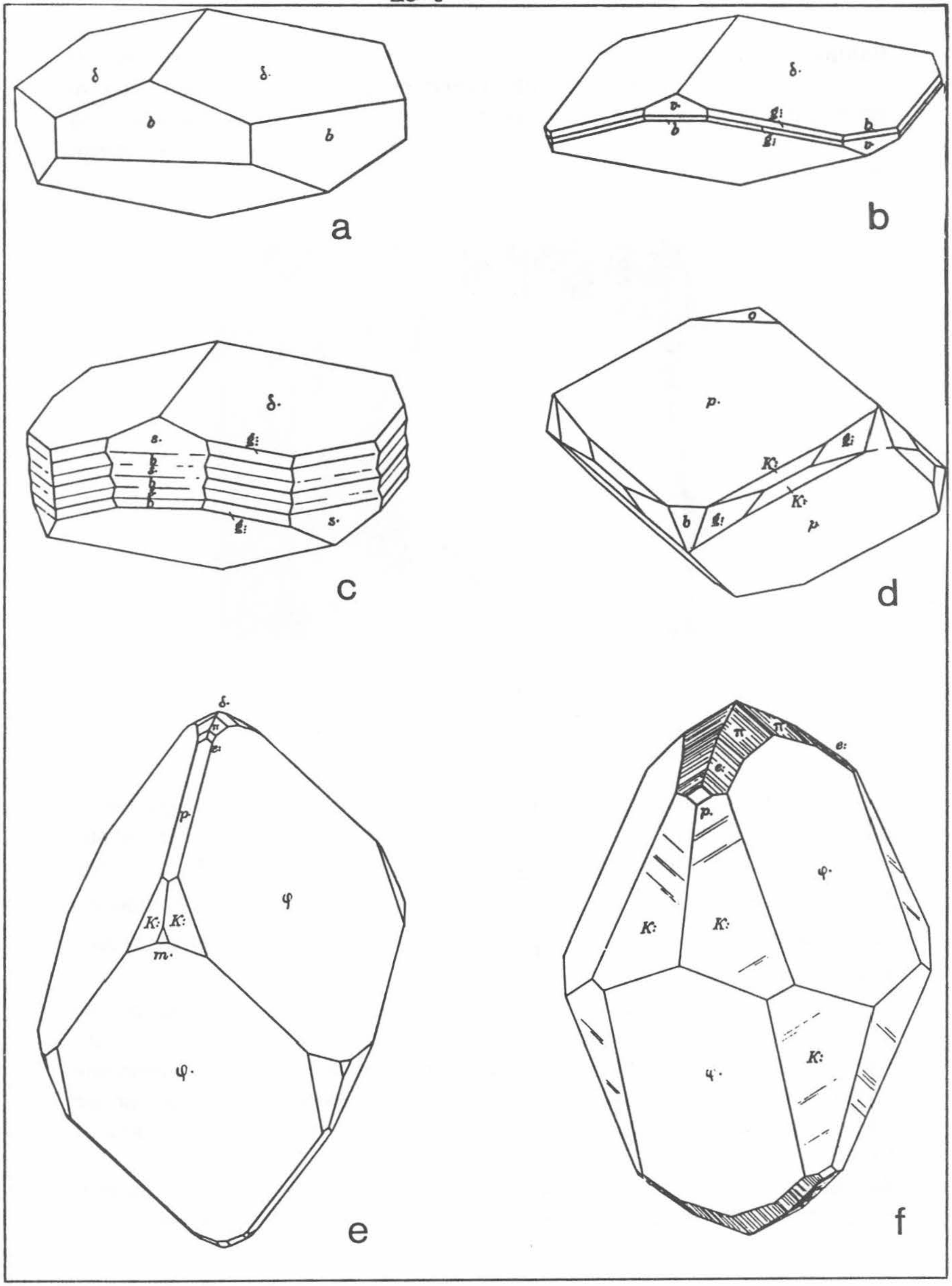


Figure 7. Crystal forms of calcite from the Sterling Mine. The second generation is shown in **d**, **e**, and **f**. The third generation is shown in **a**, **b**, and **c**. Modified from Whitlock (1910).

Dolomite: Dolomite is a relative common mineral at the Sterling Mine. It occurs as white, creamy, saddle-shaped crystals up to 5 mm, and as spherical aggregates of curved rhombohedra up to 1.5 cm in diameter (Fig. 8). Microprobe analyses show that this dolomite is actually a ferroan dolomite of composition $Ca(Mg_{.62}Fe_{.31}Ca_{.07})(CO_3)_2$.

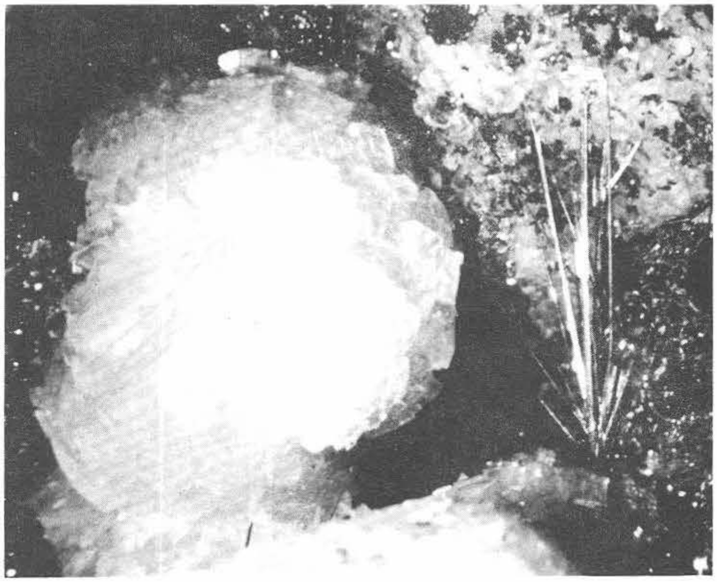


Figure 8. Spherical arrays of dolomite crystals with millerite, stilpnomelane, quartz, and magnetite pseudomorphs after hematite. The millerite spray is about 1 cm long. Oren Root Collection, Hamilton College, #H183. Photo by S. Chamberlain.

Commonly, dolomite crystals are epitaxially overgrown by light tan to dark reddish brown siderite, resulting in zoned crystals. Microprobe analyses of such zones crystals give $Ca(Mg_{.66}Fe_{.30}Ca_{.02}Mn_{.02})(CO_3)_2$ for the ferroan dolomite cores, $(Fe_{.75}Mg_{.16}Ca_{.08}Mn_{.01})CO_3$ for the light tan siderite overgrowths, and $(Fe_{.81}Mg_{.10}Ca_{.08}Mn_{.01})CO_3$ for the dark reddish brown siderite overgrowths

Ankerite is commonly listed in the literature as occurring at the Sterling Mine (e.g. Buddington, 1934; Palache et al., 1951; Roberts et al., 1974; Jensen, 1978). Neither X-ray diffraction nor electron microprobe studies revealed any ankerite at this locality. It seems likely that earlier wet chemical analyses of dolomite/siderite zoned crystals are the source of this misidentification.

Goethite: Goethite is relatively uncommon at the Sterling Mine, usually occurring as microscopic tufts of acicular golden crystals (Fig. 9) which formed relatively early in the paragenetic sequence. More rarely goethite occurs as crystalline botryoidal coatings on hematite crystals or as an alteration of magnetite pseudomorphs after hematite crystals. Massive brown "limonite" is not uncommon in the dump material.

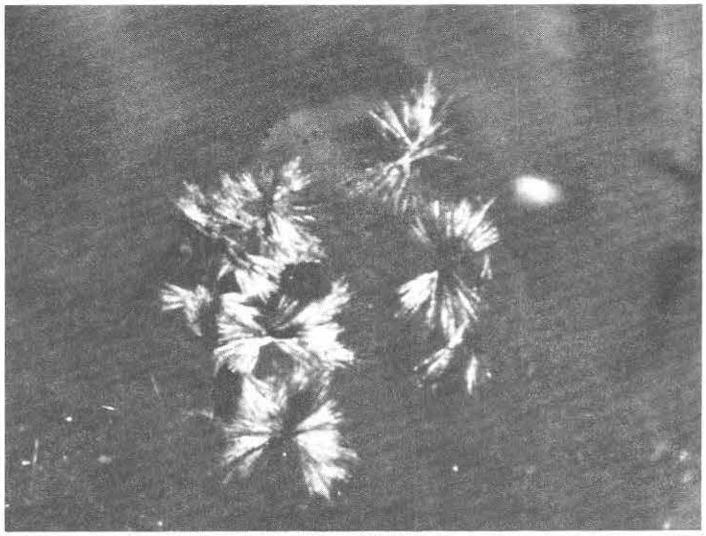


Figure 9. Golden sprays of acicular goethite on quartz crystals. The field of view is about 2 mm. Collection of Georgia and Everett Shaw, Cortland, NY. Photo by S. Chamberlain.

Hematite: Hematite was the principal ore mined, and as might be expected is common. Mostly it occurs as earthy, red masses. Steel-grey massive hematite and botryoidal (Fig. 10) and specular forms are less common. Unaltered crystals of hematite are rarely encountered because most of what appears to be hematite has been altered to magnetite.

Magnetite: Virtually all of what appears to be specular hematite is actually a replacement of hematite by magnetite. These pseudomorphs are typically small, flattened rhombohedra with black, lustrous faces Figs. 11 and 12). Although they appear to be hematite, they are strongly magnetic, have a black streak, and their X-ray diffraction pattern clearly shows strong magnetite lines.

Millerite: This is the species for which the locality is justly famous. Millerite was discovered at the Sterling Mine by Franklin B. Hough, a noted scientist, historian, and collector who lived in the nearby village of Somerville. Hough's original description (Hough and Johnson, 1850, page 287) includes:

> **It was first noted by the writer about two years since, and**

attracted his attention for its delicate capillary appearance, brilliant lustre and the difference of its crystalline form from that of sulphuret of iron, which in color and association it so nearly resembles. It occurs mostly in radiating tufts of exceedingly minute and slender crystals of brass-yellow color, and a very brilliant lustre, which when highly magnified present the appearance of flattened hexagonal prisms with striated faces, the striae being parallel with the principal faces of the prism... They occur in geode-like cavities of the iron ore, which are lined with crystallizations of spathic iron, specular iron, quartz, calcite, cacoxene, and sulphuret of iron; from among these crystals the tufts proceed, attached generally to the spathic iron, more rarely to the crystals of iron. It is not an abundant mineral; only perhaps one or two dozen specimens have been procured since its discovery... In a specimen which the writer procured of a miner,... a crystal was found completely transfixing a rhomb of spathic iron, and supporting it in air, at a distance of 1/8 inch above the inner surface of the cavity.

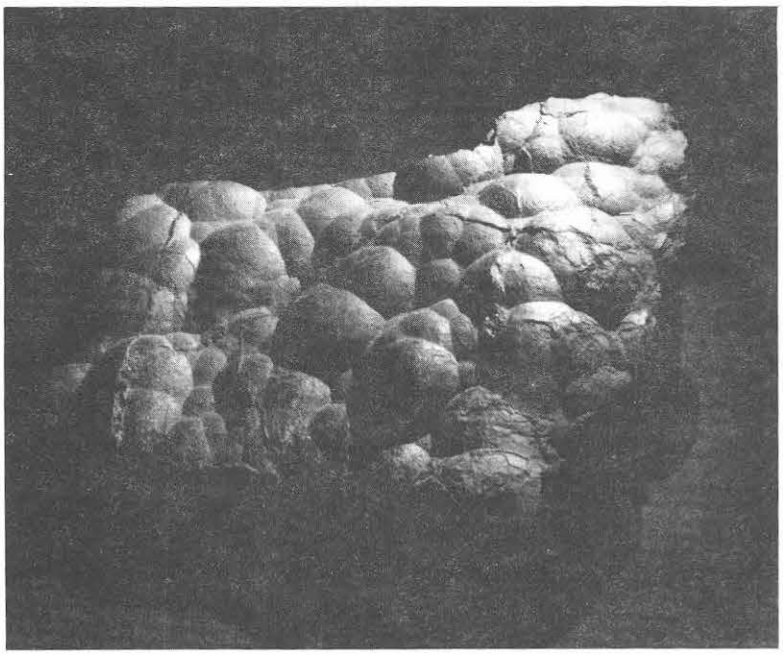


Figure 10. Red botryoidal hematite. The specimen is 15 cm in width. #H268, Oren Root Collection, Hamilton College. Photo by S. Chamberlain.

Sprays of acicular millerite crystals range from microscopic to about 3 cm in largest dimension. Individual crystals may be relatively sturdy and longitudinally striated (Figs. 8 and 13), delicate and curved, or spiraled. Occasionally smaller sprays are observed to radiate from a single thicker

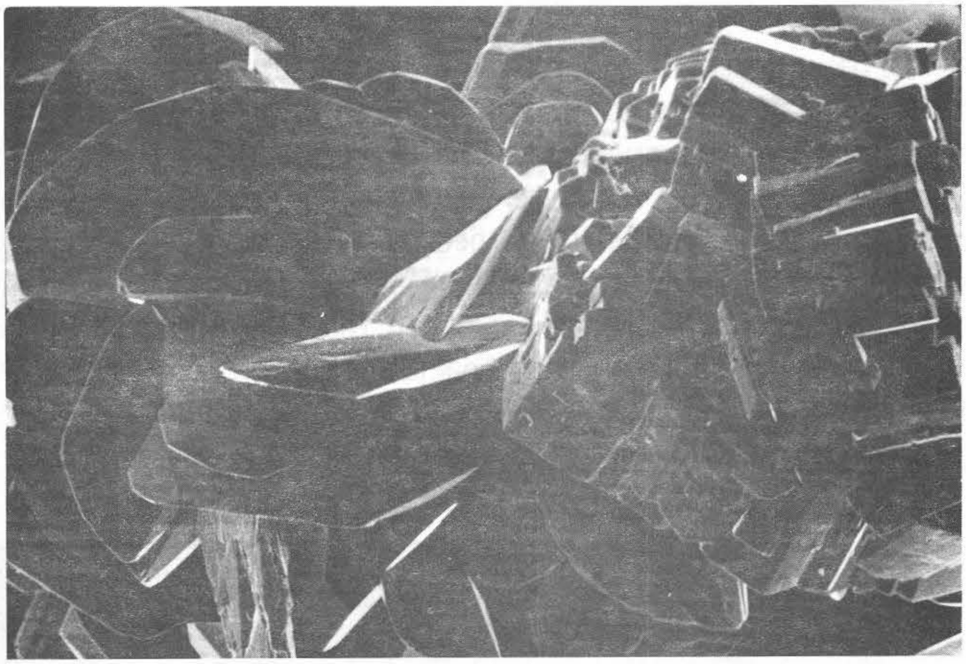


Figure 11. Black pseudomorphs of magnetite after rhombohedral crystals of hematite with pink carbonate apatite crystals. The magnification of the SEM photomicrograph is about 200x. Photo by G. Robinson.

Figure 12. Black magnetite pseudomorphs after rhombohedral crystals of hematite with pink carbonate apatite crystals. The magnification of the SEM photomicrograph is about 200x. Photo by G. Robinson.

crystal in a habit resembling a broom (Fig. 14).

Pecoraite: Some millerite at the Sterling Mine is altered to a light yellow to dark green secondary minerals. The degree of alteration ranges from thin, partial coatings to complete replacement (Fig. 15). Work on such alteration products is often difficult because they are not well crystallized. By combinations of X-ray diffraction, electron microprobe analysis, and TGA-EGA analysis, Robinson has concluded that the alteration product is largely pecoraite (Robinson and Chamberlain, 1984). The measured composition is $(Ni_{5.87}Mg_{.16})(Si_{3.76}Fe_{.30})O_{10}(OH)_8$ in excellent agreement with the theoretical composition $Ni_6Si_4O_{10}(OH)_8$.

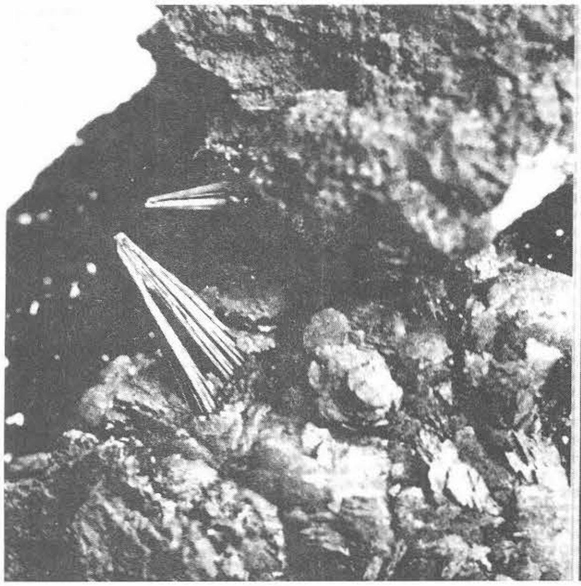


Figure 13. Millerite crystals on tan siderite crystals. The millerite spray is nearly 2 cm. #H281, Oren Root Collection, Hamilton College. Photo by C. A. Smith.

Figure 14. Millerite "broom" on magnetite pseud. after hematite crystals. Gore Collection, Harvard University. Photo by S. Chamberlain.

Pyrite: Pyrite is uncommon at the Sterling Mine. Usually it occurs as small cubes or octahedra rarely exceeding 1 or 2 mm on edge. The most interesting pyrite specimens are parallel growths of octahedral crystals on tan siderite (Fig. 16).

Quartz: Quartz is common in vugs as druses of clear crystals which rarely exceed 1 cm across. Stalactitic growths and ferruginous crystals are occasionally encountered. Some individual crystals have attenuated prism faces and resemble hexagonal dipyramids, while others have more complex morphology (Fig. 17). Quartz also occurs in the massive ore as yellow and brown jasper.

Siderite: Siderite is common at the Sterling Mine. Two color varieties are present. One is a light tan color (Fig. 16) and has a composition of $(Fe_{.77}Mg_{.13}Ca_{.10})CO_3$. The other is dark red to reddish brown and has a composition of $(Fe_{.81}Mg_{.10}Ca_{.08}Mn_{.01})CO_3$. Both types occur with about equal frequency as rhombohedral crystals up to 5 mm or as parallel groups up to several centimeters. The light tan crystals are occasionally observed to be impaled on millerite needles as described by Hough above.

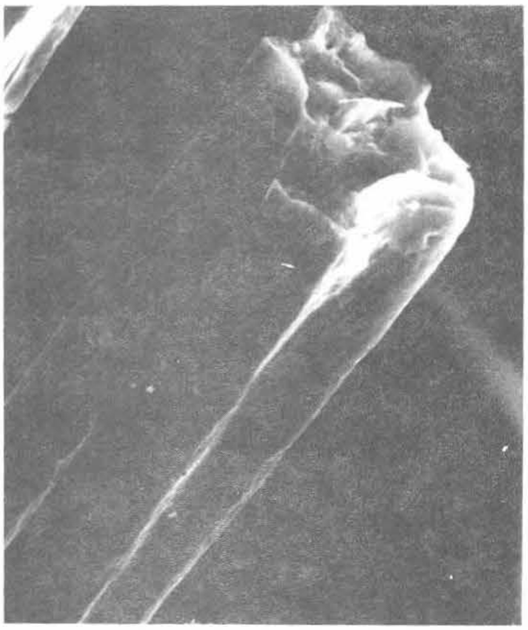


Figure 15. Pecoraite pseudomorph after millerite crystal. The SEM photomicrograph is about 1000x. Photo by G. Robinson.

Figure 16. Pyrite crystals on light tan siderite crystals. The field of view is about 1 cm. New York State Museum Collection, Albany, NY. Photo by S. Chamberlain.

Sphalerite: Sphalerite is rare at the Sterling Mine. Figure 18 shows a 6 mm, complex, twinned crystal from the Root Collection. This crystal occurs in a cavity with siderite, dolomite, calcite, chalcopyrite, quartz, and magnetite pseudomorphs after hematite.

Stilpnomelane: This species was originally identified as cacoxenite (Emmons, 1842; Beck, 1842; Hough and Johnson, 1850). It was put forth as the new species chalcodite by Shepard in 1852. A few years later, Brush (1858) finally correctly suggested that chalcodite was stilpnomelane. Both X-ray diffraction and electron microprobe analyses confirm its identity as stilpnomelane. It occurs as velvety golden-brown, green-brown, and green coatings, and small isolated spheres. Close examination shows arrays of micaceous crystals (Fig. 19) up to 2 mm across. It commonly fills fractures in the ore and the chloritic rock, and is usually found in crystal-lined cavities in the ore. It also occurs more rarely as encrustation pseudomorphs after calcite scalenohedra and an unknown orthorhombic mineral, perhaps anhydrite (Fig. 20).

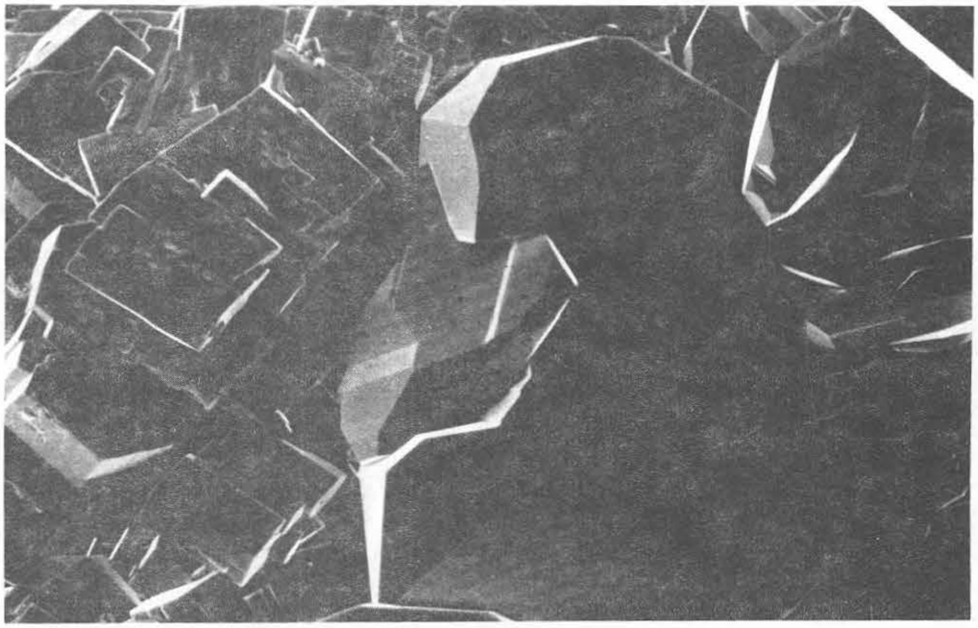


Figure 17. Colorless quartz crystals with pink carbonate apatite crystals. The magnification of the SEM photomicrograph is about 250x. Photo by G. Robinson.

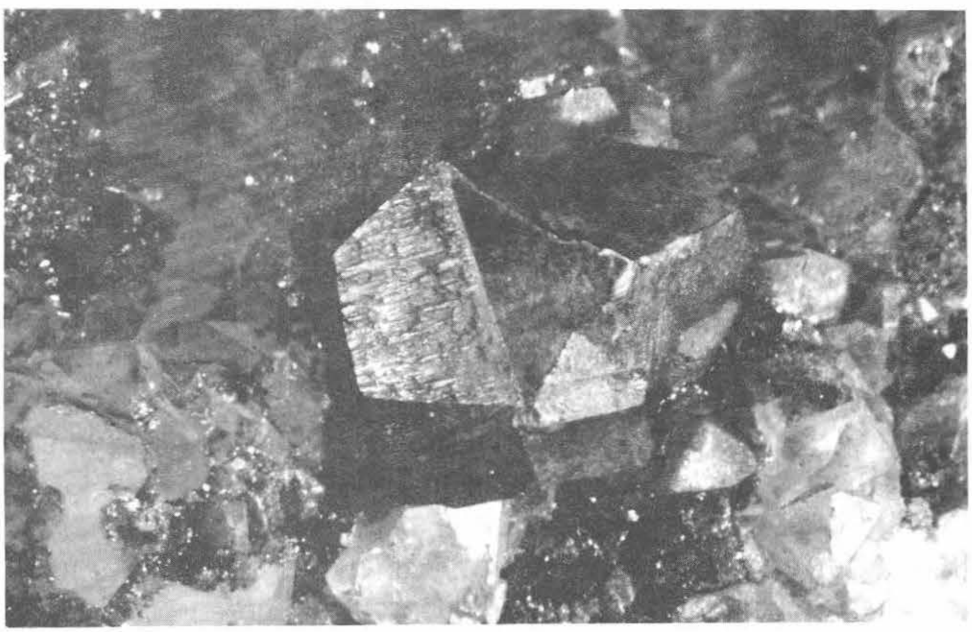


Figure 18. Twin crystals of dark brown sphalerite. The crystals are 6 mm on edge. #H228, Oren Root Collection, Hamilton College. Photo by S. Chamberlain.

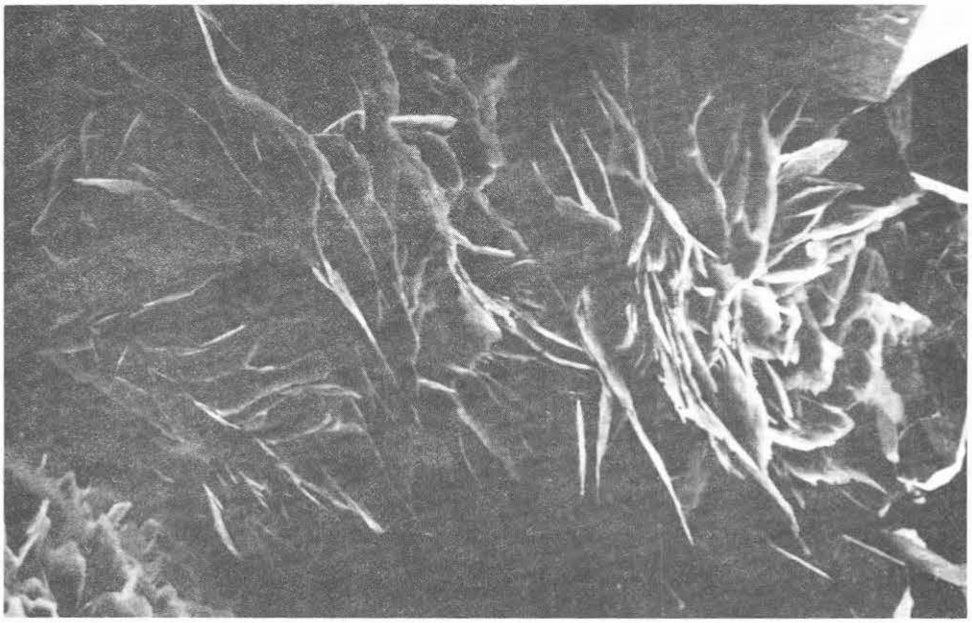


Figure 19. Stilpnomelane with quartz. The magnification of the SEM photomicrograph is about 225x. Photo by G. Robinson.

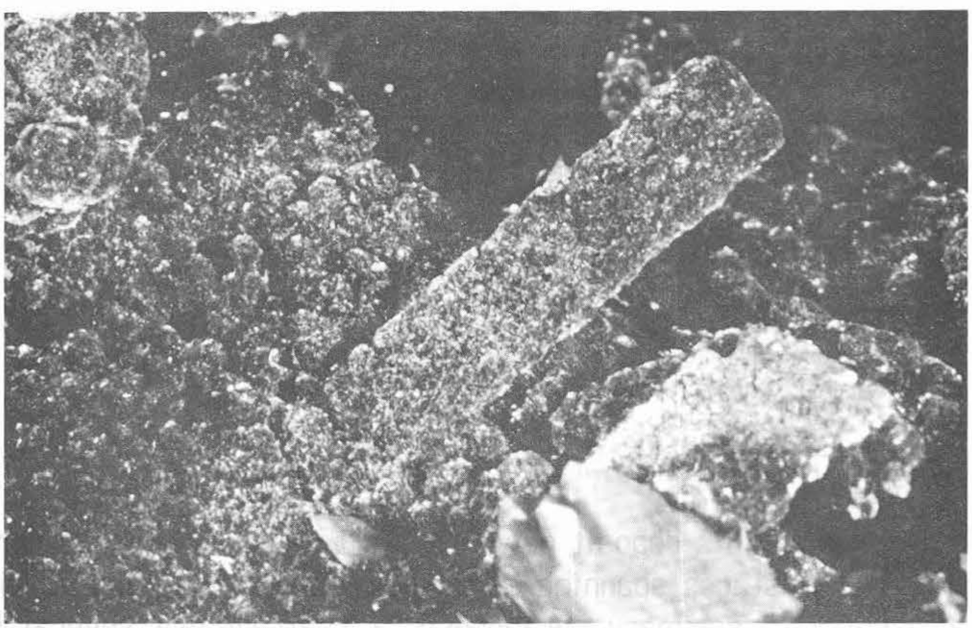


Figure 20. Encrustation pseudomorph of stilpnomelane after an unknown mineral, possibly anhydrite, with siderite. The pseudomorph is about 1 cm. #2435, Steve Chamberlain Collection, Syracuse, NY. Photo by S. Chamberlain.

Talc: An iron-rich talc has been found at the Sterling Mine. It occurs as gray, clay-like masses in the walls of some cavities in the ore, and as spheroidal aggregates of off-white microscopic crystals (Fig. 21) in others. It is most commonly associated with quartz, stilpnomelane, and magnetite pseudomorphs after hematite. Less commonly, striated molds of an unknown minerals, possibly anhydrite, are observed in the massive talc. A composite formula for this talc, based on electron microprobe, TGA, and wet chemical analyses, is $(Mg_{1.90}Fe(II)_{.78}Fe(III)_{.26}Al_{.02})Si_{4.04}O_{10.06}(OH)_{2.23}$ in excellent agreement with the general talc formula $(Mg,Fe)_3Si_4O_{10}(OH)_2$. Note that the when iron-rich talcs are synthesized (Forbes, 1969) ferric iron substitutes for Si, but in this iron-rich talc, both ferrous and ferric iron are substitution for Mg.

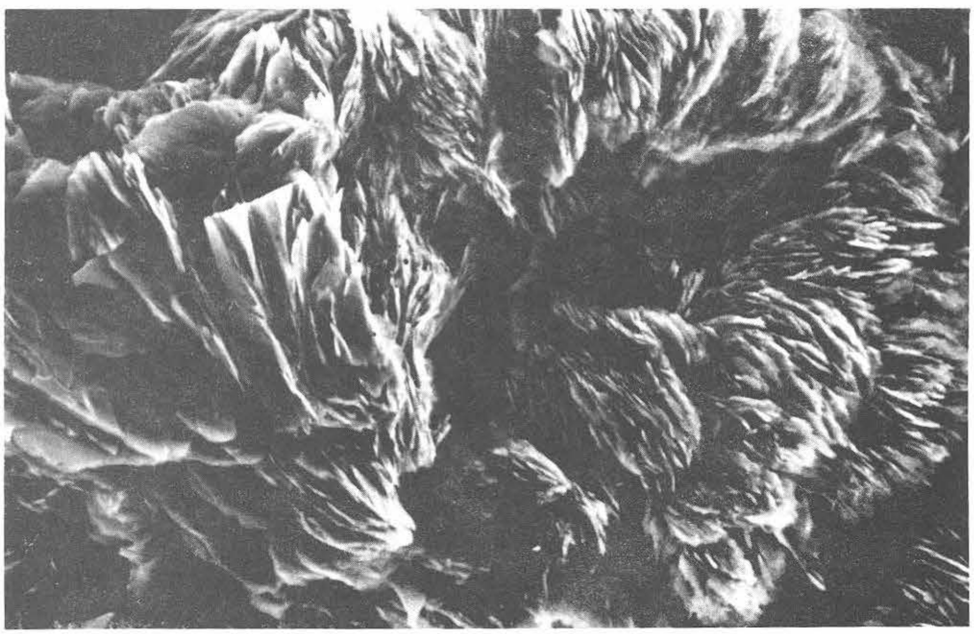


Figure 21. Creamy white aggregates of iron-rich talc crystals. The magnification of the SEM photomicrograph is about 700x. Photo by G. Robinson.

PARAGENESIS

The origin of this deposit now appears to have been quite complex, with a number of stages spanning an unknown period of geologic history. Before presenting a summary paragenesis, I shall review several topics which bear directly on this deposit's origin.

1. The deposit sits on a boundary between Grenville marble and granite gneiss. Smyth (1894b) and Buddington (1934) have shown that the

bulk of the chloritized rock in the ore body is an altered granite. Smyth suggested that the chloritization resulted from the action of iron-rich, acidic solutions arising from surficial weathering of pyritic gneisses or schists.

2. The basic structure of the ore itself is stalactitic, such as occurs from the precipitation of iron oxyhydroxides from aqueous solution at or near the surface. Furthermore, the voids or cavities in this basic stalactitic structure although subsequently mineralized, were never subjected to sufficient pressure to collapse them.

3. Thin sections of the overlying Potsdam sandstone clearly show that, contrary to the conclusions reached by Smyth from field observations, the phase of mineralization which included specular hematite, quartz, and stilpnomelane occurred after the deposition of the Potsdam. The bottom of the overlying Potsdam sandstone was mineralized from below. We believe Smyth mistakenly identified mineralized boulders in the Potsdam as pieces of detrital ore. Thin sections clearly show this is not the case.

4. Fluid inclusion studies of the quartz crystals suggest that the quartz and specular hematite was deposited at the somewhat elevated temperature of 150°C,

5. An oxidation-reduction reaction occurred sometime after the deposition of the specular hematite. These crystals have been largely reduced to magnetite. Furthermore, the reduction appears to have proceeded from the cavities outward into the walls. The major portion of the massive hematite was not affected. Both stilpnomelane and talc can be implicated in this reaction. Analyses of the degree of hematite reduction and the color and ferric/ferrous ratios of associated stilpnomelane show a clear correlation. Generally the greener stilpnomelane has $Fe_2O_3/FeO = 2.5$ and is associated with magnetite-poor vugs. The golden brown stilpnomelane has $Fe_2O_3/FeO = 16.3$ and is associated with magnetite-rich vugs. It is generally thought that stilpnomelane forms as stilpnomelane-Fe(II) and may be subsequently oxidized to stilpnomelane-Fe(III) (Zen, 1960; Brown, 1967, 1971; Hutton, 1938). Thus the association of stilpnomelane rich in ferric iron with vugs rich in magnetite suggests the coupled redox reaction:

hematite + Fe(II) stilpnomelane » magnetite + Fe (III) stilpnomelane + O_2

Likewise, our finding that in the iron-rich talc, the ferric iron substitutes for magnesium rather than for silicon suggests that the talc formed as ferroan talc and was subsequently partially oxidized to ferrian talc.

6. Although pecoraite is presently much more common on the dumps than it is in collections assembled while the mine was operating, some "old time" specimens show the alteration of millerite to pecoraite, and some specimens have been found with pecoraite pseudomorphs after millerite completely embedded in calcite. Thus, while the formation of pecoraite represents a late stage weathering, it may not exclusively be a phenomenon occurring on the dumps.

With this introduction, the paragenesis of the minerals can now be considered in three parts: 1) the minerals remaining from the original marble; 2) those remaining from the original granite gneiss; and 3) those formed during and after the deposition of the ore. Minerals remaining from the original marble include flakes of graphite and phlogopite found in a calc-silicate rock that has been partially replaced by hematite. Minerals remaining from the original granite gneiss are found as grains and plates in the green, chloritic rock and include quartz, feldspar, phlogopite, and minor tourmaline. The paragenesis of the minerals found in the cavities in the ore has been determined from examination of numerous hand specimens and while not exhaustive, should be indicative of the general pattern (Fig. 22). The observed sequence of mineral formation coupled with the observations above suggest the following five paragenetic stages: 1) stalactitic deposition; 2) ferric mineral deposition; 3) ferrous mineral deposition; 4) coupled oxidation-reduction; and 5) late deposition and weathering.

Stage 1 - Stalactitic Deposition: It is obvious that open spaces must have existed in order to have formed the crystals observed in the vugs. These voids may have resulted from the attack of the marble by acidic solution derived from decomposing sulfides or meteoric water with dissolved carbon dioxide. Some may simply have existed as fractures. The change in volume resulting from the chloritization process may have produced openings in or around the altered gneiss. Regardless of their origin, the fact remains that open spaces must have been present for solutions to enter and deposit botryoidal and stalactitic hematite. Ocherous and botryoidal hematite is a universal feature of cavity walls, and thin sections of stalactitic aggregates of later formed minerals often show a core of this hematite phase. The original precipitates may have been amorphous ferric hydroxides or oxyhydroxides that gradually recrystallized as hematite, possibly after passing through an intermediate goethite phase. This stage of deposition may well have occurred at or near surface conditions in late Precambrian time.

Stage 2 - Ferric Mineral Deposition: The minerals of this stage include

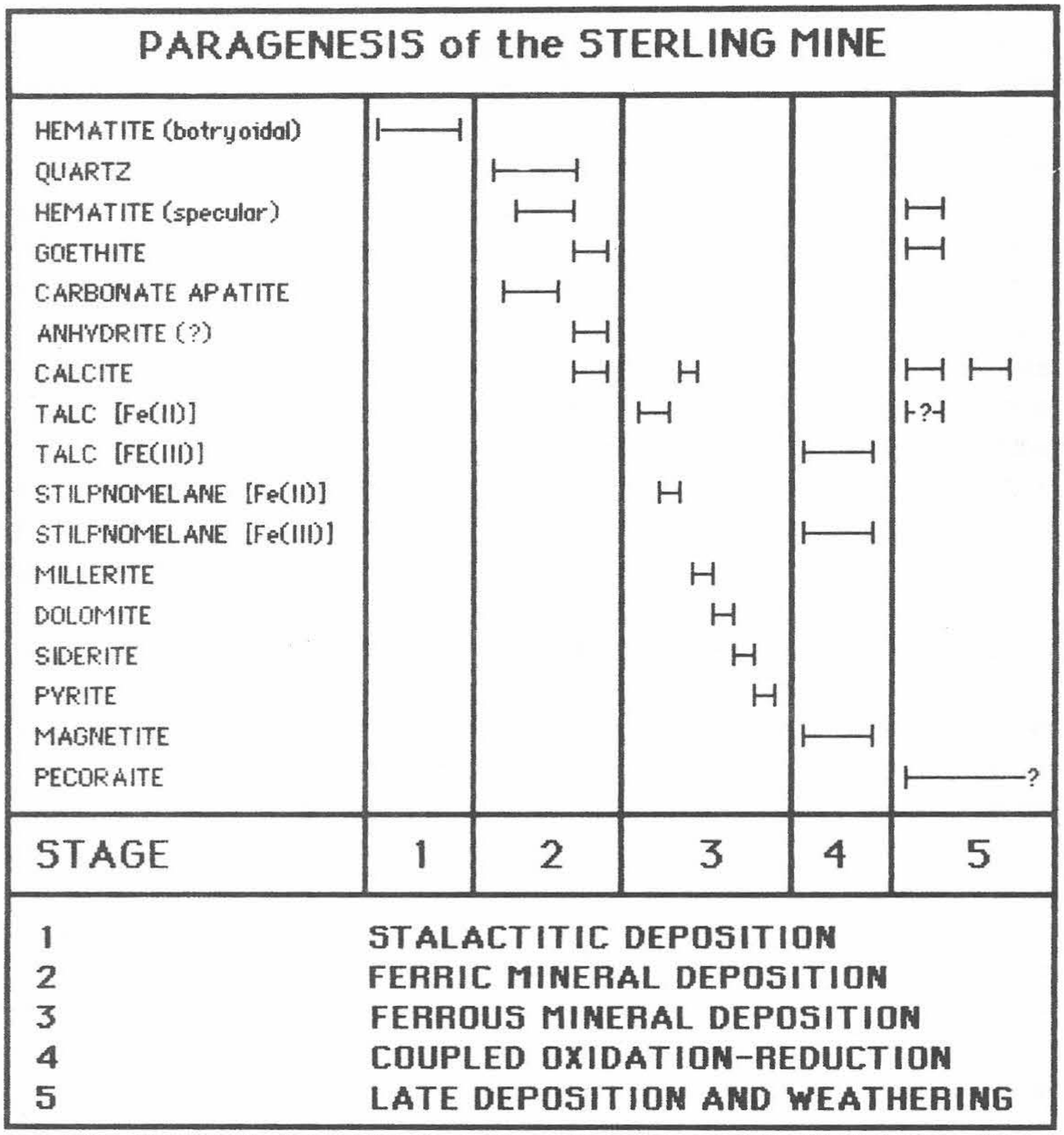


Figure 22. Paragenesis diagram for the major species occurring at the Sterling Mine. The vertical divisions between the stages represent unknown time periods.

quartz, specular hematite, carbonate apatite, crystallized goethite, calcite, and anhydrite(?). Fluid inclusion studies indicate an elevated temperature (150°C) during this phase of mineralization. The first two minerals to form were hematite and quartz. The walls of many cavities are a banded mixture of these two minerals. Often hematite crystals are embedded in quartz crystals and the quartz is encrusted with hematite, suggesting a simultaneous or alternating crystallization. The few specimens of carbonate apatite that were available for study suggest that this mineral also formed at this time. The next mineral to form appears to have been goethite, since delicate golden sprays of acicular crystals often lay partially within the terminations of quartz crystals. The anhydrite (?) laths and calcite scalenohedra that formed during this stage were subsequently dissolved away leaving crystals molds in later formed minerals. Since these minerals certainly did not form under surface conditions, considerable time may have elapsed between stages 1 and 2.

Stage 3 - Ferrous Mineral Deposition: After the final depositon of quartz, specular hematite, and goethite, a new association of minerals began to be deposited, starting with talc and stilpnomelane. As discussed above, there is good reason to believe that both these silicates formed initially as ferrous iron species, marking the beginning of a new stage in the paragenesis. After stilpnomelane, a second generation of calcite of predominantly rhombohedral habit formed, followed in order by millerite, dolomite, siderite, and pyrite. Thin sections of the overlying Potsdam sandstone clearly show replacement by hematite, carbonates, and stilpnomelane, which strongly suggests that stages 2 and 3 occurred in the Paleozoic after the formation of the sandstone.

Stage 4 - Coupled Oxidation-Reduction: The nearly universal reduction of the hematite crystals in the cavities to magnetite appears to have been a major event in the paragenesis. The evidence presented above suggests that a coupled oxidation-reduction reaction occurred in the open cavities between hematite and stilpnomelane (and talc). This reaction may have been triggered by an elevation in temperature and pressure subsequent to the deposition of the minerals involved, as stilpnomelane is a typical indicator of low grade metamorphic conditions. In the conversion of stilpnomelane rich in ferrous iron to stilpnomelane rich in ferric iron, electrons were made available to reduce some of the ferric iron in hematite to ferrous iron and produce magnetite. We have placed stage 4 at this point in the paragenesis based upon the observations that in specimens where the cavities were completely filled with calcite from stage 5, the hematite was reduced to magnetite, but in a few specimens where the cavities were completely filled with dolomite from stage 3, the hematite was not reduced.

Stage 5 - Late Deposition and Weathering: Late in the paragenesis, a third generation of "nailhead" calcite was deposited. Some finely crystallized talc also appears to have formed very late in the sequence. Small amounts of hematite, acicular goethite, and aragonite were also deposited. The remainder of events in this stage involve the alteration of earlier minerals. Malachite formed from chalcopyrite; pecoraite, from millerite; and goethite, from pyrite, magnetite, and siderite. These alterations can take place under atmospheric conditions and may still be occurring in the dumps today. At some point or points in this final stage, some vugs were completely filled with calcite, or more rarely with calcite and gypsum. As noted above, pecoraite pseudomorphs after millerite crystals have been found within this calcite, as have magnetite pseudomorphs after hematite crystals.

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PLEISTOCENE GEOLOGY, GROUNDWATER, AND LAND USES OF THE TUG HILL, AND BRIDGEWATER FLATS AQUIFERS, ONEIDA COUNTY, NEW YORK: A CASE STUDY

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INTRODUCTION

Due to new national interest in groundwater resources, increased incidents of groundwater pollution, and the lack of an adequate groundwater data base, the Oneida County Planning Department Environmental Management Council in conjunction with Syracuse University and the United States Geological Survey (U.S.G.S.) initiated an evaluation of Oneida County's groundwater resources.

Oneida County (figure 1) is situated within five of the State's major physiographic provinces (figure 2). They are the Adirondack Mountains, the Hudson Mohawk lowlands, the Appalachian plateau, the Tug Hill plateau and the Ontario Lake plain. The present land configuration is a direct result of prior glaciations. The rolling hills, surficial drainage, and valleys are evidence of late Wisconsin glaciation, extensive glaciofluvial erosion, lake impoundment and mass-wasting.

Oneida County is characterized by a dominance of glacial aquifers in a network of transecting glacial troughs with moderate relief and several hundred feet of valley fill. These aquifers are comprised of confined and unconfined sand and gravel glacio- fluvial deposits, glacio-lacustrine deposits and post glacial alluvial deposits.

The groundwater resources of Oneida County have only been slightly studied (Fairchild, 1909; Halber, 1962; Heath, 1964; Kantrowitz, 1970), geologic information has been well documented (Miller, 1909; Kay, 1953; Dale, 1963; Street, 1963; Cadwell, 1972; Wright, 1972; Krall, 1977; Chambers, 1978; Jordan, 1978; Antonetti, 1982; Franzi, 1983).

Most groundwater studies in glaciated areas have been either 'site-specific' or greatly generalized. Neither the generalized and qualitative studies nor the more detailed studies effectively assist planning with respect to the control of groundwater contamination and overdevelopment. The former is too qualitative to accurately locate groundwater recharge and discharge areas, whereas the latter often is too site-specific to have direct transfer value. The research being conducted is a "case study" designed to provide the types of hydrogeological information necessary to evaluate the groundwater resources in glaciated terrains.

A better determination of the types of groundwater flow systems that develop in glaciated areas will provide future planners with information essential to various land use activities such as the location of additional landfills, bulk storage sites, etc.

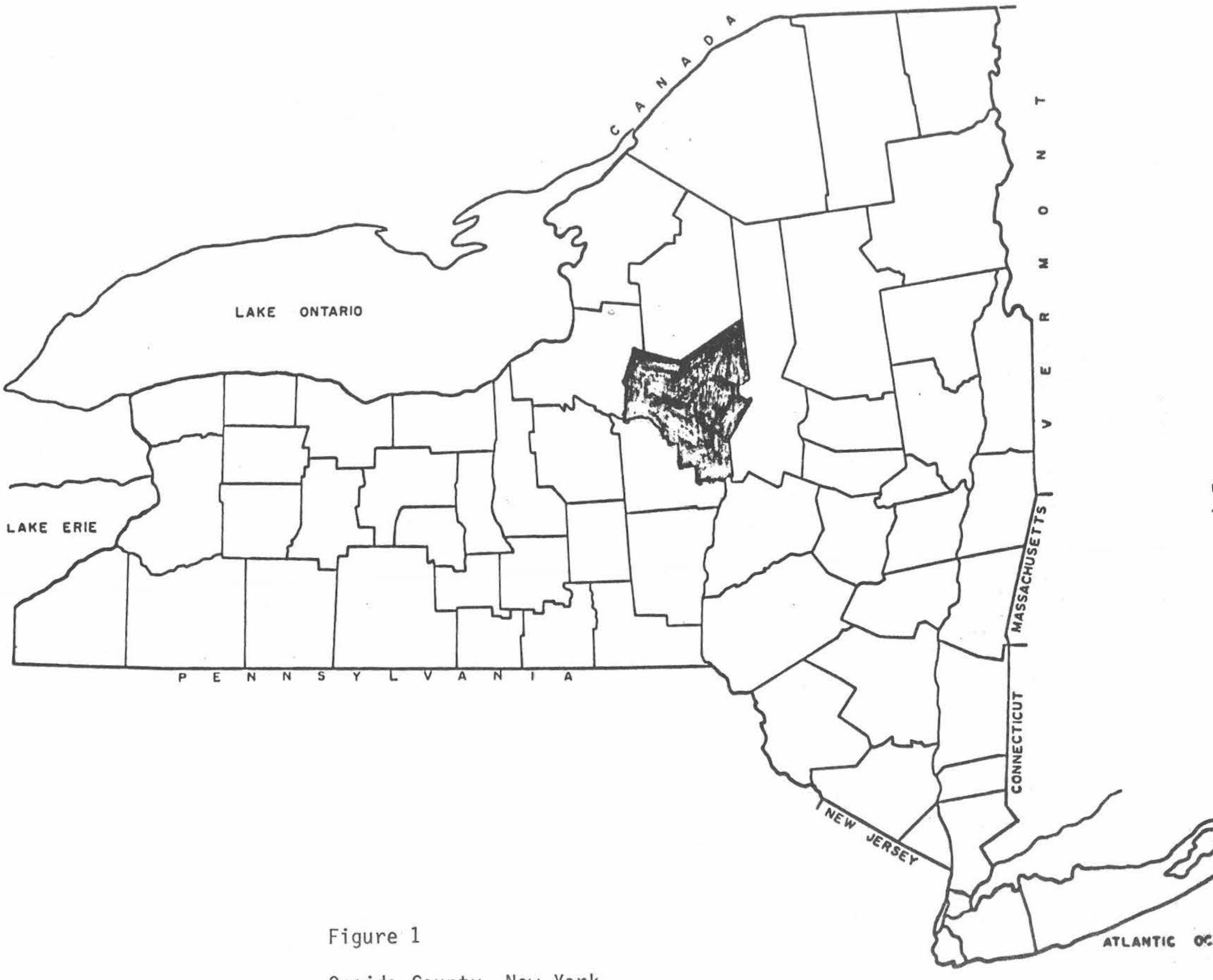

Figure 1

Oneida County, New York

This study qualitatively evaluates the hydrogeology of Oneida County, New York, and simulates, through the use of computer models, the major aquifer systems to determine the main parameters controlling groundwater movement.

The probable boundaries for the major glacial aquifers is being determined by literature evaluation of the glacial geology and interpretation of published and unpublished data from test borings and water-well records. Maps will be prepared of transmissivity, specific yield, the water table, and the potentiometric surfaces of confined aquifers.

Selection of some aquifer boundaries and the determination of the types of groundwater flow systems (regional, local) will be done by numerical simulation by finite-difference computer models. The models (Trescott and Larson, 1976) will be used as a tool to 'test' the sensitivity of groundwater flow systems to aquifer boundary conditions and to determine estimates of material parameters such as hydraulic conductivity.

Twelve aquifers have been tentatively identified in Oneida County but geologic and hydrogeologic data has been mapped and compiled only for the Tug Hill Aquifer. This, and the Bridgewater Flats Aquifer are typical examples of valley fill systems. The studies of these two aquifers is a joint effort of the Planning Department, Syracuse University, and the U.S.G.S.

Tug Hill Aquifer

Setting

The Tug Hill Aquifer occurs along the west and south sides of the Central Tug Hill area (figure 3). The aquifer extends 55 miles from Adams Center (Jefferson County) through Western Oswego County and into Oneida County, where it joins the Rome Sand Plains beyond McConnellsville and Blossvale (figure 4). The aquifer is a typical example of a valley fill aquifer. West Branch Fish Creek runs NW-SE through the center of the valley.

Surficial Geology

Field data indicate that the valley is filled with sand and gravel which grades southward into lacustrine sands. The northern part of the aquifer is comprised of beach sand and gravel and localized pockets of outwash material. Beach sand and gravels can be seen in an excavation pit cut into a beach ridge around Adams Center. Proceeding south through the valley, the beach sand and gravel grade into coarse outwash material. Along the valley walls are extensive deposits of kame terraces, which are believed to influence the recharge into the valley material. Significant quantities of coarse granular sediments comprising a kame moraine complex are found in the reach from Williamstown to Westdale. From Williamstown to Camden, several eskers trending NW-SE, are found along the valley floor. The outwash deposits grade into lacustrine deposits within this area. Proceeding south from Camden to the terminus of the aquifer around McConnellsville, the aquifer is comprised of lacustrine sands. Several well records in the southern portion of the aquifer indicate that sand and gravel overlies till and bedrock at a depth ranging from 90-130 feet. An extensive outwash fan deposited by the Mad River is found at its confluence with the West Branch of Fish Creek Valley. Additional investigation is needed to determine the depth to which this fan material extends.

Hydrology

The Tug Hill Aquifer is recharged by the substantial precipitation resulting from prevailing westerly winds sweeping across Lake Ontario and then encountering the rising Tug Hill Plateau. Recharge occurs from precipitation that falls on the aquifer and seeps to the water table as well as from seepage out of streams crossing or following the aquifer formation.

Several municipal water systems tap the aquifer to supply 6,900 people at a rate of about 1.3 million gallons per day. About the same number of people tap the aquifer through individual wells. The new DEC Salmonid Hatchery at Altmar pumps at peak load 2.3 million gallons per day, and Schoeller Technical Paper pumps on the average 1.5 million gallons per day.

Aquifer yield varies along the length of the aquifer. Kantrowitz, 1970; estimated aquifer yields to range from 2-4 million gallons per day to as much as 12-20 million gallons per day. Two very productive reaches (12-20 mgd) are from Kasoag to south of Westdale and within the Richland quadrangle.

In general, depth to the water table is approximately 20 feet, but, the water table when in close proximity to streams is found to be 2-8 feet below land surface. Throughout the aquifer many shallow dug wells tap the groundwater resource to supply potable water for individual home use. Some reported depths to the water table were as great as 35 feet and 47 feet. Many springs flow year-round along the valley margins.

Further investigation is needed for the Tug Hill Aquifer. The interaction between the West Branch Fish Creek and the aquifer needs to be determined. For many parts of the aquifer, data on saturated thickness is missing. An appraisal of groundwater quality needs to be conducted.

This information will be disseminated to the public for educational purposes, and to State and local bodies of government in order to develop long-term planning and management.

A number of municipal sewage treatment plants discharge wastewater directly into Fish Creek. In addition, several industrial and sanitary disposal sites are located on top of the aquifer. An inventory of the overlying land use needs to be conducted and a variety of strategies are needed to protect these vulnerable, hidden resources from overuse and contamination.

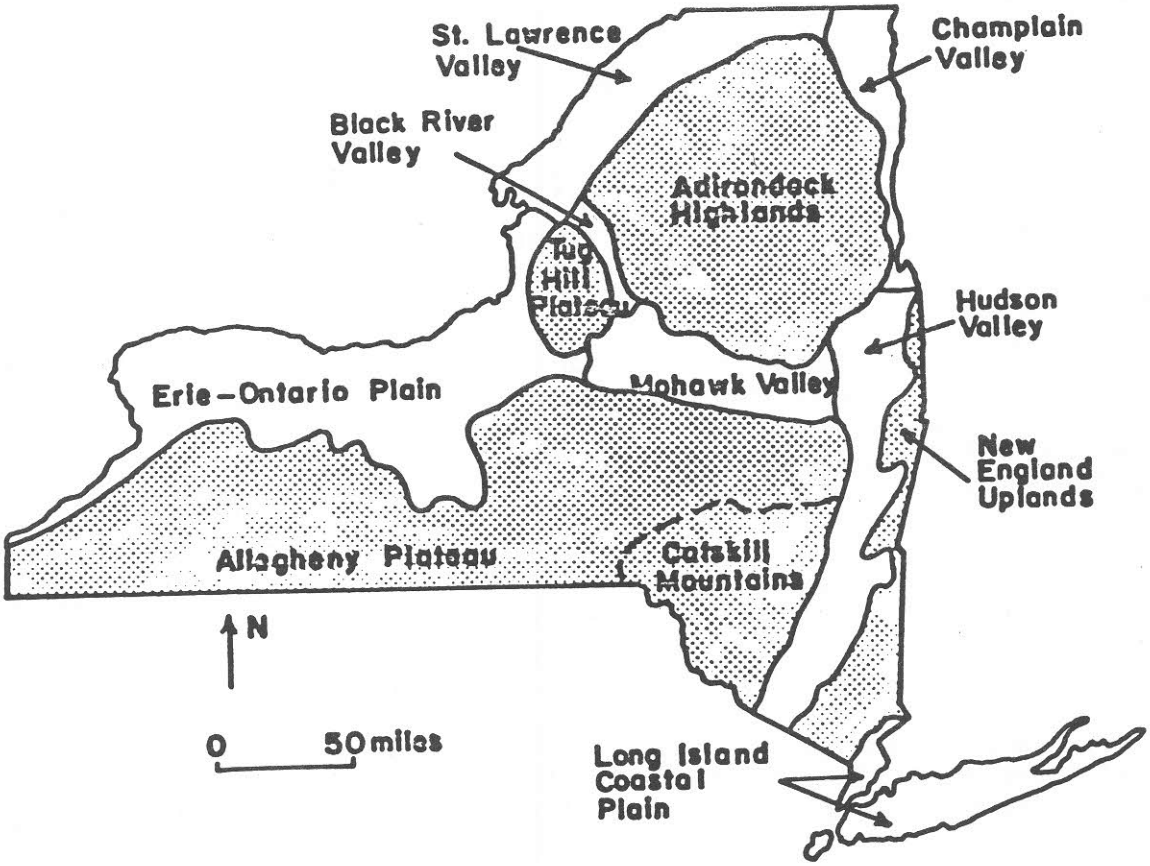

Figure 2. PHYSIOGRAPHIC PROVINCES of New York State (adapted from Olson, 1969).

Figure 3. Tug Hill Plateau

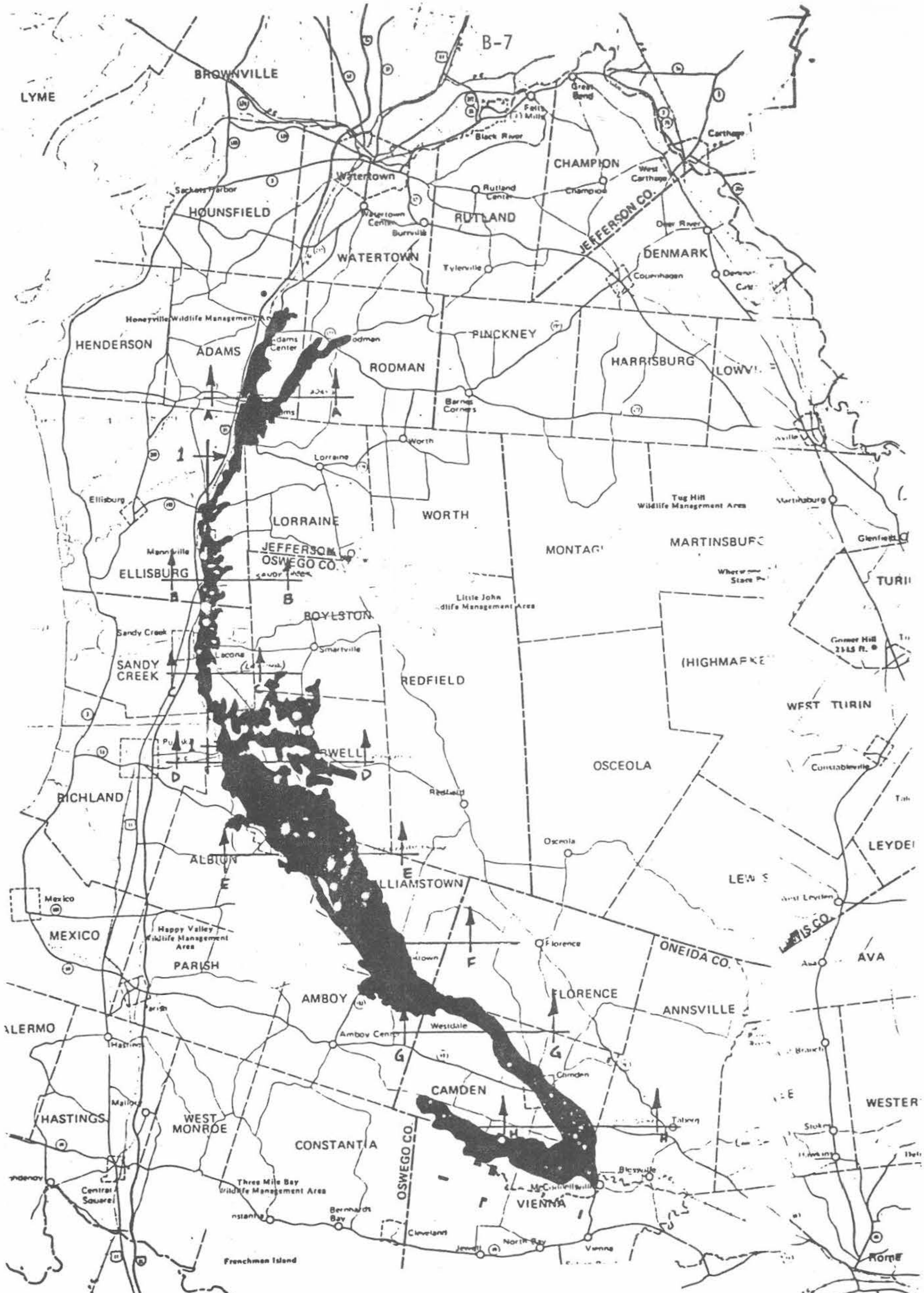

Figure 4

Tug Hill Aquifer

The Bridgewater Flats Aquifer

Setting

The Bridgewater Flats Valley, in southeastern Oneida County, is a typical example of a headwater through-valley (figure 5). The valley is approximately four miles long and one mile wide, at its head, and one-half mile wide at its outlet. It gradually slopes to the south about 60 feet from its head to its outlet. The valley floor slopes to the east, where it has been incised by the West Branch Unadilla River. Drainage north of the divide is to Sauquoit Creek. Drainage to the south is to the West Branch Unadilla River of the Susquehanna River Basin.

Surficial Geology

An analysis of the surficial geology and well records indicate that the valley is partially filled by sandy gravel which thins southward. Gravel exposures can be seen in northern Bridgewater Flats. The greatest known thickness of gravel, 47 feet, is located west of Babcock Hill close to the drainage divide. The surficial sandy gravel deposits grade southward into coarse black sand, silt and clay. Buried gravel is indicated in well logs at well locations near the southern end of the valley. Well logs for the southern end of the valley showed gravel overlain by 16 feet of clay and 22 feet of coarse black sand; another log showed 40 feet of sand. One well was drilled 394 feet through fine sand and bottomed in gravel.

Surficial clay deposits appear sporadically in the valley. The clay is considered to be lacustrine deposits laid down in a proglacial lake over outwash gravels. The southern part of the valley contains an aquifer under confined conditions. It is assumed that the buried gravel in the southern end of the valley is continuous with the surficial gravel in the northern end.

Hydrology

The estimated saturated thickness of the aquifer ranges from 10 to 40 feet at depths less than 50 feet (Hollyday, 1969). Estimated well yields (Hollyday, 1969) could range from 550 to 2,400 gal/min with a median of 1,000 gal/min. In general, the water table lies within a few feet of the land surface.

The aquifer is recharged by direct precipitation and runoff. Two major tributaries on the west side of the valley contribute recharge to the aquifer throughout the year. Kame deposits on the west side of the valley allow for very rapid infiltration of precipitation to recharge the valley deposits. The east side of the valley is underlain by a bedrock bench. Unconsolidated material covering the bedrock are mixed ice - contact and till deposits.

Land Use

The predominant land uses in the Bridgewater Flats valley are dairy farming, commercial potato and vegetable farming, cash cropping and corn

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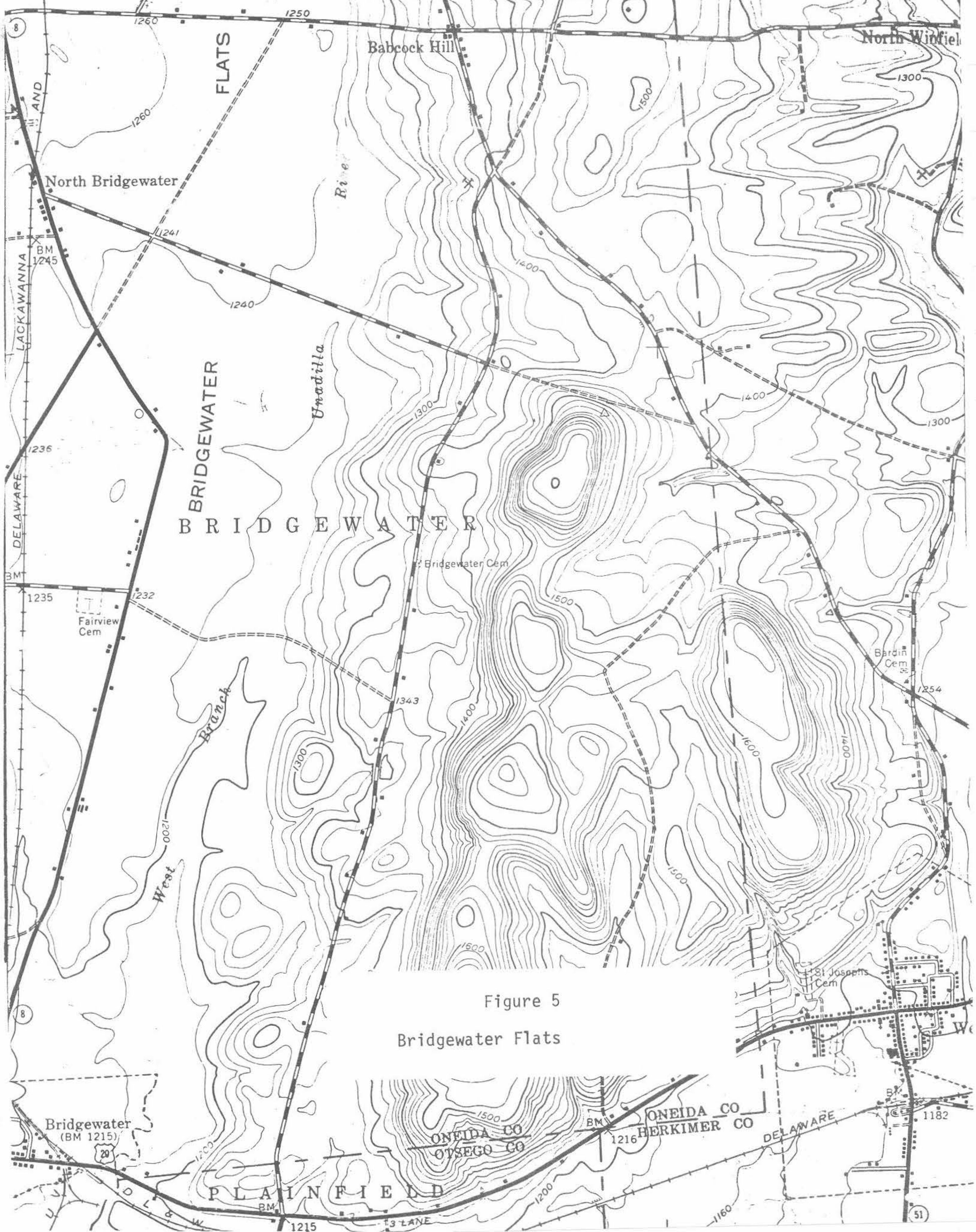

Figure 5

Bridgewater Flats

plantings for grain and silage. Commercial fertilizers and manure are applied to the soils for maximizing crop production. Therefore, nitrate loading to the groundwater may adversely impact water quality. However, at the present time, there is not extensive residential development. Homeowners, however, utilize individual water well supplies and have conventional on-site septic tank systems. The greatest potential for groundwater contamination is from the development of mobile home parks which utilize on-site conventional septic tank systems. Evaluation of adequate soil conditions and construction of alternative septic tanks and leach fields will minimize the risk of groundwater pollution from nitrates.

Further investigation on the Bridgewater Flats aquifer is being conducted. The degree to which the aquifer is connected to the West Branch Unadilla River must be determined. Recharge areas must be adequately defined, especially recharge to the confined aquifer in the southern end of the valley. Recharge rates and natural groundwater discharge areas and rates need to be determined in order to minimize the risk of overdevelopment in the future. This information will also reduce the number of poorly located, deep, costly wells. In general, the groundwater resources here have met the demands placed upon them, and appear to be ample to meet future needs.

DEGLACIATION AND CORRELATION OF ICE MARGINS, APPALACHIAN PLATEAU, NEW YORK*

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INTRODUCTION

The field trip area (Fig. 1) is located in the Susquehanna River Drainage Basin and includes parts of the Unadilla River, Wharton, Oaks, and Butternut Creek valleys. Local relief of the bedrock-drift interface reaches 1200 ft, while surface elevations range between 1050 and 2350 ft. The bedrock is predominantly Devonian sandstone, siltstone, shale (Hamilton Group) and limestone (Onondaga and Helderberg Groups). The purpose of this report, and associated field trip, is to establish and illustrate criteria for the recognition and correlation of ice margin positions in the upper Susquehanna River Drainage Basin.

This part of New York State was no doubt subjected to repeated glaciations during the Pleistocene Epoch. Multiple glaciations are suggested in the subsurface by interbedded diamictons (tills) and outwash gravels (Randall, 1972). However, the abundant glacial deposits within the field trip area formed during the advance and retreat of the Late Woodfordian ice sheet.

Six ice-marginal positions were held during Late Woodfordian time. They are, from south to north, the Wells Bridge, Oneonta, New Berlin, Cassville-Cooperstown, Middleburg, and Valley Heads ice margins. The Wells Bridge, Oneonta, and New Berlin ice margins were established during what is inferred to have been a semi-continuous retreat of the glacier. The glacier readvanced to form the Cassville-Cooperstown margin. The Middleburg ice margin developed during subsequent retreat from the Cassville-Cooperstown position. The final event within the field trip area was a glacial readvance that deposited the Valley Heads Moraine.

TOPOGRAPHIC CONTROL OF ICE-MARGINAL POSITIONS

The dissected Appalachian Plateau has a bedrock relief (exclusive of overburden) on the order of 600 to 700 ft, and in some areas as much as 900-1,200 ft. This is greater than most areas in the U. S. that were

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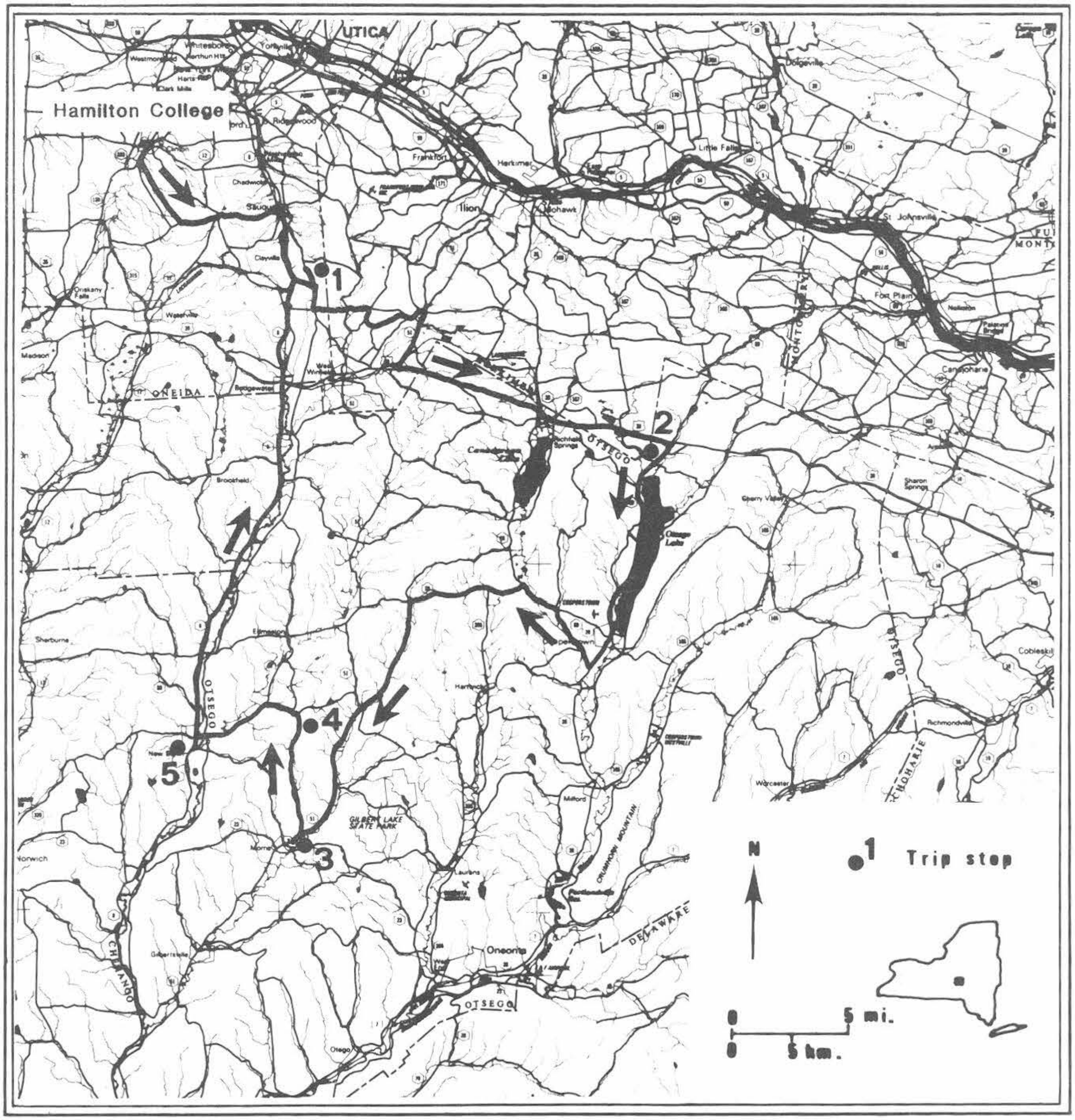

Figure 1. General field trip area.

subjected to continental glaciation and can be recognized to have influenced ice-marginal positions on a local scale. Cadwell (1972, 1973b, and 1978) describes the formation of ice lobes as glacier salients extending downvalley from the common ice margin. Fleisher (1983) further developed this concept by relating the influence of various topographic and ice-marginal configurations to depositional landform distribution.

The upper Susquehanna Drainage consists of major SSW flowing tributary streams that join the SW flowing Susquehanna to form an asymmetric drainage pattern. The regional ice flow direction was sub-parallel to the major tributary streams, thereby preferentially enlarging their valleys and increasing their relief. During deglaciation the regional ice margin trend was perpendicular to these tributary valleys and diagonal or sub-parallel to the main Susquehanna Valley. As a result, flow within the thinned ice margin adjusted to local topographic conditions. Figure 2 illustrates several possible configurations of ice-marginal positions relative to local valley trends. The location and distribution of depositional landforms commonly formed along the ice margin also are shown. The concept of topography (significant relief and valley trend) influencing local ice-marginal deposition is important for the recognition and correlation of these positions.

The major tributary valleys were not equally enlarged by the erosional effects of concordant ice flow. The larger ones probably were open to the north (hanging valleys) prior to glaciation as the result of earlier capture of the Susquehanna headwaters through headward growth of the Mohawk system. These lacked headward upland divides, thereby permitting easy access to glaciers, with subsequent valley enlargement and through-valley enhancement. Similar valley enlargement would not have occurred in valleys with headward upland divides (non-captured valleys). As a result, the longitudinal profiles of SSW flowing tributaries maintain significantly lower headward elevations in through-valleys, whereas in all non-through valleys the profiles rise to meet an upland divide.

This contrast in headward elevations had a major effect on determining which valleys could sustain active ice flow during deglaciation. As illustrated in Figure 3,the receding ice margin within a non-through valley creates thinning on the divide. This eventually reduces nourishment to the terminus allowing a negative ice budget to develop, with massive stagnation and downwasting in the valley. The resulting dead-ice environment would favor the formation of eskers, ablation moraines and local glaciolacustrine deposition resulting in deltaic outwash, kame-deltas and hanging deltas. Typical valley train and massive outwash accumulation is precluded here by effective detachment of the stagnant lobe from the hydrologic system of the main ice mass (Fleisher, 1984a). This situation is well demonstrated by the landforms and stratigraphy of the Butternut Valley.

CRITERIA FOR THE RECOGNITION OF ICE-MARGINAL POSITIONS

Deglaciation of the eastern Appalachian Plateau took place during a time-transgressive retreat between circa 16,500 and 14,500 yrs BP

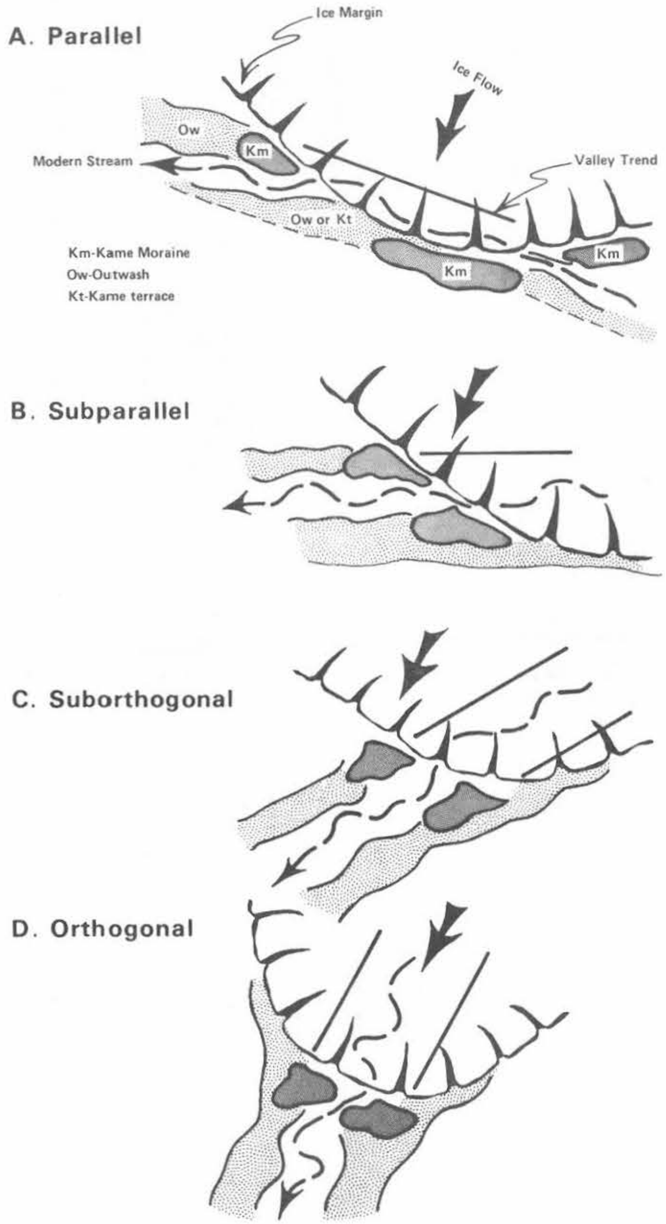

Figure 2. Factors controlling the distribution of ice-marginal landforms. The occurrence of landforms is influenced by the orientation of the ice margin relative to the valley trend. (from Fleisher, 1983, 1984a, 1984b)

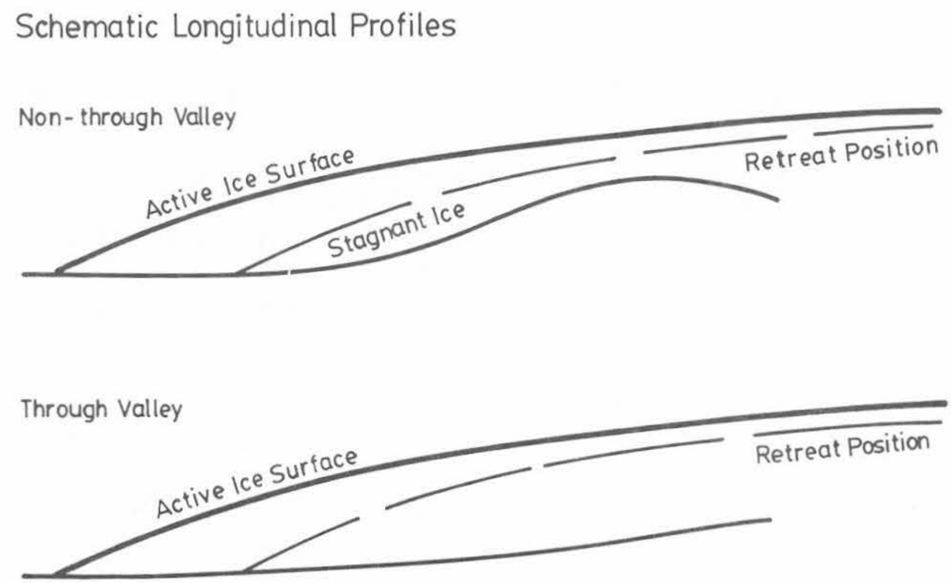

Figure 3. Schematic diagram of non-through valley and through valley longitudinal profiles and retreating glacier surface. (from Fleisher 1984a, 1984b)

as documented by Cadwell (1973a), Coates (1974), Fullerton (1980), and Fleisher (1983). At any given time the glacier terminus was draped across the landscape along a single, continuous trend.

The term "ice margin" refers to the location of the glacier margin during episodes of retreat (as in recessional positions) or readvance (as in end positions). Landforms representing the ice margin are both depositional and erosional, and are presumed to form synchronously along any given margin. These landforms appear in two distinct topographic facies (Fleisher 1984a, 1984b). The recognition of ice-marginal landforms and their short-distance correlation establishes positions held during glacier retreat or readvance. These ice-marginal positions are used in regional correlation.

Valley Floor Facies

1. Kame moraines and associated outwash - Kame moraines consist of hummocky, kame and kettle topography that spans the full valley width and is dissected by modern drainage. They generally stand 80 to 100 ft in relief above the common valley floor and may have served to dam the valley for a period of time following glacier retreat. Kame moraines consist primarily of gravel (generally coarse, displaying all degrees of sorting and stratification), but diamict may also be present. They form along the margin of actively flowing ice by glacial-glaciofluvial accumulation in an ice-contact environment. Massive quantities of outwash are associated with kame moraines and form valley trains or delta terraces. The Wells Bridge Moraine across the Susquehanna Valley between Oneonta and Sidney is a good example and is illustrated in Figure 4 (Fleisher 1977, 1984b).

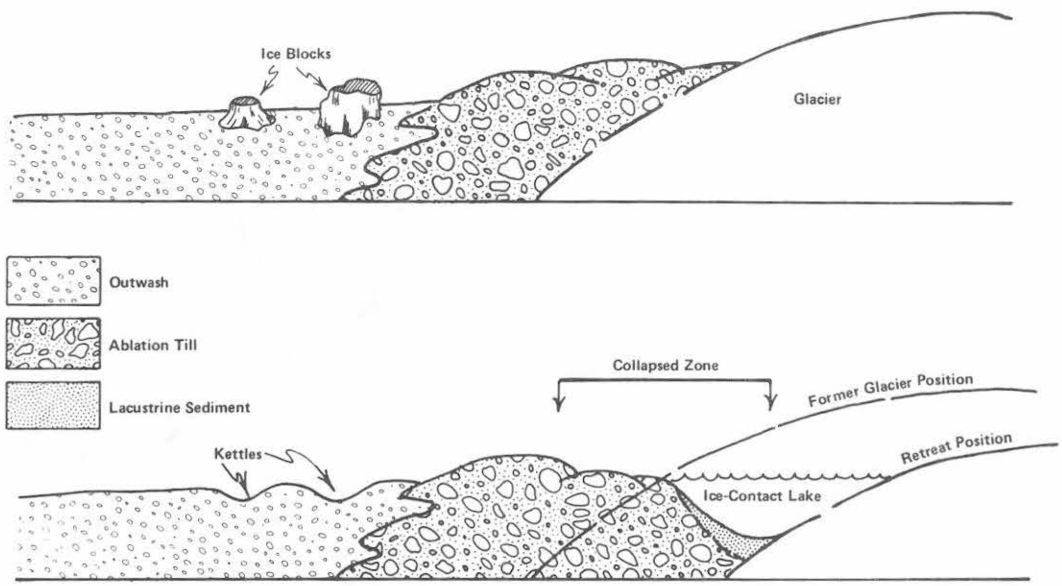

Figure 4. Longitudinal cross section of Wells Bridge kame moraine and associated outwash (from Fleisher 1977).

2. Pitted outwash and dissected valley train - While these landforms commonly are found in association with kame moraines, they also occur independently. They are considered to be ice-marginal landforms for two primary reasons.
    a) Their surfaces grade upvalley to an elevation above the modern flood plain that could neither have been established during post-glacial deposition, nor could they be erosional remnants of a once more extensive valley train. Erosion to that degree would be inconsistent with landforms in other parts of the same valley. This indicates ice must have been present at that position, from which meltwater was discharged and outwash deposited (Fleisher 1983, 1984b).
    b) Kettles and other collapse features indicate deposition in contact with ice.

    Good examples are in the Bridgewater Flats valley (Stop 1), and in the Unadilla Creek valley at the confluence with Wharton Creek in the village of New Berlin (Stop 4).

3. Dead-ice sink - This is a landform and stratigraphic feature similar in origin to a kettle, but on a scale that occupies most of the valley floor width. Dead-ice sinks are recognized by two landform characteristics that are diagnostic when they occur together. They are,
    a) an anomolously wide flood plain with many tight meanders and/or very poor drainage.
    b) associated massive outwash (dissected valley train or delta terraces) in both the upvalley and downvalley directions (Fleisher 1983, 1984a, 1984b).

    When large masses of stagnant ice (dead-ice) become detached by glacier retreat and are buried in part or totally by outwash, subsequent melting will create a progressively developing void (sediment sink) that is kept full of lacustrine and/or fluvial sediment as it grows. Large, detached ice blocks were associated with some ice-margins, thereby making the dead-ice sink an ice-marginal landform. An example can be seen at the mouth of Wharton Creek, between New Berlin and Pittsfield, as shown in Figure 5

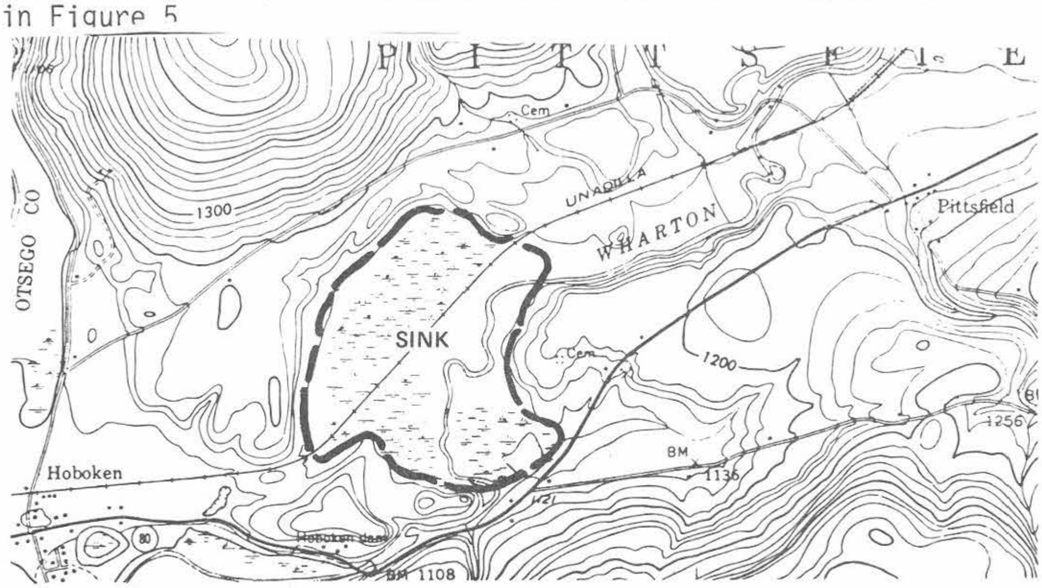

Figure 5. Topographic map expression of dead-ice sink (from Fleisher 1984a).

Valley Slope and Divide Facies

1. Isolated kames and ablation moraines - Slope wash, mass wasting and erosion by meltwater associated with ice on the divides tends to hinder the preservation of depositional landforms. Lodgement till is the most common overburden here, except for areas across which ice margins can be traced. The term "ablation moraine" is used to signify areas of limited hummocky topography, with only moderate to low relief. They appear as "patches" of morainic terrain and lack an association with typical outwash landforms. They are interpreted to indicate small scale areas of limited downwasting (Fleisher 1984b).

2. Underfit streams - These are low order tributaries (usually 1st, 2nd, or 3rd) or drainageways that occupy valleys which appear inordinately large for the size of the modern stream. They are interpreted to have developed through the benefit of meltwater discharge related to a nearby ice margin. They are often associated with discontinuous patches of ablation moraine. Examples are not common, but Brook Creek, a tributary of the Unadilla River north of New Berlin, is a particularly good example (Fleisher, 1984b).

3. Upland meltwater channels - These are not easily distinguished from glacially developed cols, but can be associated with an ice margin when found in association with meltwater gravels. They also will reflect a shape more commonly eroded by water rather than ice. An upland meltwater channel in the field trip area is the Cedarville col at Cedarville.

IDENTIFICATION OF RECESSIONAL AND READVANCE ICE MARGINS

The topographic expression of a recessional moraine (formed during retreat) is very similar to an end moraine (formed by readvance). They can be distinguished, however, on the basis of their stratigraphy. Evidence for a readvance includes lodgement (meltout) till over outwash, distortion of stratified drift by overriding ice, and compaction of fine-grained stratified drift by the weight of the overriding ice. Examples of distorted and compacted stratified drift will be examined at Stop 2 on the field trip. Surface exposure of these not only are uncommon but also ephemeral, therefore, stratigraphic evidence as revealed in water well logs and test borings becomes increasingly important.

The stratigraphic distinction between recessional and end moraines is illustrated in Figure 6. Basically, both consist of coarse sand and gravel, which grade laterally downvalley into a normal association with outwash (valley train or delta terraces) that, in this diagram, prograde into an ice-contact lake. Within a recessional moraine the outwash is interstratified with lacustrine silt and sand. However, readvance to an end moraine position would place till and outwash (both consisting of sand and gravel in the logs) stratigraphically above lacustrine sediment. Interstratification versus superposition of gravel and lake sediment is the diagnostic evidence for distinguishing recessional from end-moraine positions.

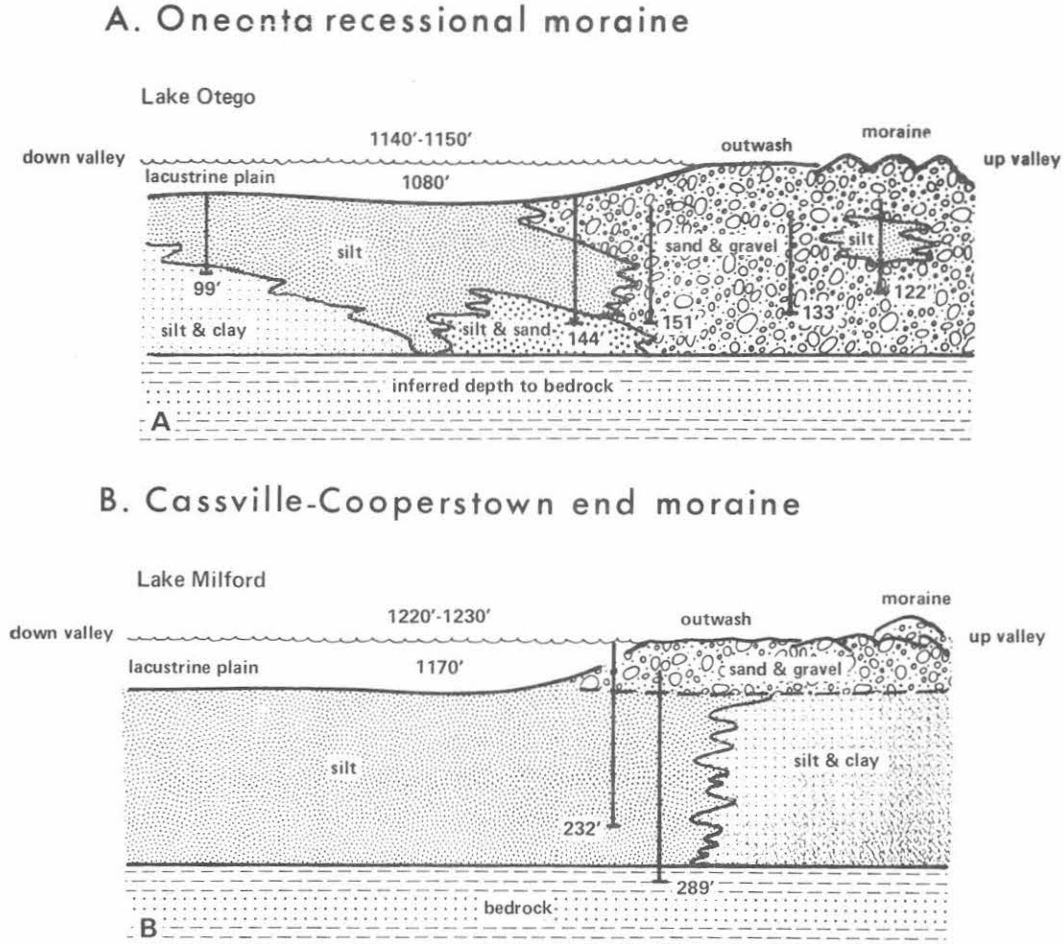

Figure 6. Stratigraphic distinction between recessional and end-moraine (from Fleisher 1983, 1984b).

A - outwash gravel interstratified with lake sediment.

B - outwash gravel overlies lake sediment.

The Oneonta and New Berlin ice margins (Fig. 7) are both recessional in origin. The thickness and stratigraphic relationship of the sediments at a recessional margin is controlled by the specific depositional environment. This is demonstrated by the interstratification of outwash with lacustrine sand, silt, and clay at the Oneonta margin illustrated in Figure 6A.

The Cassville-Cooperstown Moraine and the Valley Heads Moraine are end moraines, each forming during a readvance of the Woodfordian glacier. The depositional environment associated with the formation of the Cassville-Cooperstown ice margin is illustrated in Figure 6B. Thick outwash gravels (50-60 ft) were deposited above 150 ft of lacustrine silts and clays that are interpreted to continue upvalley beneath the moraine, thereby suggesting the glacier readvanced into a proglacial lake. Commonly well records are not available to complete the stratigraphic interpretation for an ice margin. Therefore, additional evidence must be used to reconstruct the environment of deposition associated with the ice margin.

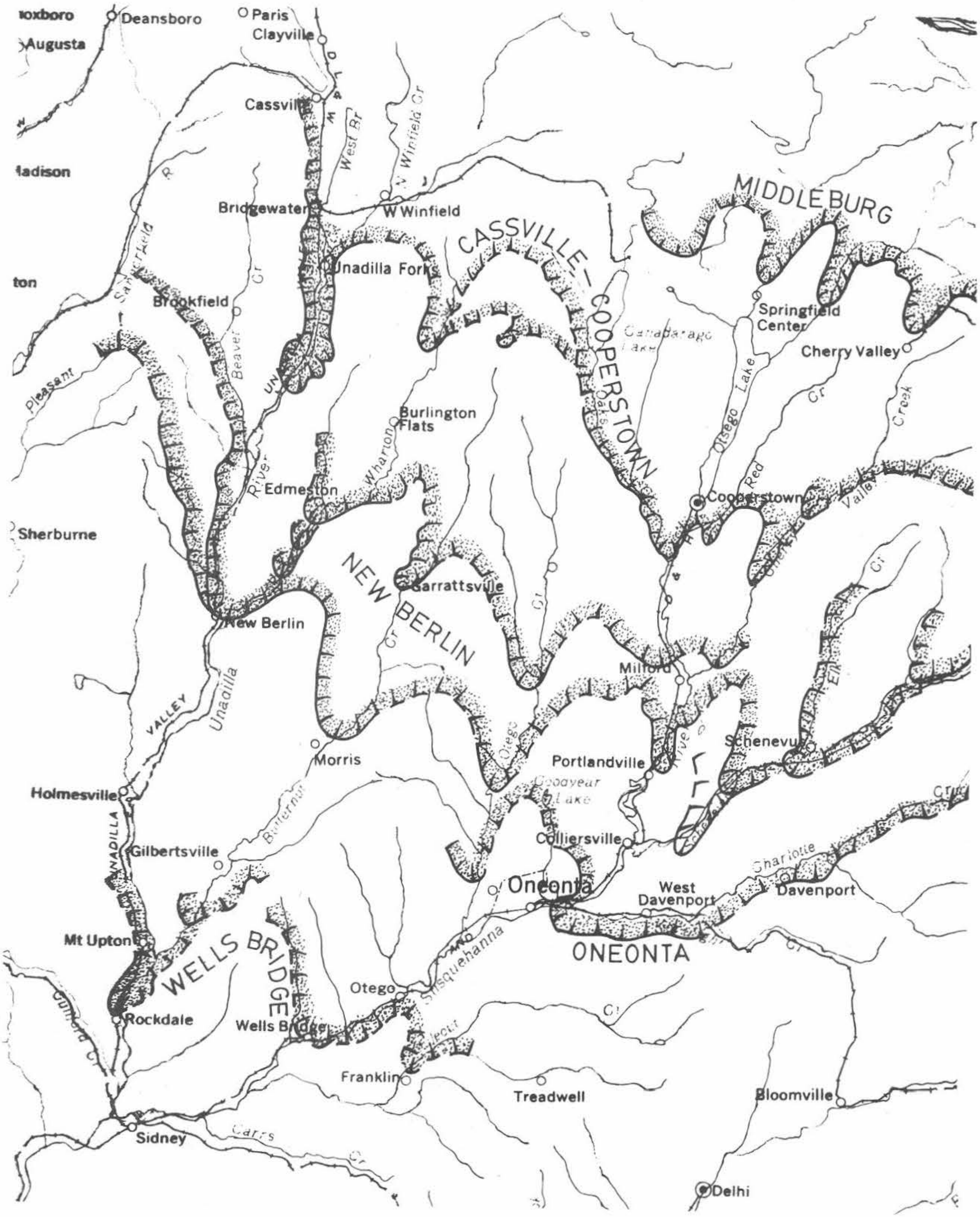

Figure 7. Ice-marginal positions in the upper Susquehanne Drainage (from Fleisher 1984a, 1984b).

While the field evidence does not clearly demonstrate a readvance, it does suggest that the mode of origin for these deposits is not retreating ice. For example, the existing well records in the Bridgewater Flats area, south of Clayville suggest thick outwash gravels (20-60 ft) with only thin layers of interbedded silt and clay (10 ft). Such a stratigraphic record does not illustrate evidence for a readvance as previously described. Therefore, additional criteria were developed to test the Valley Heads as a readvance.

Stratigraphic evidence in the form of stratified drift within kames adjacent to the Bridgewater Flats outwash plain permits relative age to be determined. Here the kames are generally 100-160 ft above the valley floor and were deposited during retreat of the Cassville-Cooperstown ice, while a marginal ice tongue (salient) occupied the valley. Meltwater streams flowed between the glacier and the adjacent valley walls forming kame terraces; kame deltas were formed in ponded areas. As deglaciation continued, the ice margin retreated several miles, leaving blocks of stagnant ice in the Bridgewater Flats Valley. These blocks melted prior to, or during, the glacier readvance to the Valley Heads marginal position. The thin layers of silt and clay, recorded in the well logs, were deposited in the kettle depressions during deposition of the sand and gravel outwash plain.

Previous correlations of the Valley Heads Moraine system were suggested by Chamberlin (1883), mapped in detail by Tarr (1905), and formally named by Fairchild (1932). The Valley Heads margin was traced to within several miles of Clayville by Fairchild (1912, 1932), Denny (1956), Denny and Lyford (1963), Fullerton (1971) and Cadwell (1973a). This correlation is consistent with our interpretation of the field trip area. The Valley Heads ice margin has been recognized in the headwaters of the Chenango River at Pratts Hollow to include a large kame moraine. Similar types of landforms are used to trace this margin eastward to Clayville.

CORRELATION OF ICE-MARGINAL POSITIONS

Six distinct ice-marginal positions have been recognized, correlated, and named. In chronologic order, from youngest to oldest, they are:

Valley Heads Moraine - this is the least well represented of all mapped margins, preserved only in the Bridgewater Flats valley north of Clayville. This was originally named by Fairchild (1932).

Middleburg margin - named for the ice marginal position originally mapped by LaFleur (1969); this margin is represented by a traceable moraine from Cobleskill to Richfield Springs.

Cassville-Cooperstown Moraine - named for the moraine originally mapped by Krall (1972) and Fleisher (1983, 1984b); this margin is represented by the most continuously traceable moraine and is the only one produced by a readvance.

New Berlin margin - a closely spaced, dual margin consisting of parallel trending marginal landforms, well expressed in both valley and slope facies of the Unadilla drainage (Fleisher 1983, 1984b).

Oneonta margin - named for the assemblage of landforms in the vicinity of Oneonta along the Susquehanna and Charlotte Creek valleys (Fleisher 1983, 1984b).

Wells Bridge margin - named for the well developed Wells Bridge Moraine in the Susquehanna Valley (Fleisher 1983, 1984b).

Figure 7 illustrates the ice margin positions and their correlation within the Susquehanna Drainage Basin. The correlation chart of Table 1 indicates the regional correlation of the ice margins.

TABLE 1 CORRELATION CHART
(modified from Fleisher 1983)

Susquehanna Drainage	Schoharie Drainage	Catskill Mountains
(Fleisher 1983, 1984b and present study)	(LaFleur, 1969)	(Cadwell, 1983)
Valley Heads Moraine		
Middleburg margin	Middleburg readvance	Middleburg margin
Cassville-Cooperstown Moraine	Middleburg readvance	Middleburg margin
New Berlin margin	Prattsville readvance	Glacial Lake Grand Gorge phase
Oneonta margin	Tannersville readvance	Wagon Wheel Gap margin
Wells Bridge margin		

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ROAD LOG

Miles	
0.0	Start at intersection of Rt. 412 & Rt. 12B (south), proceed south on Rt. 12B
1.0	Hard left turn on Grant Rd. (not Post St.)
1.6	Grant Rd. becomes South St. where Martin Rd. joins from the left
4.0	T-intersection, turn left
5.8	Intersection with Rt. 12, turn right (south) on Rt. 12
6.3	Enter Village of Paris
6.4	Bear left toward Sauquoit on Paris Green Rd. and then immediately turn left on Paris Hill Rd. (mileage 6.5)
9.3	Cross intersection with Oneida St. at STOP sign, Sauquoit
9.6	Turn right onto access ramp to Rt. 8 (south)
10.5	Highway begins to climb valley wall behind Valley Heads Moraine
12.4	Hummocky crest of moraine on both sides of road
13.2	Turn left on Stone Rd.
14.5	Turn left (north) at T-intersection onto Holman City Rd.
15.85	Entrance to Ludlow Sand & Gravel Company Quarry via dirt road on the right - STOP 1.

STOP 1 LUDLOW SAND AND GRAVEL COMPANY

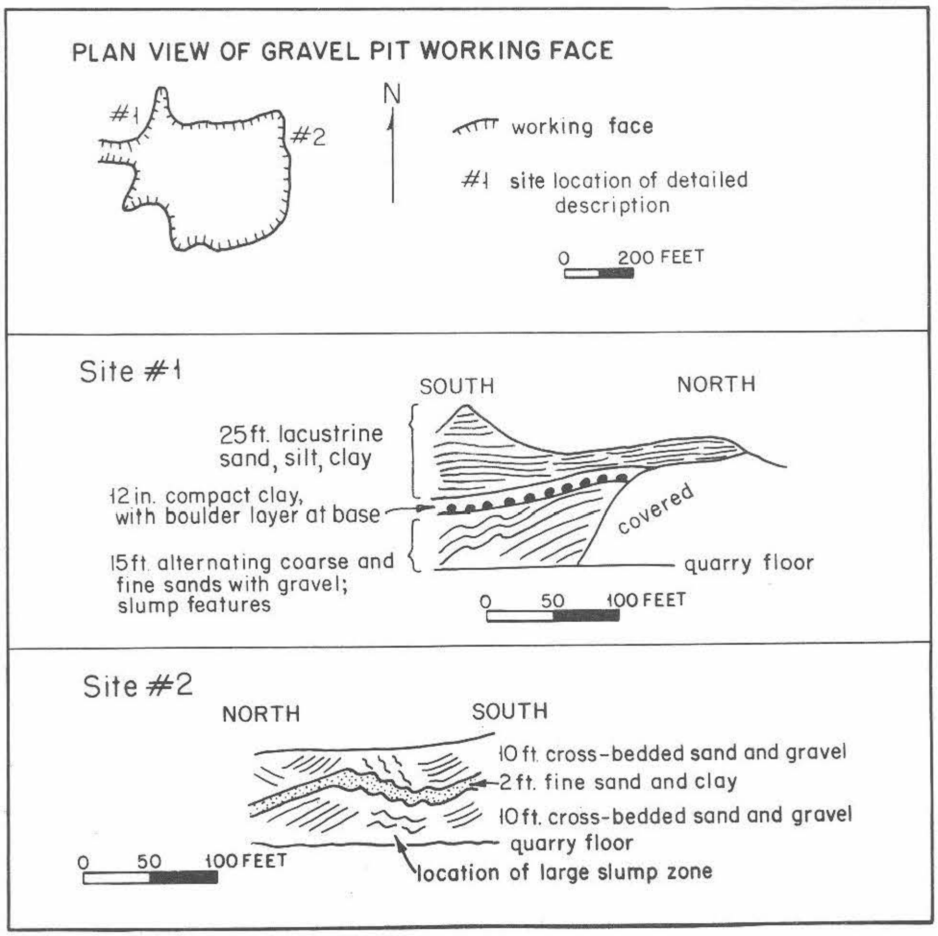
This gravel pit is located at the Valley Heads Moraine. Past exposures have illustrated the complex environments of deposition associated with this ice-marginal position.

Materials: well sorted, stratified sand, silt and clay; coarse cross bedded gravels; delta foreset gravels; laminated lake clay; poorly stratified fine sand; compact clay diamict (till).

Sedimentary structures: cross-bedding; channel lag gravels; cut and fill; large foreset beds; ripple drift laminations; slumped sands, gravels and silts.

Environment of deposition: Large(?) lake with meltwater streams of variable discharge, with deposition of gravels, sands and silts; near glacier terminus (possibly ice-contact) with deposition of sediments over blocks of glacier ice.

Noteworthy characteristics: ice-contact collapse features; thick, poorly stratified sands; well stratified, interbedded sands and gravels; channel cross bedding; delta foreset beds; compact clay layer with boulders concentrated at the base. This is diamict, deposited during an oscillation of the ice margin.



Return to entrance of gravel pit; turn left (south) on Holman City Road

16.05 Segment of Valley Heads Moraine (on left and right)

16.55 On left are ice contact, crossbedded outwash gravels

16.75 Pitted outwash on right

17.20 Stone Road, on right

18.10 STOP sign at Babcock Hill intersection
Turn left onto Babcock Hill Road, toward Cedarville

18.40 Intersection, continue on Babcock Hill Road

18.75 Enter Herkimer County, North Winfield

19.75 STOP sign, in town of North Winfield
Continue straight on Babcock Hill Road

20.85 Intersection, turn right onto Brace Road

21.55 Turn left onto Cross Road

22.15 STOP sign at intersection with Meeting House Road, turn left

22.25 PICTURE STOP: "Smith Esker" on right. This esker can be traced for almost a mile. Continue on Meeting House Road.

23.85 STOP sign. Turn right onto Babcock Hill Road, continue toward Cedarville

24.85 STOP sign. Turn right onto NY Rt. 51 (south)

24.90 Cross Cedarville meltwater channel

24.95 Turn right, stay on NY Rt. 51 (south)

27.75 Cross railroad tracks, small gravel pit on right in outwash gravels

28.20 Junction Route 20. Turn left (east)

29.20 Enter Otsego County

35.40 Intersection with Rt. 28; proceed east on Rt. 20

35.60 Enter Richfield Springs

36.30 Rt. 167 (north) enters from left; proceed east on Rt. 20

42.10 Traffic light intersection, turn right on Rt. 80 (south

Sign for Village of Springfield Center

42.35 Turn right on dirt road entrance into Hussey Quarry (unmarked, across from swampy pond) - STOP 2

STOP 2 HUSSEY QUARRY

This gravel pit contains a diamict (till) that documents the Cassville-Cooperstown readvance. The main things to see here are the stratigraphic units and the nature of their deformation.

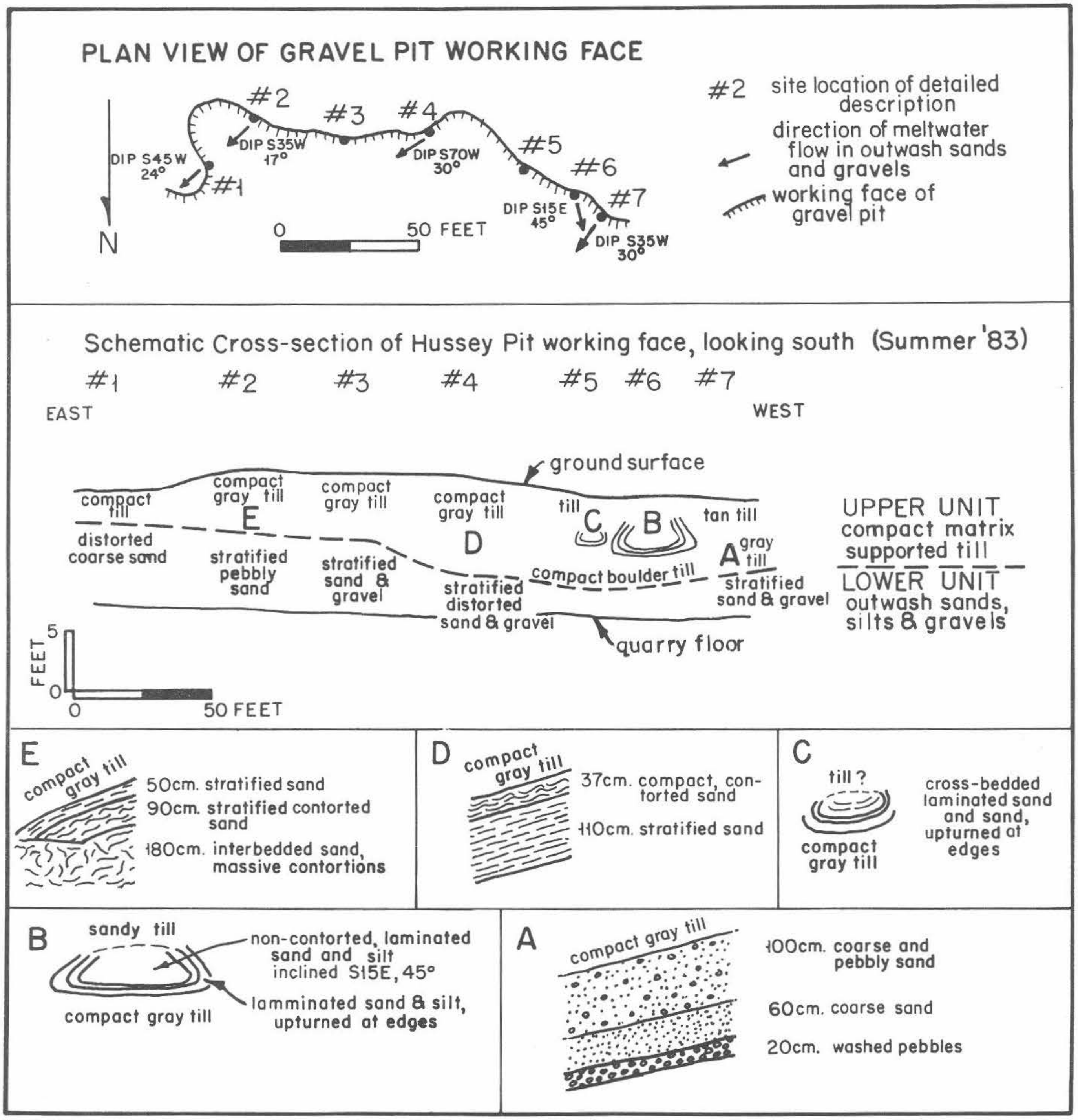
There are three fundamental stratigraphic units exposed in the south wall of the quarry (see STOP 2 diagram). They are 1) a compact, fine-grained, matrix-supported gray diamict, which overlies 2) a tightly compact stratified drift containing fine gravels, interbedded coarse sands, and well sorted, highly distorted fine sands, and 3) a lower gravel unit containing well washed medium to coarse grained pebbles, cobbles and boulders. The compact stratified drift occurs as distorted lenses of cross-bedded sand and silt within the diamict. The lenses are upturned at their edges and contain distorted sand and silt and show thickening of each unit along fold axes. This thickening suggests that the sand and silt were NOT frozen during deformation. The underlying gravel is not well enough exposed to determine if it too is distorted.

The following questions must be addressed.

1. Origin of the compact stratified drift -
    a. Were the sand and silt lenses sheared up from the lower outwash and included in the diamict by overriding ice movement?
    b. Were the lenses deposited at the same time as the diamict?
    c. Were the lenses deposited after the diamict and subsequently distorted by the diamict squeezing up and around them?
2. Causes of compaction and distortion -
    a. Were the lens edges folded during emplacement through shearing?
    b. Were the lenses deposited within the diamict with subsequent glacier overriding and deformation?
    c. Were the lenses distorted during "dewatering" of the diamict?

STOP 2 diagram - Stratigraphic units and nature of deformation in Hussey Quarry on the following page.

STOP 2 diagram. Stratigraphic units and nature of deformation in Hussey Quarry.



42.80 Continue south on Rt. 80 through Village of Springfield Center

44.90 Rt. 80 traverses hanging delta to Lake Cooperstown

51.90 Village of Cooperstown; follow Rt. 80 & Rt. 28 through Cooperstown

52.80 Proceed south on Rt. 28 past intersection with Rt. 80 west and Rt. 28 north

54.70 Turn right onto Otsego County Rt. 26, which climbs up the kame and kettle topography of the Cassville-Cooperstown Moraine (excellent view of moraine to the left)

54.70 PICTURE STOP - Proceed west on Rt. 26.

58.35 Intersection with Rt. 80 (west) and Rt. 28 (north)

60.55 Turn left on Rt. 80 (west) & Rt. 205 (south)

62.50 Rt. 205 turns south; continue west on Rt. 80

66.70 Turn left in hamlet of Burlington at blinking light on Rt. 51 & Co. Rt. 16

68.75 Esker on valley floor to the left, associated with ablation till

72.25 to 72.45 Esker complex on left

72.90 Turn left (east) in hamlet of Garretsville on Co. Rt. 16 and cross valley

73.20 Turn left on dirt road

73.25 Entrance to New Lisbon Landfill - along New Berlin ice margin - OPTIONAL STO

73.30 Back on Co. Rt. 16, heading south

73.90 Bear right on dirt road and right again back down onto valley floor

74.40 Intersection with Rt. 51, turn left (south)

74.55 Hill on left is the upvalley end of an esker

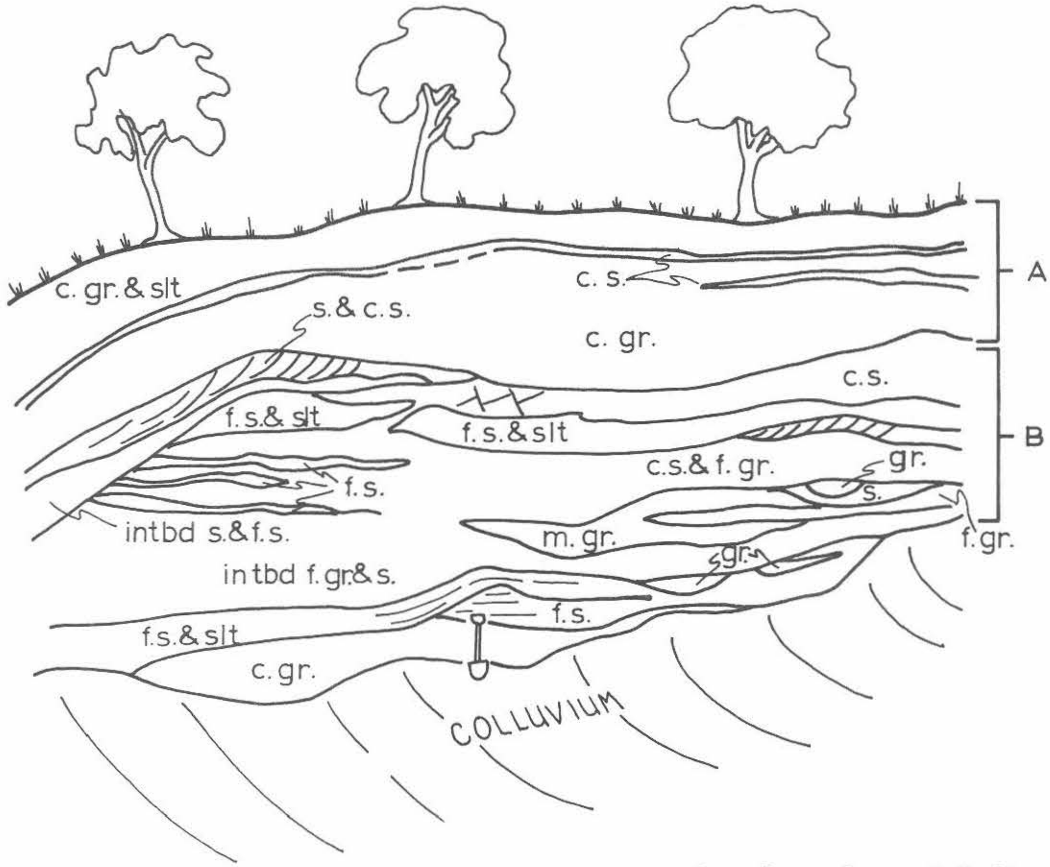
75.35 Additional eskers are seen downvalley

78.45 Entrance on left to quarry in deltaic outwash establishing a body of water in Butternut Creek valley at approximately 1150'

80.9 Road enters from right, Co. Rt. 49

81.10 Road on the left through split rail fence and across corn field; bear left at bottom of hill and through woods to quarry entrance, which cannot be seen from Rt. 51 - STOP 3

STOP 3 MORRIS KAME QUARRY diagram and description.



shovel scale = 3.5 ft.

STOP 3 Diagram. Ice-contact stratified drift in a kame .75 mile east of Morris, on the west side of Butternut Creek

In this quarry are exposed deposits typical of non-through valley deposition (stagnant ice blocks) and the associated conditions of glacier retreat. This kame and other landforms across the valley to the south depict an ice-contact environment dominated by local ponding and glaciofluvial deposition. The diagram above illustrates the northwest-facing working exposure in 1983.

stratigraphic units	sedimentary structures
A - Coarse gravel fining upward to silty matrix Interbedded coarse sand lenses and layers	A - Crude cross bedding, graded bedding and imbricate clasts. Average thickness is 8 feet.
B - Well sorted sand, interbedded fine gravel and sand, sorted sand and interbedded fine sand and silt	B - Cut and fill, channel lag gravel, collapse displacement, cross bedding, draped bedding, slump tilting. Average thickness above quarry floor (general elevation of 1140 feet) is 12 feet.

Miles

81.10 Return to Rt. 51 and turn right (north)

81.95 Turn left on Otsego Co. Rt. 49, which climbs up the Aldrich Creek valley. Aldrich Creek is an underfit stream occupying a valley that was enlarged by meltwater discharge along the New Berlin ice margin. Patches of discontinuous ablation till are found scattered along the valley.

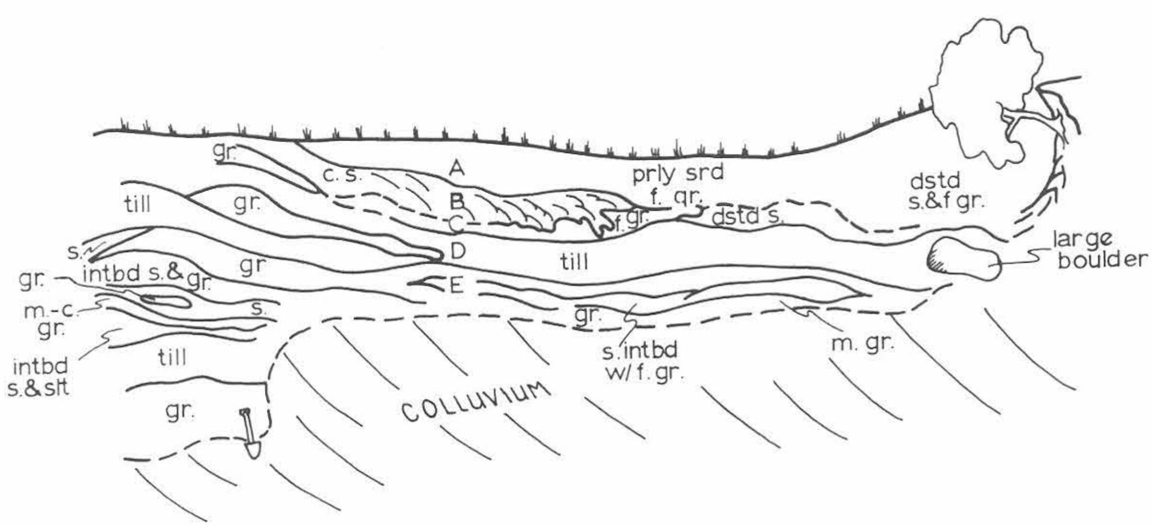
83.45 House and outbuilding on the right are situated on ablation till

87.55 Intersection with Ramey Road on left and Burski Road on right (the names of both are unmarked at intersection), proceed straight (north)

87.65 Entrance to gravel quarry on right - STOP 4

STOP 4 DIVIDE FACIES OF ICE-MARGINAL DEPOSITS

This semi-active small quarry lies along the trend of the New Berlin ice margin. Typically, ice-marginal positions are correlated across divides by recognizing erosional features such as cols, meltwater channels or underfit streams. Here, however, is a rare view of the materials constituting a depositional landform, which resembles an ice-contact kame. The diagram below illustrates the association of stratigraphic units at STOP 4.



shovel scale = 3.5 ft.

Stop 4 Diagram. Stratigraphy and ice-contact structure of divide facies deposits.

Stop 4 Description on following page.

STOP 4 Description.

Unit	Thickness	Description
A	36"-48"	poorly sorted fine gravel, distorted in places
B	24"-30"	well stratified coarse sand, distorted by slump, collapse or injection from below
C	6"-12"	fine gravel with sand and silt matrix
D	30"-40"	compact diamict with clay matrix and striated cobbles
E	36"-?	washed gravels interbedded with a few layers of well sorted sand (few inches thick) which seems compact

Miles

Proceed north on Otsego Co. Rt. 49

88.35 Turn left at intersection with Otsego Co. Rt. 17 and proceed west. Rt. 17 follows along the New Berlin ice margin as it descends from the divide through the small hamlet of Cardtown and toward Pittsfield.

89.75 Turn left on Rt. 80 (west) and proceed toward New Berlin

90.1 Kame moraine complex on the right is separated from valley wall by an outwash channel

90.9 Hamlet of Pittsfield. Highway climbs up the back side of a kame moraine-outwash complex that swings across the valley marking a brief hesitation along the retreating New Berlin ice margin. Several similar complexes are found northward along Wharton Creek toward Edmeston.

91.7 Rt. 80 descends the pitted outwash surface and follows the southern margin of a prominent dead-ice sink.

92.05 View of dead-ice sink on the right.

92.50 Bear left on Musk Road up a small hill (unmarked dirt road at bridge over Wharton Creek). Don't follow Rt. 80 across bridge!

92.95 Another dead-ice sink can be seen on the valley floor at 2 o'clock

93.40 Intersection with Otsego Co. Rt. 18 (unmarked) at YIELD sign. Turn left and follow Rt. 18 for .2 mile.

93.60 Intersection with Otsego Co. Rt. 13, turn right (west)

93.95 New Berlin valley train on left - OPTIONAL STOP

Proceed west on Rt. 13

94.20 Cross Unadilla River, enter Chenango Co. and Village of New Berlin

94.40 Intersection with Rt. 8 at YIELD sign. Turn left and proceed south on Rt. 8

94.85 Entrance to New Berlin Gravel Quarry on the right - STOP 5

STOP 5 NEW BERLIN GRAVEL QUARRY

The gravel quarry has been developed at two levels within a deltaic valley train. The upper level is primarily within the topset beds and the lower level (adjacent to Rt. 8) exposes massive foreset beds. Limestone and associated chert are major lithologic components constituting approximately 26% of all clasts (14% limestone, 12% chert). Postglacial leaching and reprecipitation of carbonates accounts for the various stages of induration found within some of the gravel units.

The topsets consist of 20 to 25 feet of moderately to poorly sorted, horizontal cobble gravels that contain some coarse sand lenses. Coarse lag gravels are common and depict a cut and fill origin. Crude channel cross bedding also is evident. Some of the sand lenses contain small-scale collapse features which are too small to have disturbed the associated gravel.

The underlying foreset beds are inclined at 28° to the southeast (S 30°-50°E) and generally consist of a finer pebbly gravel interbedded with sand. The gravel units consist of both well sorted, clast supported, matrix-free layers and sandy, pebbly fine gravel. The sand ranges from fine to coarse, with occasional silt layers up to 1 foot in thickness. No major collapse features or contorted beds were exposed in 1983.

This deposit is interpreted to be part of a massive deltaic valley train that was prograded downvalley from the New Berlin ice margin (just .5 mile to the north) into an ice-contact lake. The dam was located 15 miles downvalley in the vicinity of Mount Upton and Rockdale. The elevation of the dam corresponds to the topset-foreset contact elevation here. A major dead-ice sink marks the ice-marginal position at the head (upvalley extent) of the valley train. These depositional landforms and their association are helpful in recognizing the valley facies of ice-marginal positions.

Turn around and proceed north on Rt. 8

95.60 Traffic light intersection of Routes 8 and 80. Continue straight on Rt. 8 (north) and Rt. 80 (west)

96.85	Proceed north (straight ahead) on Rt. 8 at intersection where Rt. 80 (west) turns to the left
98.10	Highway follows base of kame terrace scarp for the next 2.5 miles. Kame terrace stands at 1300 ft. The kame terrace represents outwash shed along an ice margin that paralleled the Unadilla River valley, from New Berlin to Columbus Quarters. Associated kame moraines also occur along the valley, as at South Edmeston on the east side of the valley and Columbus Quarters on the west side. Late glacial discharge from the north (possibly associated with the Valley Heads readvance) dissected all of these ice-marginal landforms.
101.30	Highway climbs onto outwash associated with kame moraine
101.75	Columbus Quarters intersection with Chenango Co. Rt. 41. Kame-moraine complex to the right. Continue north on Rt. 8
102.00	Highway descends north side of kame moraine, which is part of the New Berlin ice margin
104.40	Enter Madison County; proceed north on Rt. 8 to Bridgewater
106.70	Breached and eroded remnant of Cassville-Cooperstown Moraine can be seen on the valley floor
113.00	Outwash plain forming the valley floor on the right developed from Valley Heads discharge originating from the north
113.90	Enter Oneida County
114.50	Junction with Rt. 20 at Bridgewater; proceed north on Rt. 8
116.50	Outwash surface on the right grades northward to Valley Heads ice-marginal position near Cassville, 3 miles to the north
119.00	Leave Rt. 8 (which bears to the right) and proceed straight ahead toward Cassville
119.30	Turn left in Cassville on Summit Road
119.90	Road crosses characteristic hummocky topography of Cassville-Cooperstown Moraine for about one mile
121.30	Proceed straight toward Paris on Doolittle Road
124.40	Sign for Paris
124.55	STOP sign intersection with Rt. 12, bear right (north) on Rt. 12
125.30	Turn left toward Clinton on Fountain Street and enter Town of Marshall. Road crosses trend of Valley Heads Moraine, which is not well expressed in the topography here.
129.15	Sign for Village of Clinton
129.50	STOP sign intersection with Rt. 12B, turn left at Clinton Village Square to YIELD sign and proceed straight

129.55 Traffic light intersection

129.70 Traffic light intersection of Rt. 12B (south) and College Street (Rt. 412) where road log began. Continue straight for entrance to Hamilton College.

N O T E S

Trip B-8

Two Till Sequence at Dugway Road Exposure Southwest of Clinton, New York

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INTRODUCTION

During the waning stages of the Wisconsinan ice sheet, the southwestward ice movement across the Adirondacks changed to a strong flow around both sides of the Adirondack Mountains. The Ontario lobe moved southwestward up the St. Lawrence Valley and spread southward across the Ontario Lowlands, whereas the ice east of the Adirondacks moved southward along the Champlain-Lake George trough and expanded southward down the Hudson Valley and westward up the Mohawk Valley. Chamberlin (1883), Fairchild (1912) and Brigham (1931) suggested that these two lobes were contemporaneous in the area south and west of Utica, N.Y. To fit his meltwater chronology, Fairchild (1932) proposed that the Mohawk lobe wasted away prior to the Ontario lobe, allowing free drainage down the Mohawk Valley while the Ontario ice receded from the Syracuse-Utica area. Krall (1977) supports the latter view based on drumlin realignment, morainal orientations, and the superposition of tills at the Dugway Road exposure, 6 km southwest of Clinton, N.Y.

THE DUGWAY ROAD SECTION

The Dugway Road exposure consists of a dark gray till, with some interspersed yellow-brown layers, overlain by weak red till (Figure 1). Between the two tills is a single boulder layer underlain by a yellow-brown material somewhat sandier than either of the two tills. The top of the exposure consists of finely laminated (varved?) clays.

Ice movements can be inferred by till fabric and till color. A fabric taken in the red till shows a N 31°W trend, whereas a fabric taken 2 m below the boulder layer shows a trend of N 22°E. Two other till fabrics were taken, one just below the contact of the two tills (N 2°W) and the other 1.3 m below the contact (N 52° E). The dark gray color of the lower till can be attributed to southwestward movement of the Mohawk lobe across the Utica Shale, Frankfort Shale, Clinton Formation and the Lockport Formation, all of which have black or dark gray to blue-gray components (Dale, 1953). See inset map on Figure 1. The upper till was deposited by the Ontario lobe moving southeastward, a direction that would have taken the ice over red Vernon Shale for nearly 7 km before reaching the Dugway Road position. This evidence implies Mohawk retreat prior to advance of the Ontario lobe.

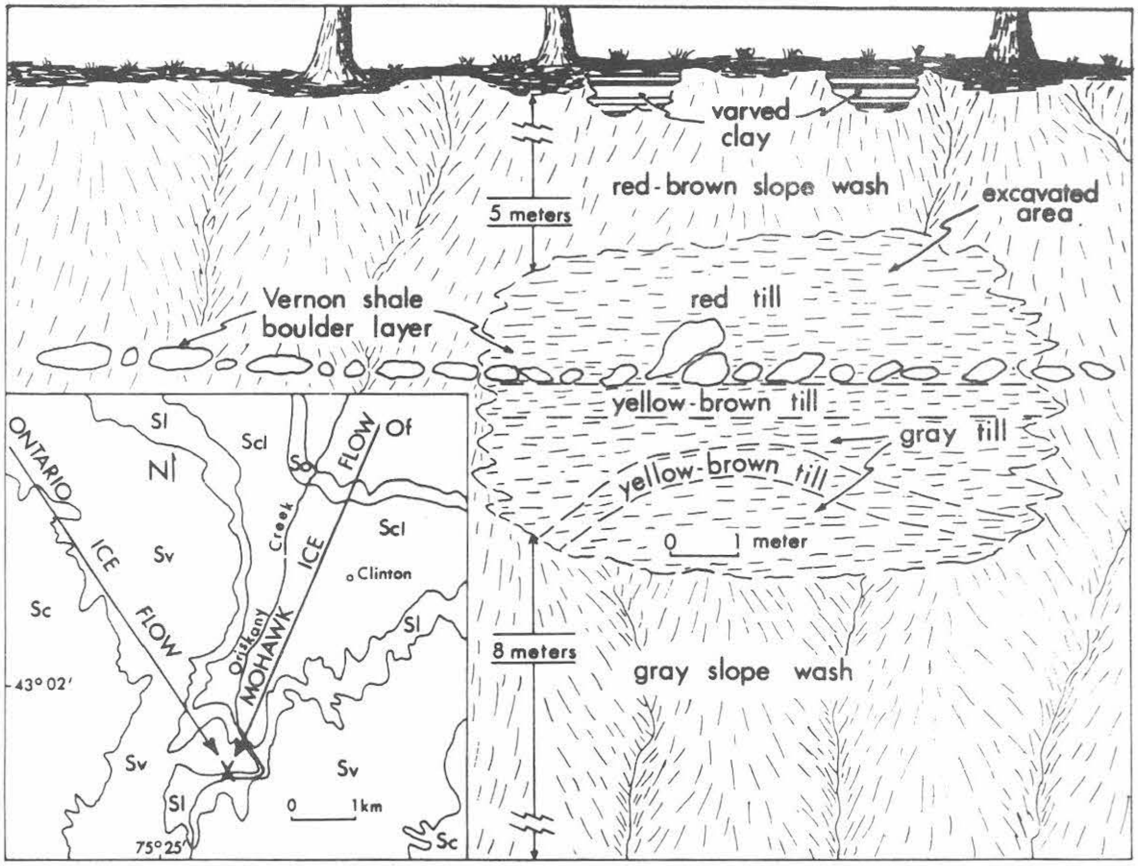

Figure 1. Sketch of the Dugway Road exposure. Inset map shows location. Geology is after Dale (1953). Of = Frankfort Shale; So = Oneida Conglomerate; Scl = Clinton Formation; Sl = Lockport Formation; Sv = Vernon Shale; Sc = Camillus Shale. Assumed ice-flow directions based on till-fabric orientations at exposure. Adapted from Krall (1977).

QUESTIONS

1. Does the boulder layer and the yellow-brown sandy material represent an ablation till deposited by the Mohawk lobe, a weathered zone of the lower till, or a deposit by the later Ontario lobe?

2. Is there a "significant" time interval between the deposition of the two tills?

3. Does the Mohawk till correlate with the West Canada Till or the Hawthorn Till of Muller, et.al. (1983) and Ridge and Franzi (field trip C-10 of this publication)?

4. What is the cause of the lower yellow-brown layer in the Mohawk till?

5. The Dugway Road exposure owes its existance to stream divergence of Oriskany Creek, which leaves the broad Oriskany Valley to the west, flows due east for more than a kilometer and then cuts northward through a narrow bedrock gorge (Figure 2). What caused this divergence?

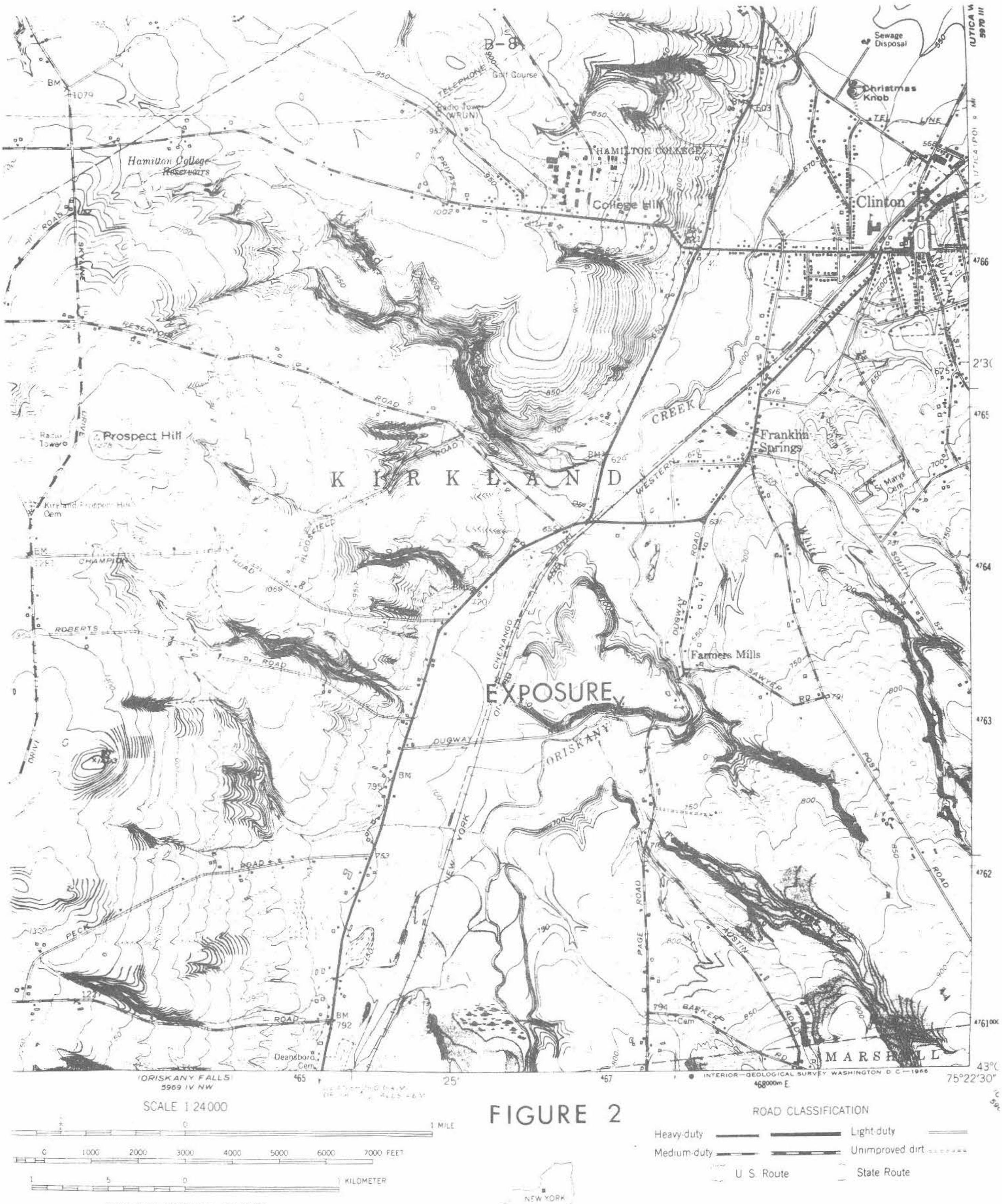
Hamilton College Reservoirs
HAMILTON COLLEGE
College Hill
Clinton
Christmas Knob
Sewage Disposal
Golf Course
Radio Tower (WRUN)
Prospect Hill
Kirkland Prospect Hill Cem
K I R K L A N D
Franklin Springs
St Marys Cem
Farmers Mills
EXPOSURE
CREEK
WESTERN
CHENANGO
ORISKANY
NEW YORK
DUGWAY
ROBERTS
CHAMPION
PECK
ROAD
SAWYER RD
PAGE ROAD
POST ROAD
BARKER
Deansboro Cem
MARSHALL
TELEPHONE
PRIVATE
SKYLINE DRIVE
RESERVOIR
ORISKANY FALLS
5969 IV NW
INTERIOR—GEOLOGICAL SURVEY WASHINGTON D C—1966
75°22'30"
SCALE 1:24 000
1 MILE
7000 FEET
1 KILOMETER
CONTOUR INTERVAL 10 FEET
DATUM IS MEAN SEA LEVEL
ROAD CLASSIFICATION
Heavy-duty
Light-duty
Medium-duty
Unimproved dirt
U. S. Route
State Route
NEW YORK
QUADRANGLE LOCATION
CLINTON, N. Y.
SW/4 ROME 15' QUADRANGLE
N4300—7522.5/7.5
THIS MAP COMPLIES WITH NATIONAL MAP ACCURACY STANDARDS

FIGURE 2

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SEDIMENTOLOGY AND FAUNAL ASSEMBLAGES IN THE HAMILTON GROUP OF CENTRAL NEW YORK

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INTRODUCTION

The Middle Devonian (Givetian) Hamilton Group in Central New York State consists of approximately 1600 ft. (approximately 490 meters) of richly fossiliferous interbedded shales, siltstones and sandstones. During Givetian time, the central New York region lay to the west (present geography) of the Appalachian mobile belt. The sediments derived from this orogen were transported to the west and northwest into an epicratonic foreland basin. Other studies (Brett et al., 1983; Baird, 1979) suggest that deposition in western New York took place on a topographically subdued bottom. Regionally, paleoslope was SW-directed, but local swells and depressions of 20-100 km wave length were present (Baird, 1979). In the region of this study, the Hamilton has been the subject of numerous paleontological studies, from the pioneering work of James Hall through the biostratigraphic studies of Cooper (1930, 1933). More recently, the integration of paleontological data and detailed sedimentologic/stratigraphic studies have resulted in the development of depositional/paleoecological models (e.g. Brett et al., 1983). The purpose of this trip is to introduce participants to the varied faunas and lithofacies of the Hamilton and to document the cyclic arrangement of faunas and facies. It is this cyclicity that provides us with a key to understanding the sedimentological evolution of the Hamilton Group.

Fig. 1. Location map of study area. Stippled pattern illustrates distribution of marine (Hamilton Group) Givetian rocks in New York.

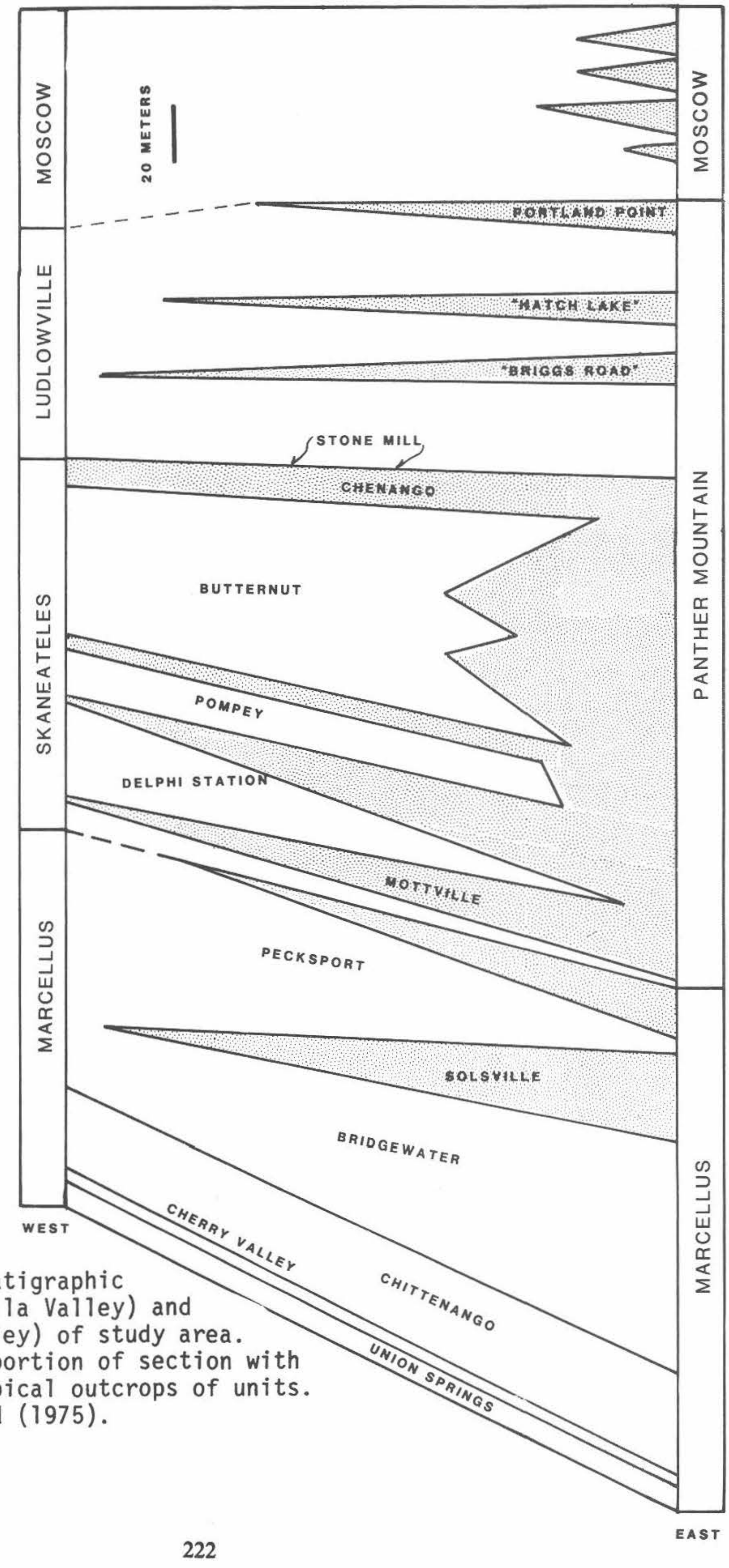

Fig. 2. Schematic lithostratigraphic section east (approx. Unadilla Valley) and west (approx. Cazenovia Valley) of study area. Stippled pattern indicates portion of section with greater than 50% sand in typical outcrops of units. Thicknesses based on Rickard (1975).

STRATIGRAPHY

The stratigraphy developed by Cooper (1930, 1933), and modified by Rickard (1975) is presented in Figure 2. The subdivision of the Hamilton Group in the region of this study was developed by Cooper in west-central New York, where a number of thin limestones provide lithologically prominent units that serve as formation-level boundaries. Correlation of these limestones with laterally equivalent thin limestones, shell beds and sandstones permitted Cooper to carry a four-fold subdivision of the Hamilton Group to the east, through the area cf the Chenango Valley. Loss of key horizons and rapid facies changes prevented Cooper from successfully carrying this four-fold subdivision beyond the Unadilla Valley (approx. 50 kms to the east of the area of this study).

The basal Chittenango Member of the Marcellus Formation rests atop the underlying Seneca Member of the Onondaga Limestone. Rickard (1975) has documented the regionally diachronous character (youngest in western N.Y.S.) of this contact. Locally, the contact is quite sharp, with interbedded calcareous shales and thin, dark limestones occurring immediately above the contact. The Cherry Valley Limestone Member of the Marcellus, possessing an abundant benthic fauna, separates the poorly fossiliferous Chittenango Member from the dark, poorly fossiliferous Union Springs silty shale.

The remainder of the Hamilton Group, including that portion of the Marcellus above the Union Springs shale, the Skaneateles, the Ludlowville and Moscow Formations, is characterized by rather regular, cyclic arrangement of lithologies, sedimentary structures and accompanying fauna. This existence of this cyclicity is immediately evident on simple inspection of the stratigraphic column in Figure 2. Beginning with the silty shales of the Bridgewater Member, a regular repetition of silty shales and sandstones recurs throughout the Hamilton Group. These include the Bridgewater-Solsville; Pecksport-Mottville; Delphi Station; Pompey; Butternut-Chenango plus Stone Mill; two (unnamed) shale-sandstone cycles in the Ludlowville; upper Ludlowville-Portland Point; plus at least 4 cycles in the Moscow Formation. These alternating silty shale, siltstone, and sandstone units form the basis of the member-level subdivisions of the Hamilton by Cooper. Within these major silty shale-sandstone cycles, smaller-scale "subcycles" are recognizable.

The internal stratigraphy of many of the Hamilton Group cycles in the study area consists of basal dark, grey fissile silty shales with low faunal diversity, grading upward into coarser-grained shaley siltstones with abundant marine fauna, followed by fine-grained silty sandstones possessing a somewhat lower faunal diversity. The sandstone units exhibit a variety of current and/or wave formed primary structures. Internally, the sandstone units often coarsen upward and contain pronounced internal scour surfaces marked by thin shell concentrations. In some instances, the coarsest sandstone of a cycle is capped by thin limestones and/or shell-rich units (e.g. Chenango-Stone Mill; Upper Ludlowville-Portland Point). These sandstone units, with associated bioclastic grainstones, are regionally correlative with limestone units in west-central New York State, and record conditions that were boradly contemporaneous

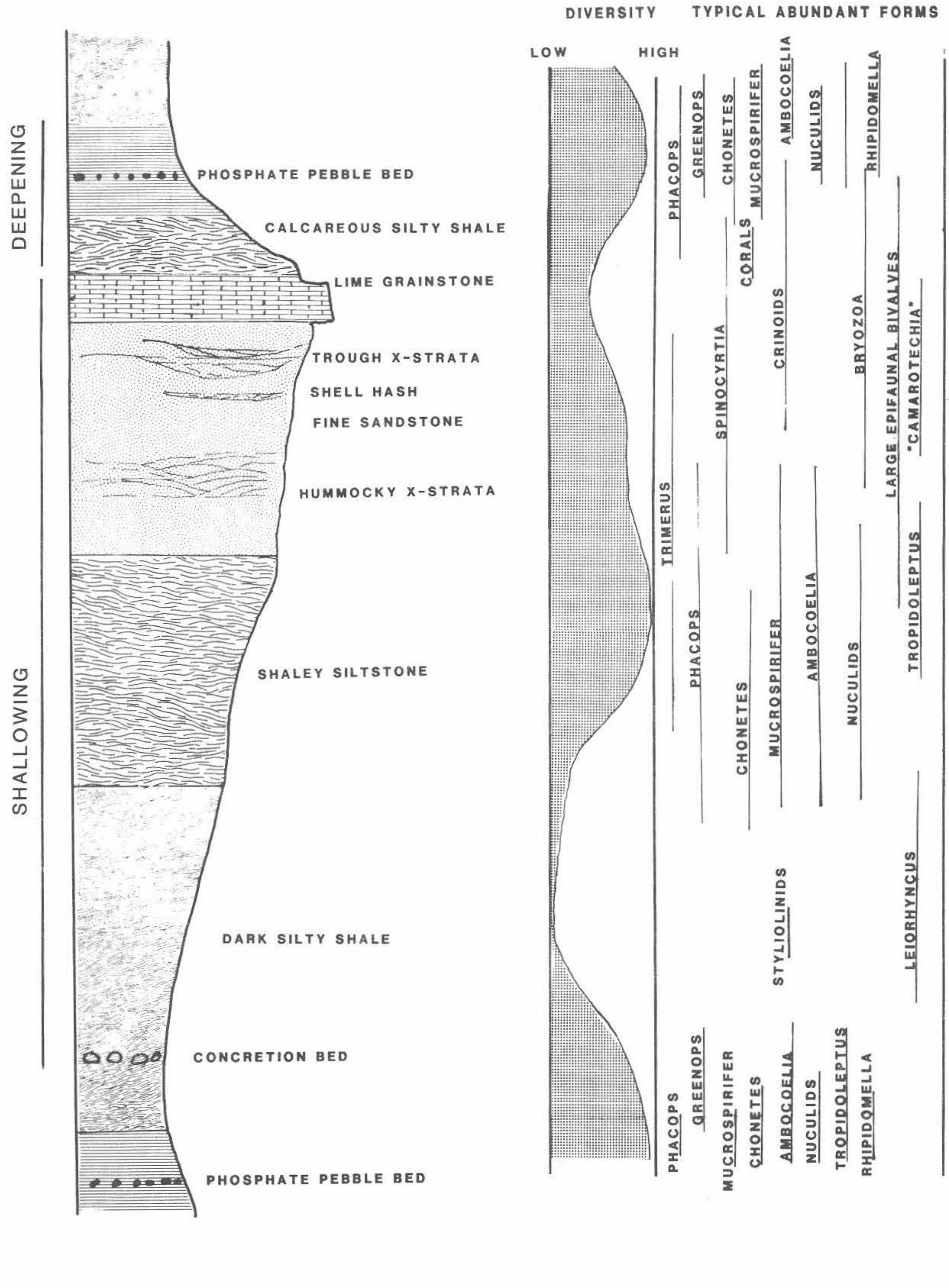

Fig. 3. Idealized Hamilton Group Cycle illustrating arrangement of various lithofacies and faunal elements. Note that many of the features shown are not present in all cycles. Vertical scale is variable.

on a basinal scale. Above the sandstones (or overlying bioclastic grainstone), finer-grained facies appear. Calcareous shaly siltstones with a relatively diverse fauna give way to finer silty shales, followed in some portions of the section by dark silty shales with a more limited fauna. Phosphate pebble beds commonly occur in the silty shales overlying sandstone units. The phosphate pebble beds exhibit considerable lateral continuity.

On a regional scale, some sandstone units can be traced west to the Skaneateles Valley (approx. 100 kms west of the study area). To the east, sandstone units merge into sandstone-dominated facies east of the Cooperstown Valley (approx. 75 kms east of the Chenango Valley) as intervening finer-grained silty shales become subordinate. The stratigraphic system that emerges, is one of sandstone "units" extending to the west within otherwise mudrock-dominated facies. Further, the sandstone units of the Central New York area are in part stratigraphically equivalent to the thin limestones that form the stratigraphic framework of the western N.Y. Hamilton Group. The cycles identified within the Hamilton Group of Central New York are asymmetric, in that if the coarsest sandstone units or bioclastic grainstones are considered the "core" of each cycle, the overlying "fining upward" portion is considerably thinner than the succeeding "coarsening upward" section. Further, it is clear that the cycles reflect alternating "shallower" and "deeper" water depositional conditions.

Internal Features of the Hamilton Group Asymmetric Cycles

For the purpose of discussion, the Hamilton Group cycles can be subdivided into 5 major facies types (Figure 3): basal dark silty shales; shaley siltstones; silty sandstones; bioclastic grainstones (often absent) and calcareous silty shales. While a particular cycle may not exhibit each facies component and many subtle vertical alterations of facies exist, outcrop-scale variations usually permit resolution of position within a larger cycle.

Basal dark silty shales:

Common attributes of this facies include low faunal diversity, general lack of current-formed primary structures and relatively high organic contact. General bioturbation is often present, but difficult to observe due to compaction. Pyritized burrows and fossil fragments are common. The typical fauna consists of *Leiorhyncus* sp. (common), styliolinids (especially in Marcellus Formation), *Ambocoelia* (near contact with overlying silty shale facies), plus rare *Chonetes*, *Paleozygopleura*, *Greenops*, *Tornoceras*, *Spyroceras* nuculids and plant fragments. Quantitative measures of both diversity and equitability indicate a stressed environment. The high organic content of this facies suggests low oxygen levels on the substrate, and oxygen availability was likely a diversity-limiting factor. Thin ripple-laminated siltstones are often present in this facies, and may document local- to regional-scale storm-related turbidite events. Concretions, often containing authigenic metallic sulfides, barite, celestite and dolomite in Septarian fractures, document pre-compaction cementation in this facies.

Shaley siltstones:

This facies consists of grey to brown crumbly to blocky weathering, slightly calcareous shaley siltstones, which exhibit upsection increase in faunal diversity within a given cycle. The basal portion of this facies contains a low to moderate diversity fauna dominated by Ambocoelia, Chonetes, nuculid bivalves, and Mucrospirifer. Greenops is typically present, but is replaced in relative abundance by Phacops in coarser variants of this facies. As the overlying sandstone facies is approached, faunal diversity rapidly increases, particularly through the addition of the larger brachiopods (Spinocyrtia, Tropidoleptus) semi-infaunal bivalves, bryozoans, and crinoids. The highest diversity and equitability is found near the siltstone-sandstone transition. Thin shell-rich horizons are common in this facies, but are often masked by subsequent bioturbation. Evidence of traction current reworking is limited to shell beds and rare ripple-laminated beds.

Silty sandstones:

The fine-grained silty sandstone units exhibit abundant current-formed-primary structures including ripple cross-lamination, hummocky-cross stratification, trough cross-stratification, symmetrical and asymmetrical ripples and internal scour surfaces floored by shell debris. The fauna consists of the larger brachiopods and bivalves, plus bryozoans and crinoids. Horizontal and vertical burrows and feeding traces are common, but complete bioturbation is rare.

Soft-sediment deformation (pillow structures, "flow rolls") are present in some sandstone units, and indicate the re-adjustment of sands rapidly deposited over watery muds. Rip-up clasts of mud-rich lithologies are common.

Faunal diversity is considerably lower in the sandstones than in underlying shaley siltstones, but equitability remains high.

Bioclastic grainstones:

This lithology is comparatively rare in the section, but is critical to interpretation of Hamilton Group cycles. Typically the grainstones occur as laterally discontinuous lenses consisting of abraded shell debris. Crinoid plates and brachiopod and bivalve fragments set in a matrix of medium-grained sand is the dominant lithology. Clast of cemented limestone and rare phosphate pebbles and shale clasts are also present. In one instance (Stone Mill Member) the top portion of the limestone contains an "in situ" fauna with abundant rugose and favositid corals and bryozoans. Because of the fragmental nature of the fossil material and the clearly transported nature of the assemblage, faunal diversity and equitability have not been determined.

Calcareous silty shales:

Rather abruptly overlying the sandstone or bioclastic grainstones are grey, calcareous silty shales. This facies exhibits increased faunal diversity and equitability compared to the underlying sandstones, and is dominated by large brachiopods, bivalves and bryozoans. Thin (2-15 cms) phosphate pebble beds are common, usually occurring as a single, laterally continuous bed. The phosphate pebble beds consist of a mixture of broken and whole shells, 0.5 - 1.0 cm subrounded phosphate pebbles, phosphatic steinkerns of trilobites and brachiopods and phosphatic fish plates that form a distinctive resistant pavement in quarry exposures. These beds clearly record periods of very limited clastic sediment input and lengthy exposure of the sediment-water interface to early diagenetic and chemical processes. The abraded shell material and rounded phosphatic clasts indicate at least sporadic traction current activity. The dominant faunal elements of the calcareous silty shales are similar to those found in the shaley siltstone facies of Hamilton Group cycles, reflective of similarity of environmental conditions.

Detailed discussion of particularly significant faunal elements and sedimentological/early diagenetic features are included in stop descriptions.

ENVIRONMENTS OF DEPOSITION

The asymmetric cycles in the Hamilton Group record deposition of terrigenous clastic sands and muds under conditions of varying water depth, wave and current reworking of the bottom, oxygen levels and substrate stability. Regional facies patterns and local vertical facies changes suggest that the Hamilton Group of the study area is not a product of "classic" deltaic deposition. Paleocurrent directions and regional facies patterns indicate that the clastics were derived from a tectonic source land to the present southeast, yet the cyclic sedimentation pattern does not appear to result from simple progradation of deltaic facies onto a marine platform. External eustatic sea-level fluctuations appear to better explain both vertical facies patterns and the temporal correlation of sandstones in central New York with thin limestones to the west. We suggest that each shale-siltstone-sandstone-shale cycle is the product of relatively rapid sea level rise, followed by deposition under relatively stable sea-level conditions. The dark silty shales which the base of each cycle were deposited during periods of maximum water depth, minimum reworking of the bottom by waves and currents and input of mud by suspended load deposition. With continued deposition, perhaps accompanied by slight sea level fall, water depths shallowed to first permit better agitation of bottom waters and colonization of the substrate by progressively more diverse marine faunas. With continued shallowing, influx of traction load sands and the production of hummocky cross-stratification, scour and fill, ripple lamination occurred as waves and currents influenced the bottom. Maximum shallowing is represented by the coarsest sandstones and/or bioclastic grainstones present in the cycles. Relatively rapid sea level rise following deposition of these facies resulted in diminished input of coarse terrigenous clastics and renewed deposition of muds inhabited by a diverse marine fauna. Phosphate pebble beds record virtual cessation of clastic input as rapid sea-level rise occurred. The eventual recurrence of dark silty shales marks the maximum deepening prior to the next upward shallowing depositional phase.

This model then suggests that the thin bioclastic limestones (e.g. Stone Mill Member, Portland Point) are the shallowest-water facies present in the section. This proposal stands in sharp contrast to the conclusions of McCave (1973), who suggested that the limestones are the products of deposition following maximum shallowing, when clastic input had been terminated by drowning of deltaic distributaries to the east. Our model correlates limestone deposition with maximum shallowing, when shell material was concentrated by wave and current winnowing of finer terrigenous clastic muds and sands. Rapid deepening, and drowning of sediment input systems correlates with the very low clastic depositional rates indicated by phosphate pebble beds and shell-rich horizons. Periods of maximum water depth are recorded by dark, silty shales with lower faunal diversity.

The suite of primary structures and the abundance of shell-lag horizons in sandstone units strongly suggest that the dominant bottom-current effects were storm related. Although tide-dominated facies have been recognized in the Hamilton Group to the east of the study area, evidence for tidal current effects in this area are absent. Storm-generated waves and bottom currents were responsible for the seaward transport of muds and sands from coastal areas to the east, the generation shell lag concentrations, hummocky cross-stratification (Bourgeois, 1980) and may have been responsible for the turbidite-like laminated siltstones found sporadically in the otherwise finer-grained, deeper water facies.

RATES OF DEPOSITION

At least 13 major asymmetric cycles can be identified in the Hamilton Group in central New York. Assuming that the Givetian deposition spanned approximately 7 million years, the duration of each major cycle was approximately 7 million years, the duration of each major cycle was approximately 500,000 years. This period is consistent with the estimated duration of sea-level oscillations identified elsewhere in the Mid-Paleozoic of the Appalachian Basin (e.g. Dennison and Head, 1975). Smaller-scale sea-level oscillations are superimposed on these longer periods of fluctuations.

The average depositional rate of the Hamilton Group is approximately 70 meters/10^6 years or less than 0.1 mm/year. This rate is considerably lower than that calculated for the subsidence rate (= long term probable net deposition) of modern sedimentary basins, which varies from 0.3-2.5 mm/year. This data indicates that Hamilton Group deposition took place upon a relatively stable cratonic basement, and thus, the subsidence behavior of the basin more nearly resembled the "tectonic quiesence" of the Late Silurian and early Devonian, characterized by deposition of platform carbonates, than the crustal instability and rapid basin subsidence evident in the late Devonian and Carboniferous history of the Appalachian Basin.

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ROAD LOG - Hamilton Group
Robert Linsley and Bruce Selleck

Note: The road log begins at the intersection of N.Y.S. Route 46 and Pratt Road, approximately 1.5 miles south of the Village of Munnsville. From Clinton, proceed south on N.Y.S. Rt. 12B, through Deansboro and Oriskany Falls to U.S. Rt. 20. Then west on Rt. 20 to Pine Woods, then north on N.Y.S. Rt. 46 to begin road log. Mileage to nearest 0.05 odometer miles.

Cum. Miles	Comments
0.0	Intersection of N.Y.S. Rt. 46 and Pratt Road. Proceed south on Pratt Road. Low hills in center of valley are part of the Munnsville kame-moraine complex. On the eastern valley wall, an active limestone quarry in the Helderberg Group and Lower Onondaga Limestone is visible.
0.55	Turn right (west) onto Stockbridge Falls Road.
1.50	Outcrops of Upper Helderberg Group and basal Onondaga Limestone (Edgecliff Member) on right.
1.80	Outcrop of Seneca Member on right.
2.25	Outcrops of Cherry Valley Limestone on right.
2.75	**Stop #1** - Upper portion of Chittenango Shale-Member of Marcellus Formation - Roadcut on left.

Stratigraphy/Sedimentology

This roadcut exposes a typical section of the black silty shales that dominate the lower Marcellus Formation. Fissile, organic-rich silty shales record deposition on an anaerobic/dysaerobic marine substrate. The extreme limitation of oxygenation and bottom circulation during deposition of this facies suggests that the waters of the basin may have been stratified in a manner analogous to the modern Black Sea. Large carbonate-cemented concretions are well-exposed mid-way up the outcrop face, and cearly demonstrate that pre-compaction cementation occurred. Note that the laminae in the silty shales are deformed by compaction around the concretions. Septarian fractures in the concretions have yielded finely crystalline dolomite, calcite and barite. Note that the surfaces of some concretions appear to have preserved burrows and steinkerns of gastropods. Is there a linkage between the presence of the burrows/fossils and the formation of the concretions?

Fauna

The fauna at this stop is exceedingly limited, dominated by Styliolina with occasional specimens of Nowakia, Leiorhychus and "Euryzone". The styliolinids and Nowakia occur abundantly on particular bedding planes that occur sporadically throughout the section. It is presumed that these are planktonic forms for they are occasionally found in almost all of the lithologies of the Hamilton Group, including the apparently inhospitable environment suggested by these non-bioturbated black shales. It has recently been persuasively argued (Yochelson, in press) that Styliolina is not a mollusk, and although he did not assign it to any other higher taxon, we are persuaded that it is safest to assume that it is not a member of an extant phylum, and Fisher's assignment to the Cricocondarida seems as sound as any. Nowakia occurs much more rarely (about 1% of the fauna), but is found with the styliolines and presumably shares a similar mode of life even though taxonomically it is a tentaculitid rather than a styliolinid.

Leiorhychus and "Euryzone" are not usually found in association on the same bedding plane as Styliolina, but scattered throughout the section. Two very different interpretations have been placed on the paleoecology of these genera. It has been suggested that "Euryzone" could have had a mode of life similar to the extant genus Ianthina, which builds a bubble raft to which it clings until it bumps into its prey, the Portuguese Man-of-War, Porcellia. It has similarly been suggested that perhaps Leiorhychus was attached by its pedicle to floating pieces of wood. These interpretations thus attribute the occurrence of these genera to being introduced from outside the immediate environment. However, since both species are restricted to the black shale environment of the Hamilton we suspect that they are indeed adapted to maintaining a marginal existence in an environment that is an abomination to other faunal elements.

(Note: we have placed quotation marks around "Euryzone" because we have never seen a specimen that was determinable below the ordinal level of Archaeogastropoda. Other members of the genus Euryzone are associated with reef deposits and the anomalous environment of the Hamilton form suggests that it is probably not appropriately assigned to this genus.)

Cum. Miles	Comments
2.75	Continue west on Stockbridge Falls Road
3.45	Turn left (west) onto unnamed road
4.90	Turn left (south) onto Glass Factory Road
5.45	Cross Oneida Creek
6.10	Turn right (west) onto Fearon Road
6.45	Turn left (south) onto Swamp Road
6.90	**Stop #2** - Bridgewater and Solsville Members of Marcellus Formation.

The basal section of the exposure here consists of dark, fissile to blocky silty shales which bear a rather limited fauna. The section coarsens upward, corresponding with gradual increase in faunal diversity. The basal portion of the Solsville Member is difficult to access at this stop, but a fair selection of typical Solsville forms can be found in talus mid-way up the exposure.

The preservation of the fossils at this locality is quite exceptional. Thin sections of the material shows that growth lines are well preserved, suggesting that the calcitic shells preserved here are probably original material. Some aragonite is still preserved, but most has been altered to calcite. Still even the mollusks exhibit good growth lines which suggests that the replacement of aragonite is a very precise molecule for molecule substitution.

This unit is also the site of some rather rare and unusual fossils including the monoplacophoran *Cyrtolites* and the bellerophont *Praematarotropis* and soft bodied preservation of annelids (Cameron, 1967).

The fauna is dominated by the brachiopods *Spinocyrtia* in the upper sandier facies, and *Mucrospirifer* in the middle siltier layers, along with the bivalves, *Ptychopteria flabellum*, *Gosseletia triquetra*, and a variety of nuculids, gastropods including *Bembexia sulcomarginata* and *Palaeozygopleura hamiltoniae* plus a variety of orthoconic cephalopods.

One unusual aspect of the preservation in this quarry is the fact that a very large percentage of the bivalves are preserved with both valves intact. This is no surprise for nuculids which are infaunal and typically are entombed in the sediments which prevents their valves from gaping open upon death. However, for bysally attached semi-infaunal forms like Gosseletia, mobile semi-infaunal forms like Grammysia and epifaunal genera like Ptychopteria, it is unexpected to find both valves intact. It would seem possible that quick burial may have been an intermittant, but relatively common cause of death in this assemblage. This interpretation is also consistent with the rather large escape burrows that are abundant in this unit.

Many faunal elements in this unit are restricted to single bedding planes that may repeat throughout the quarry. For example, Palaeozygopleura has been found on only three horizons, but within those horizons they may be very abundant. The axis of coiling of the shells of Palaeozygopleura in this quarry are randomly oriented allowing us to infer that they have not been aligned by current action. Yet twice as many are found in an aperture down position as in an aperture up position. Since all positions of this shell exhibit equal hydrodynamic stability, we infer that the shells were occupied at the time of death, and the orientation of the shells reflects the life position.

Other bedding planes are dominated by Bembexia sulcomarginata, a pleurotomarian which was probably an algae grazer or possibly a deposit feeder. Many of the shells of Bembexia in this unit are encrusted with a trepostomatous bryozoan (Leptotrypella). Most frequently it is the upper surface of Bembexia that is encrusted, though some specimens exhibit encrustation of the base while the spire remains clean. Very rarely is a specimen found in which the encrustation spreads very far from one side onto the other. We suspect that encrustation primarily developed on dead shells on Bembexia. Bembexia is a genus that is only found in the Marcellus and Skaneateles Formations of the Hamilton Group while Palaeozygopleura is found throughout.

Cum. Miles	**Comments**
6.90	Continue south on Swamp Road
9.60	Intersection with U.S. Rt. 20 in Village of Morrisville; turn left (east) then immediately right (south) onto River Road.
13.70	Turn right onto N.Y.S. Rt. 26 in Eaton, N.Y.
15.00	Turn left (south) onto Bradley Brook Road
17.40	Turn right (west) onto Soule Road
17.45	**Stop #3** - Quarry entrance on left. Park on side of road. Basal Moscow Formation.

Stop #3. This quarry exposes the "interior portion" of an upward-coarsening cycle of the Hamilton Group and can be generalized as consisting of an upper sandy facies dominated by bivalves and a lower silty facies dominated by brachiopods. Separating the two facies is a thin coquina consisting predominantly of the shells of Spinocyrtia granulosa.

The upper sandy unit is dominated by large epi-byssate bivalves and large semi-infaunal bivalves as well as larger spiriferoid brachiopods. There are also occasional specimens of the burrowing trilobite Trimerus (Dipleura) deKayi and large specimens of Phacops rana.

The lower silty portion of dominated by brachiopods, particularly Chonetes, Devonochonetes, Athyris Rhipidomella, and Mucrospirifer. Bivalves are also abundant, commonly infaunal nuculids or mobile semi-infaunal forms (such as Grammysia). The dominant trilobites in this facies are Greenops and smaller specimens of Phacops. Gastropods are represented by Glyptotomaria and Ruedemannia, the bellerophont Retispira and Palaeozygopleura. Bembexia drops out of the fauna after the Skaneateles. However, the Palaeozygopleura shells in this quarry are barely recognizable for the shells are completely encrusted by the trepostomate bryozoan Leptotrypella. The aperture is always open, but the shell is completely encursted all the way around which is quite surprising since Palaeozygopleura is a shell dragger, incapable of lifting its shell from the substrate. This leads us to conclude that either Leptotrypella was capable of extending its colony down into the sediment (which no modern bryozoan can do) or that there was a secondary occupant (a sipinculid?) of the dead Palaeozygopleura shell that was capable of elevating the shell off of the substrate so that the bryozoan could establish itself everywhere except the aperture. In some individuals the encrustations became much thicker on one side of the shell than the other, always the adapertural side is thickened, which leads us to believe that the secondary occupant is still occupying the shell, but the mass of bryozoans has forced him to resort to a shell dragging mode, thus killing the bryozoans on the apertural side of the colony.

There are also massive colonies of other trepostomatous bryozoans in the lower part of the quarry.

Cum. Miles	**Comments**
17.45	Turn around, head east on Soule Road
17.50	Turn right (south) onto Bradley Brook Road
18.40	Turn left (east) onto Geer Road
19.20	**Stop #4** - Geer Road - Quarry entrance on left. Park on side of road. Basal portion of Moscow Formation.

Stop #4. Stratigraphically and faunally this locality is similar to Stop #3. The faunal elements are essentially as described for the previous locality. Some of the rarer elements of the fauna of both quarries include Hyolithes, occasionally with helens and operculum in place, and two different phyllocarids with their bivalved cephalothorax and segmented abdomen with the telson composed of three spike-like projections.

A prominent surface near the basal portion of the section exposes a phosphate pebble bed containing abundant brachiopods and rare fish plates.

Cum. Miles	**Comments**
19.20	Continue east on Geer Road
20.10	Intersection with Kenyon Road. Continue east on Geer Road.
20.80	Intersection with Lebanon Reservoir Road. Turn left (east).
22.80	Intersection with River Road. Proceed across River Road onto Armstrong Road.
23.80	Intersection with Lebanon Street - turn left (N).
26.80	Main Street light - Village of Hamilton, intersection with N.Y.S. Route 12B.
27.00	Turn left (E) onto E. Kendrick Avenue - Pass entrance to Colgate University
27.20	Turn right (E) onto Hamilton Street. Colgate campus on right.
29.75	Bear left at "y" intersection
30.40	Turn left (N) onto Poolville Road.
32.10	Hamlet of Hubbardsville; continue north through intersection onto Quarterline Road.
32.90	Turn right (E) onto Rhodes Road.
33.50	Turn left (N) onto Cole Street (becomes Cole Hill Road).
35.40	**Stop #5** - Roadside Quarry and rubble: Mottville Member of Skaneateles Formation:

Sedimentology:

The exposure at this stop nicely exhibits the shaley siltstone-to-sandstone transition of the Mottville Member. The lowermost portion of the outcrop consists of relatively fissile to blocky shaley siltstones dominated by small brachiopods. The section coarsens rapidly upward to silty, fine-grained sandstones which exhibit some cross-lamination and rare hummocky bedding.

Fauna

This stop is dominated by the higher energy faunal elements that characterize the silty sandstones of the Hamilton Group. These include the large epifaunal byssate pectinoid bivalves including Glyptodesma, Leiopteria, Pterinopecten, Actinopteria and Limoptera as well as semi-infaunal bivalves such as Grammysia. The typical trilobite of the unit is the large burrowing trilobite Trimerus, whose punctate carapace comprises one of the dominant skeletal elements of this locality. Complete specimens are fairly common, most commonly not enrolled and not infrequently oriented on edge or on end as some other unusual position relative to the sediments. It is suggestive that the area was occasionally subjected to events, which involved rapid burial. Other common elements include the brachiopods Spinocyrtia, Mucrospirifer, Camarotoechia and Devonoproductus. The gastropods are not well preserved here, but include Bembexia and Palaeozygopleura.

End of Trip

(To return to Clinton, N.Y. proceed north on Cole Hill Road to intersection with Pleasant Valley Road (approx. 1 mile). Turn left (N) onto Pleasant Valley Road. Proceed north to intersection with U.S. 20. Turn left (W) onto U.S. 20. Proceed on U.S. 20 to intersection with N.Y.S. Rt. 12B. Turn right (N) onto N.Y.S. 12B and proceed north to Clinton.)

The Late Wisconsinan Glaciation of the West Canada Creek Valley

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INTRODUCTION

This field trip focuses on the glacial stratigraphy of the West Canada Valley and the interpretation of the Late Wisconsinan history of the region. The Late Wisconsinan stratigraphic record in the West Canada Creek Valley (Figure 1) is unique in central New York in terms of its completeness and in the complexity of the glacial events that it represents. The significance of such lithostratigraphic documentation of the glacial record is underscored by the lack of success in developing a regional glacial chronology based on morphostratigraphic relationships (Coates, 1976; Calkin and others, 1982). Correlations based solely on morphologic evidence are often ambiguous due to the time transgressive and composite nature of many glacial landforms. The glacial deposits of the West Canada Creek Valley provide an important lithostratigraphic basis for correlations to the well-developed time-stratigraphic and geologic-climate classifications in the eastern Great Lakes Region (Dreimanis and Karrow, 1972; Dreimanis and Goldthwait, 1973).

The West Canada Creek Valley is an area where the sedimentology of readvance deposits and relative chronology of simultaneous readvances from several different ice flow directions may be studied. Ice from the Ontario Lobe (Black River and Oneida Sublobes) and the Hudson-Champlain Lobe (Mohawk Sublobe) as well as Adirondack through flow and possible Adirondack ice cap sources encroached on this region during the Late Wisconsinan (inset on Figure 1).

The field trip is designed to give an introduction to as many of the glacial stratigraphic units as possible and to show their sedimentologic diversity. It can in no way, however, show all the stratigraphic relationships which were used to determine the glacial history of the region. This would require too many stops for a one day excursion. In addition, many exposures that would be more informative than those shown on this trip are not readily accessible or are inappropriate for a group of this size. Important localities which are not included as stops on the field trip are discussed in

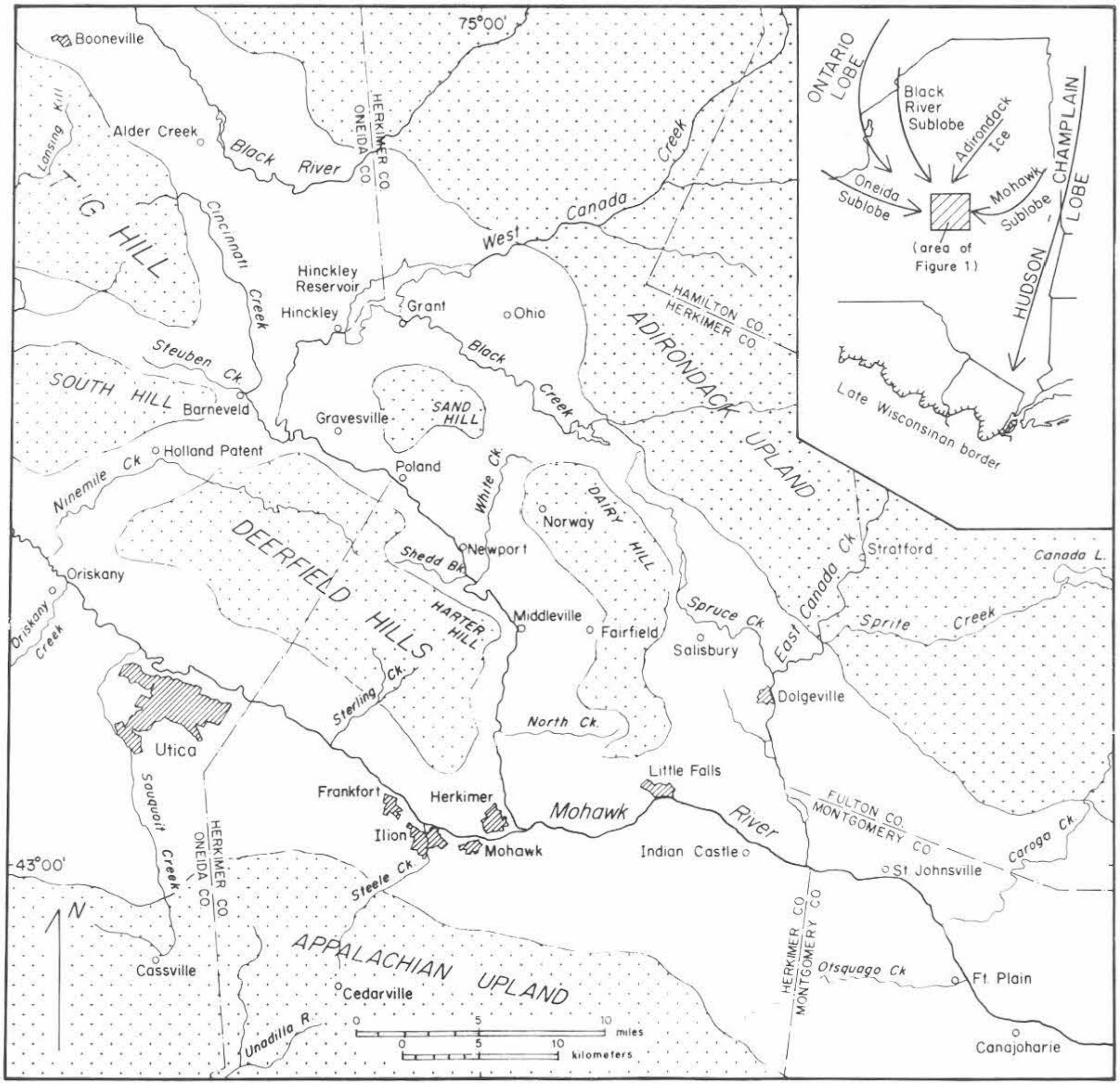


Figure 1. -- Location map of the West Canada Creek Valley and surrounding regions. Upland areas are patterned. This map serves as a base map for Figures 8, 9 and 11 thru 13. Inset map shows general flow direction of glacial lobes and sublobes that affected the region during Late Wisconsinan time and the maximum extent of Late Wisconsinan glaciation.

the text.

PREVIOUS WORKS

Even prior to the acceptance of the Glacial Theory, glacial features were recognized in the Mohawk Valley by Vanuxem (1842). The existence of a "Mohawk Valley glacier", which flowed from east to west in the Mohawk Valley, was first proposed by Dana (1863). Chamberlin (1883, 1888) cited striation evidence in support of a Mohawk Lobe moving eastward in the valley and this idea was later amplified by Brigham (1898, 1908). Chamberlin (1883) first recognized that the Adirondack Upland impeded ice flow and forced lobation of the ice-front into adjacent valleys. Miller (1909) reached a similar conclusion from his studies of ice movement and glacial erosion in the Black River Lowland. Brigham (1898) provided well-log data for deposits along the western Mohawk River and described morphologic features while Cushing (1905) provided brief descriptions of deposits in the Little Falls area. It was generally agreed that while the Mohawk Lobe showed evidence of having been active during advance, the area was probably deglaciated by downwasting and overall stagnation, a behavior that was responsible for thin drift on the valley sides and a conspicuous lack of morainic features (Brigham, 1911, 1929). Fairchild (1912) speculated on valley-wide ice-dammed lakes which extended into the Black River Lowland. These lakes drained across the Appalachian Plateau and were ponded by the receding Ontario and Hudson-Champlain ice masses. Brigham (1911, 1929) was critical of the lake chronology proposed by Fairchild. He recognized meltwater impoundment as playing a major role in the deglaciation of the Mohawk Valley at lower elevations (less than 200 m) but he saw no evidence for widespread lacustrine deposits at higher elevations (200 to 400 m) which should have existed if the proglacial lakes drained into the through-valley system of the Appalachian Upland.

Fullerton (1971) and Krall (1977) were the first to present stratigraphic evidence for large-scale readvances that occurred in the western Mohawk Valley region. Fullerton (1971) studied deposits near the eastern-most margin of the Ontario Lobe (Oneida Sublobe) in the Mohawk Valley. Some of the sections that he described are included on this field trip. Fullerton named the expansion of Ontarian ice the "Indian Castle Readvance" and correlated deposits of this readvance and the Valley Heads Moraines of central New York. Older deposits, some of eastern provenance (Mohawk Sublobe), were described in stratigraphic sections but Fullerton did not formulate a comprehensive glacial history for the pre-Indian Castle events.

Krall (1977) traced the Cassville-Cooperstown Moraine of the Mohawk Sublobe along the Appalachian Plateau from Cooperstown to Cassville where it is truncated by a younger Valley Heads Moraine of

Ontario provenance. He identified till of eastern provenance beneath Valley Heads deposits at least as far west as Clinton.

Studies in adjacent regions support the concept of readvance events in the western Mohawk Valley. Stratigraphic work in the Hudson Valley during the late 1960's and early 1970's has shown that the Hudson-Champlain Lobe experienced cycles of advance and retreat (Connally and Sirkin, 1973). Well logs in the Schoharie Valley (LaFleur, 1969) and stratigraphic sections in the Mohawk Valley (LaFleur, 1979) record at least two advances of the Mohawk Sublobe. The general character of Late Wisconsinan deglaciation in the Mohawk Valley is compatible with the record of Late Wisconsinan ice-front oscillations in the Great Lakes region (Dreimanis and Goldthwait, 1973; Frye and Willman, 1973; Wright and others, 1973; Johnson, 1976; Evenson and others, 1977b)

APPROACH TO THE PROBLEM

Detailed mapping and stratigraphic investigations in the Mohawk Valley throughout the past decade, coupled with recent developments in sediment-facies analysis have greatly influenced recent interpretations of the glacial history of the region. These interpretations have been summarized by Muller and others (in review).

Late Wisconsinan Age of Deposits

Earlier work in the West Canada Valley shows tills of possible Middle or Early Wisconsinan age (Fullerton, 1971). This concept stems from an idea that prevailed in the that Late 1960's that Late Wisconsinan glaciation reached its maximum extent north of where the border is now placed (inset on Figure 1). Recent studies in Pennsylvania and New Jersey show that Late Wisconsinan (Woodfordian) ice extended well into eastern Pennsylvania and New Jersey (Crowl, 1980; Crowl and Sevon, 1980; Cotter and others, in press). If this border is indeed Late Wisconsinan, it is unlikely that portions of the Mohawk Valley, Tug Hill or southwestern Adirondacks would have remained ice-free through the whole Late Wisconsinan. All the tills previously thought to be Early or Middle Wisconsinan deposits in the West Canada Creek Valley are now known to occur as the upper-most till deposits at some localities. Also, they are all associated with ice-marginal deposits which have not been subsequently overridden. Uncovering of the surfaces of these tills must post-date the Late Wisconsinan covering of the region. Therefore, all deposits thus far encountered in the West Canada Valley and shown on this field trip are considered Late Wisconsinan (Figures 2 and 3). In the absence of any absolute dates these stratigraphic relationships provide an explanation of this age designation.

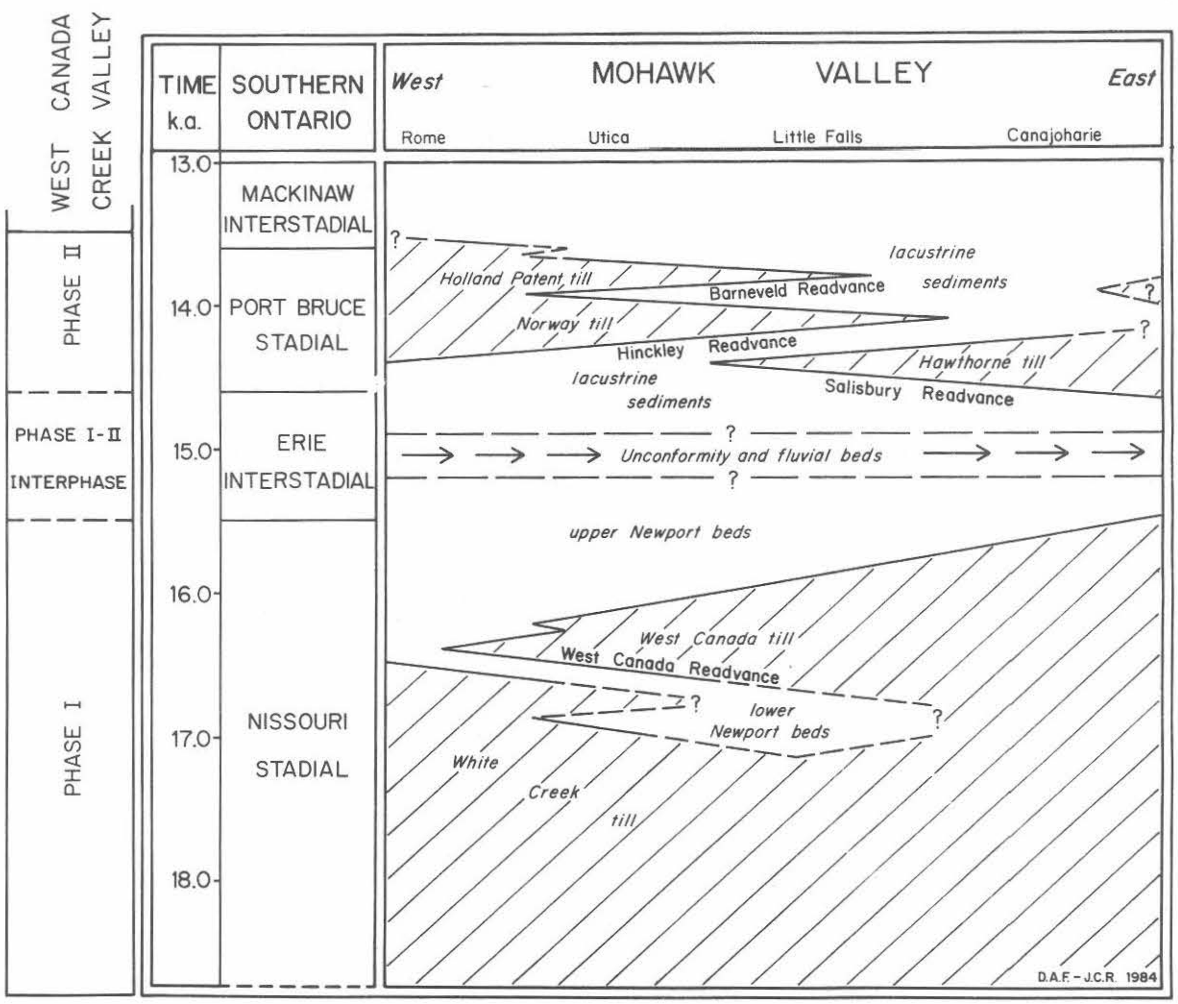


Figure 2. -- An east-west time-distance plot of readvances in the western Mohawk Valley (Muller and others, in review) with correlations to geologic-climate units of southern Ontario (Dreimanis and Karrow, 1972; Dreimanis and Goldthwait, 1973) as inferred by Muller and others (in review). Glacial lakes, tills and other deposits associated with readvances and retreatal phases are shown in italics. Column at left shows the deglaciation-phase terminology used in this report.

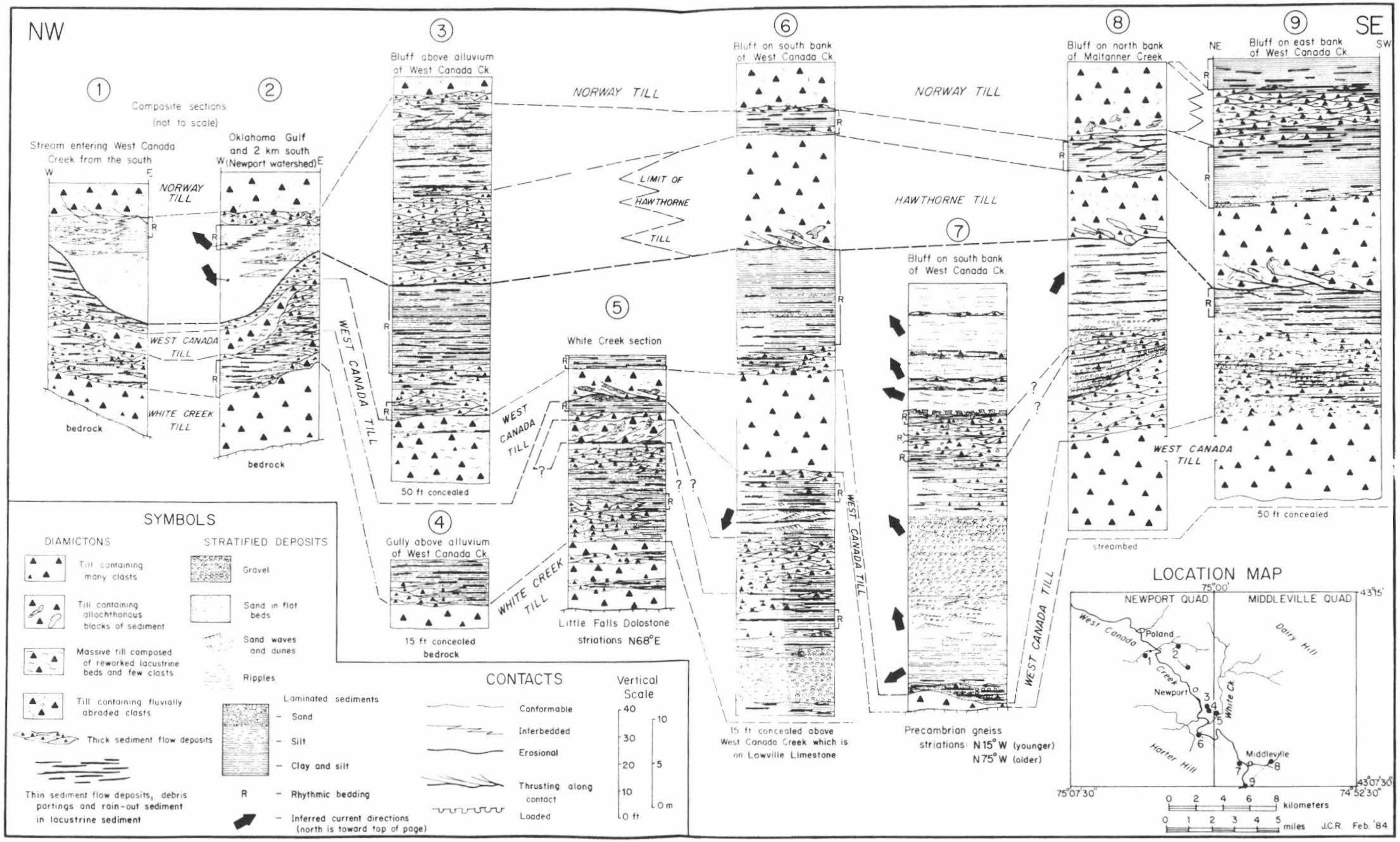


Figure 3. -- Representative stratigraphic sections from the West Canada Creek Valley (Newport and Middleville quadrangles) showing inferred correlations from northwest to southeast. Sections 1 and 2 are composite sections and are not drawn to scale. (STOP 1 is at Section 9; STOP 2 is at Section 5)

Provenance Investigations

A critical part of the formulation of the glacial history of the region is the development of provenance criteria for the different ice flow directions. Provenance data have been lacking in the West Canada Valley. Distinguishing the different source terranes in the field is very difficult and requires supporting laboratory analysis. Although compositional differences between till units have been observed within individual exposures, the observed variations are usually not consistent between widely-spaced localities. Lateral changes in underlying units may cause rapid changes in till composition. All provenance inferences must be made with supporting information at hand, i.e. bedrock maps, lithology of underlying glacial units and striation or till fabric (Foresti, 1984) data. Types of compositional data which have been used to distinguish till units are pebble and granule counts (Lykens, 1984), total carbonate analysis, dolomite and calcite percentages, major-element whole-rock geochemistry, and trace and minor-element analyses including Fe, Ti, Rb, Sr, Zr, Ba and Mn (Franzi, 1984). Heavy mineral data were used by Antonetti (1982) and Loewy (1984) but these data proved to be ambiguous for distinguishing between prospective metamorphic source terranes. The provenance of tills and lacustrine deposits is presently being investigated in the lower West Canada Valley using pebble counts, carbonate analysis and clay mineral X-ray analysis (Ridge, in prep.).

In the lower West Canada Valley, lacustrine sediments of Oneida Sublobe provenance can be recognized in the field as those that contain rain-out sediment which is red. In general, lacustrine sediments of Ontario provenance are lighter in color than Mohawk Sublobe sediments. Lacustrine sediments derived entirely from the Mohawk Sublobe do not contain red rain-out sediment and are darker in color. Rhythmites between the Hawthorne till and lacustrine sediment flows associated with the Norway till (STOP 1; Section 9 on Figure 3) show these differences. The lower half of the rhythmites contains upward-thinning varve couplets and drab, dark gray colors. Midway through the rhythmites, the couplets begin to thicken upward, show lighter colors, and contain red drop sediment. This succession shows a transition in the varves from an eastern (Mohawk Sublobe) provenance at the base to a western (Oneida Sublobe) provenance at the top.

In the western Mohawk Valley, Ontario Lobe tills may have a red color because of Vernon Shale and red lacustrine sediments that underlie these tills to the west. Tills from the east or of Mohawk provenance are never red-colored and have a more drab appearance. Unfortunately, these simple criteria cannot be used in the West Canada Valley where both Mohawk and Oneida Sublobe tills are gray.

Sediment - Facies Analysis

The sedimentologic distinction between till and sediment flows in diamict sequences is important to the reconstruction of the glacial history of the West Canada Valley. Particularly where deep lacustrine waters (as deep as 225 m) bordered the Late Wisconsinan ice sheet, thick sediment flows may have been deposited directly from the glacier and from valley sides. Delineation of the physical limits of till sheets requires the ability to differentiate tills from sediment flows. The works of Boulton (1968), Evenson and others (1977a), Lawson (1979), May (1977), and Hicock and others (1981) serve as guides to the recognition of sediment flows and tills.

Gravel deposits in the Mohawk Valley near Little Falls and in the West Canada Valley, previously identified as outwash (subaerial) deposits (Fullerton, 1971), are now recognized as subaqueous outwash (Rust and Romanelli, 1975) or subaqueous fan deposits (Boothroyd, 1984). The distinction between subaerial and subaqueous gravels is critical to discerning periods of lacustrine drainage and interstadial episodes.

Paleomagnetic Investigations

Analysis of paleomagnetic declinations preserved in fine-grained lacustrine sediments have provided an important time-based stratigraphic tool for testing correlations between nearby stratigraphic sections. Secular declination changes measured in the depositional remanent magnetism (DRM) are used for testing time equivalence between units. This technique is particularly applicable to the West Canada Creek Valley where exposures of suitable lacustrine sediments span most time intervals, exposures are abundant, and reproducibility can be demonstrated for the lacustrine units. Metamorphic rocks in the Adirondacks supply an abundance of fine-grained magnetic minerals (predominantly magnetite) to the glacial deposits of the area. For this reason remanent signals have high magnitudes and are very stable. Remanent signals from Mohawk and Adirondack provenance deposits have on the average 10 to 100 times higher magnitudes than Oneida Sublobe deposits. These differences probably result from disimilar magnetic mineral concentrations and provide another criterion for provenance studies. A summary of preliminary paleomagnetic results is shown on Figure 4.

Paleomagnetic secular variation curves may be used to test regional correlations within distances of about 500 kilometers from the study area. The western Mohawk Valley data are being compared to declination data from the Genesee River Valley of western New York (Brennan and others, 1982; Brennan and others, in prep.; Braun and others, 1984). Presently, the declination records from both

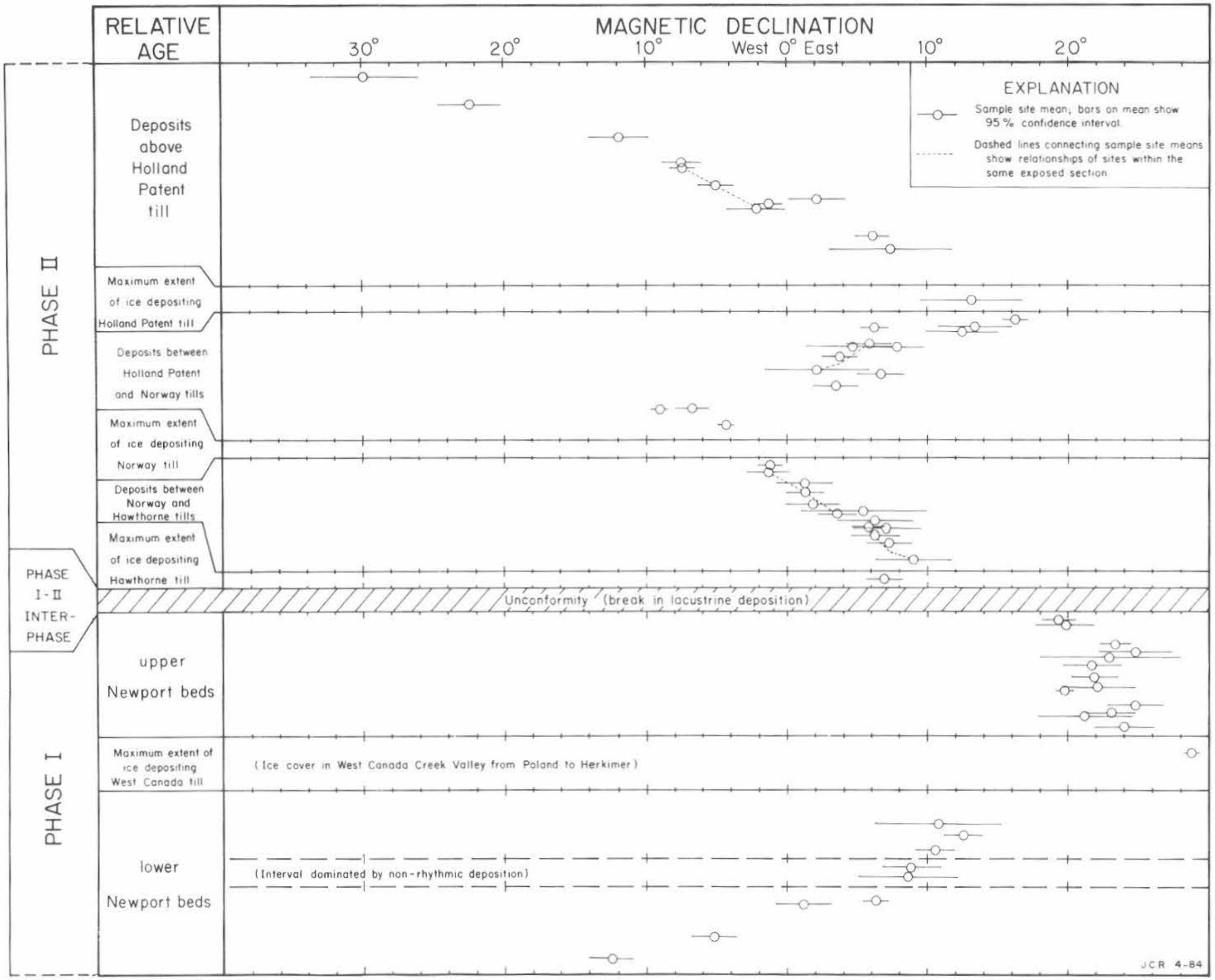


Figure 4.-- Paleomagnetic declination plotted against relative time as determined from relative ages of units from which sample cores were taken. Each point on the plot represents the mean value of 7 to 16 cores taken at one sample site or horizon.

areas are too incomplete to allow a decisive check on correlations. For periods during which the records may overlap, as indicated by correlations within the Valley Heads Moraines, the declination records are in agreement.

GLACIAL STRATIGRAPHY

The glacial stratigraphy of the West Canada Valley can be subdivided into two units of glacigenic deposits separated by an erosional unconformity and fluvial beds. The two units represent separate phases of Late Wisconsinan ice-front oscillation during general deglaciation. Deposits within the units are conformable except for local erosional disconformities which resulted from overriding glaciers. For discussion purposes only, these units will be referred to as deposits of: Phase I, the lower set of deposits, and Phase II, the upper set of deposits. The interval between them will be referred to as the Phase I - Phase II Interphase (Figure 2). The use of the terms phase and interphase are informal and without proven relationship to the geologic-climate classification of Dreimanis and Karrow (1972).

Phase I

White Creek Till

The lowest unit of Phase I deposition is the White Creek till which typically is stony, sandy and compact and contains a high proportion of metamorphic clasts. In exposures the lower contact of the till either lies directly on bedrock or is concealed. The till contains no reworked lacustrine sediment except in exposures along White Creek (STOP 2) where the upper part of the till is interbedded with sediment flows and laminated lacustrine deposits. Ice-flow indicators (till fabrics and striations) associated with the White Creek till indicate variable ice flow ranging from south to east. Southerly flow directions probably represent overriding of the area by thick ice, a conclusion that is supported by high proportions of Adirondack lithologies in the till. Southeasterly flow directions probably developed as a result of topographically controlled ice flow which is reflected by a greater proportion of local bedrock clasts in the till. Along White Creek (near STOP 2), striations (N70E) indicate flow from the west (Oneida Sublobe). These deposits may reflect a local oscillation of valley controlled ice flow at the margin of the receding Oneida Sublobe.

Lower Newport Beds

Lacustrine units of the lower Newport beds overlie the White Creek till. The unit generally becomes finer-grained upward and consists of cobbly to silty sediment flows, laminated silts and sands, and silty rhythmites. Along the headwater tributaries of White Creek the rhythmic couplets thin upward above the White Creek till but thicken and become more clayey near the base of the West Canada till. The occurrence of diamict layers of possible glacial and sediment-flow origin, suggests a glacier-proximal environment and possible readvance during deposition of the lower Newport beds (Sections 5 and 6 on Figure 3). In the upper White Creek Valley this part of the lower Newport beds is represented by thick turbidity and debris flow deposits that separate a lower silty rhythmite facies from an upper clayey facies. No provenance data or ice-flow indicators are available but current directions inferred from megaripples in sandy beds associated with the unit suggest a western source or Oneida Sublobe affinity (Section 6 on Figure 3).

West Canada Till

The West Canada till is an eastern (Mohawk Sublobe) provenance till which lies between the upper and lower units of the Newport beds. Facies of the till are highly variable in the West Canada Creek Valley from Middleville to Poland. Near Middleville (STOP 1; Sections 7, 8 and 9 on Figure 3) it is dark gray to black, stony and has a silty matrix. To the northwest in the West Canada Creek Valley the unit is characterized by a sparsely stony, silty to clayey till facies (STOP 2; Sections 1 thru 6 on Figure 3). The silty and clayey facies is displayed along the tributaries of White Creek and in the Shedd Brook area where the till overlies clayey sediments of the lower Newport beds. The increase in clay content and decrease in stoniness is in the inferred direction of ice flow. At higher elevations along the southwestern flank of Dairy Hill, the West Canada till exhibits a stony character and is sandier than in the West Canada Valley. This facies may reflect thinner and sandier sequences of lacustrine sediments beneath the till in upland areas or greater contributions of metamorphic Adirondack lithologies along the northern margin of the till sheet in upland areas.

Upper Newport Beds

The upper Newport Beds are dominated by upward-thinning, clayey and silty rhythmites which contain rippled sand beds and turbidites (Sections 3 and 6 on Figure 3). Thick sequences (as thick as 40 m) of subaqueous outwash sands and gravels and sediment flows may be found associated with depo-centers at the receding Mohawk Sublobe ice front. Subaqueous fan deposits are well exposed along West Canada Creek at Middleville (Section 7 on Figure 3). Current directions in these deposits indicate a source to the southeast. At many other sections in the West Canada Creek and White Creek valleys paleocurrent flow to the northwest has been inferred from rib and furrow structures and flutes on the parting planes of finer-grained laminated silts and clays.

Meltwater Deposits in the Upper West Canada Creek Valley

Recession of the Black River Sublobe during the waning stages of Phase I is recorded by proglacial sediments in two exposures in the Cincinnati Creek Valley, approximately 1.5 km east-northeast of Barneveld (STOP 4; Figure 5). The Phase I sediments at both exposures are truncated by the Norway till (Phase II) which, in turn, is overlain by younger lacustrine and deltaic sediments. The variability in the stratification and the texture of the Phase I lithofacies, the presence of lacustrine sand and rhythmites within the stratigraphic sequences, and the strong, unidirectional paleocurrents are consistent with the interpretation of these deposits as subaqueous outwash (Rust and Romanelli, 1975) or subaqueous fan deposits (Boothroyd, 1984). Paleocurrents, clast provenance, and upward fining of texture suggest that the units may have been deposited as esker fans from a northward-retreating (Black River Sublobe) ice front (Figure 6).

The configuration of the proglacial lake into which the Phase I sediments at Cincinnati Creek were deposited must be inferred from isolated deposits which survived subsequent glacial events. Ice-marginal delta deposits at Bailey and Ninety Five hills (STOP 6), approximately 10 km northeast of Barneveld, record an episode of deltaic sedimentation, from Black River Sublobe source, into glacially ponded water at an elevation in excess of 475 m. These deposits are almost 33 m higher than the deltas deposited into glacial Lake Miller (Phase II), and thus, record an earlier, higher level lake phase. Accordant elevations on Sand Hill, a kame (delta?) complex 1 km northeast of Cold Brook, suggest that deposition was controlled by the same lake level. These ice-contact drift facies are probably closely synchronous with the deposition of a thick sequence of lacustrine rhythmites which unconformably underlie Phase II deltaic sediments in bluff sections along the southern shore of the Hinckley Reservoir. These data indicate the existence of an extensive high-level lake in the upper West Canada Creek Valley during the initial stages of deglaciation. Such high levels of impoundment require extensive ice dams in the Oneida and Mohawk lowlands in order to prevent southward drainage through lower cols of the Appalachian Upland. The impoundment, therefore, may be associated with the readvance of the Mohawk Sublobe which deposited West Canada till in the lower West Canada Creek and Mohawk Valleys.

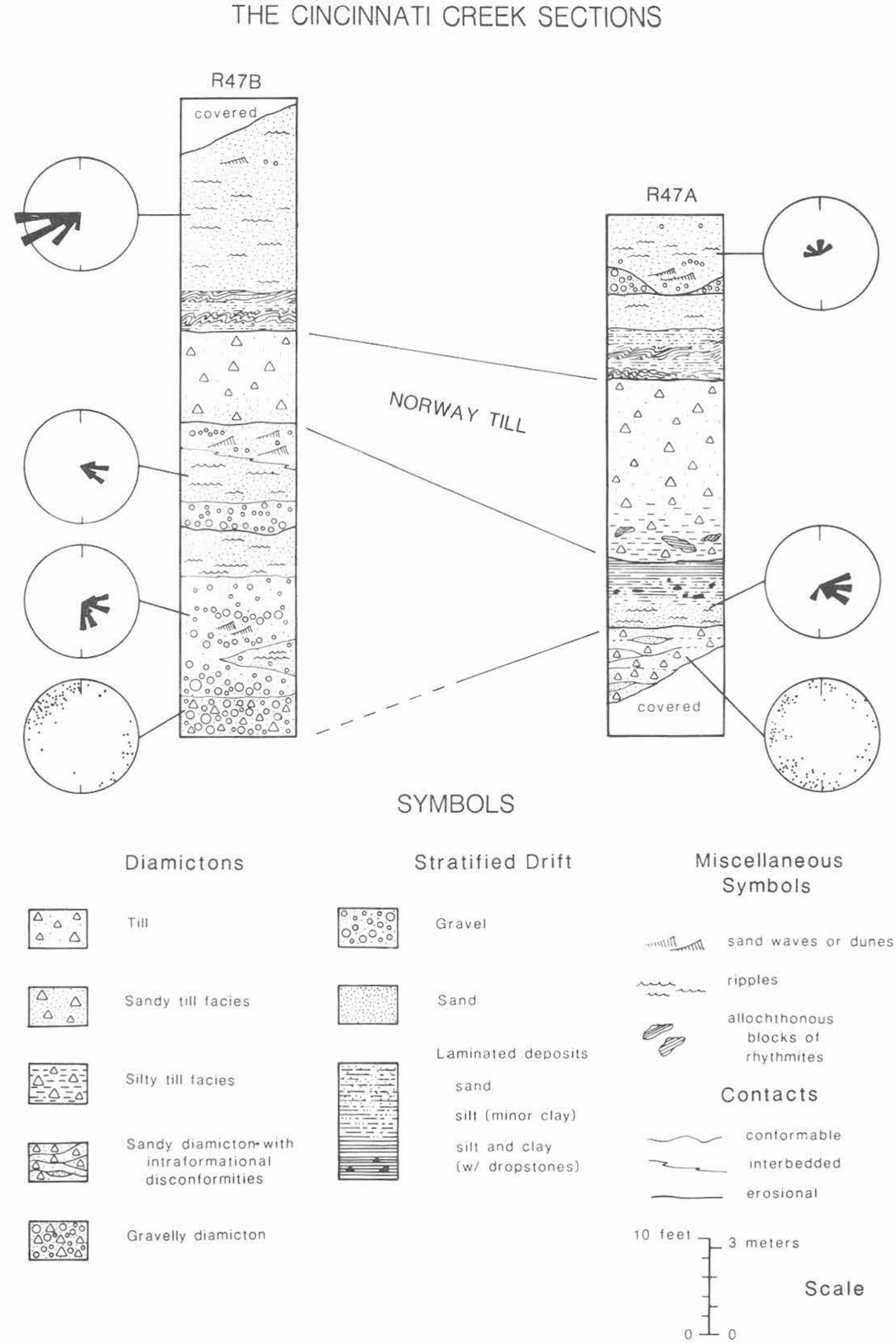


Figure 5. -- Representative stratigraphic sections of the exposures along Cincinnati Creek, showing inferred correlations. Dip-line plots of elongate clasts and rose diagrams are drawn with north toward the top of the page. The outer circles of the rose diagrams represent five current measurements.

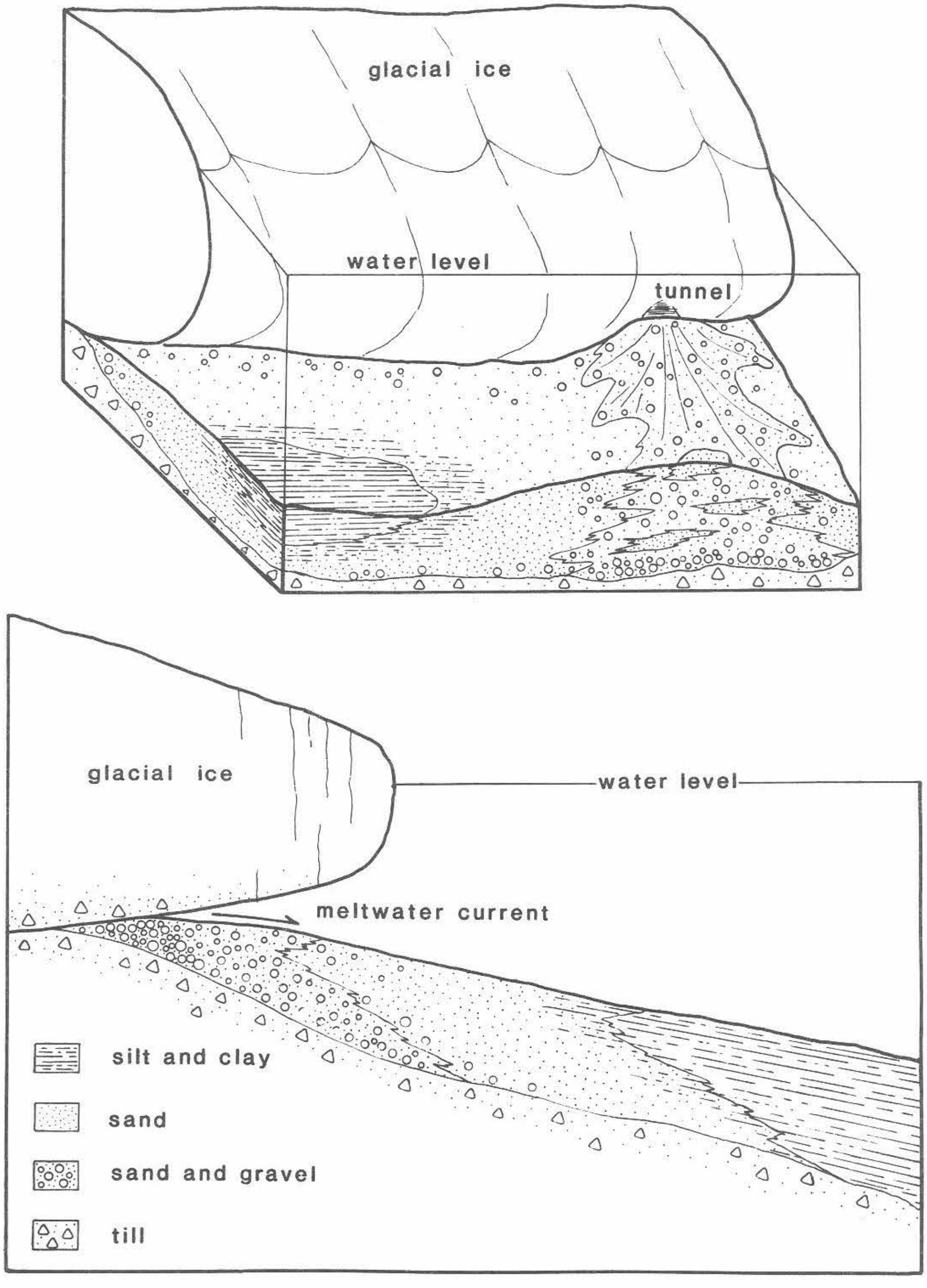


Figure 6. -- Inferred esker-fan sedimentation interpreted from sediment facies observed in sections along Cincinnati Creek (STOP 4; Figure 5) in Phase I sediments.

Phase I - Phase II Interphase

Deposits which underlie the Hawthorne till show evidence of an erosional unconformity and fluvial deposition. Evidence for the erosional period, here referred to as the Phase I-Phase II Interphase, is outlined below and in Figure 7.

1. The nature of the lower contact of the Hawthorne till suggests that the underlying sediments were subglacially eroded and deformed by the glacier that deposited the till (STOP 1; Sections 6 and 9 on Figure 3). At many exposures the till truncates rhythmites of the upper Newport beds, which contain upward-thinning varve couplets, and no intervening proglacial sediments are present. The apparent lack of proglacial sediment may indicate extensive subglacial erosion as the Mohawk Sublobe advanced into the West Canada Creek Valley. Alternatively, proglacial sedimentation may have been focused into the deeper portions of the lake basin by density underflows emanating from the ice-front. The apparent lack of proglacial sediment may reflect non-deposition in the higher elevations of the lake basin as well as subglacial erosion.

2. Outside the limits of the Hawthorne till, proglacial lacustrine sediments, which underlie the Norway till and are, in part, time-equivalent to the Hawthorne till, show an unconformable contact with the upper Newport beds (Sections 1 and 2 on Figure 3). Within the limits of the Hawthorne till sheet, they have a conformable contact with the top of the Hawthorne till (Sections 6, 8 and 9 on Figure 3). Immediately beyond the limits of the Hawthorne till, sediment flows which appear to be proglacial sediments related to the Hawthorne till have an unconformable relationship with the Newport beds (Section 3 on Figure 3).

3. In the Mohawk Valley, 6 km south-southeast of Little Falls, the Hawthorne till and associated sediment flows overlie fluvial gravels (Lykens, 1984). The gravels display crossbedding and imbrication which indicate a paleocurrent flow to the east during a period of subaerial sedimentation. The upper contact of these gravels marks a transition upward from fluvial to lacustrine conditions during a readvance of the Mohawk Sublobe.

Differences in the thickness and physical character of glacial sediments in the West Canada Creek and Mohawk River valleys are probably related to differences in elevation of exposures in the two areas. In the West Canada Valley, the upper surface of the Newport beds is only exposed at elevations of 213 to 381 m while fluvial gravels in the Mohawk Valley are exposed at elevations no higher

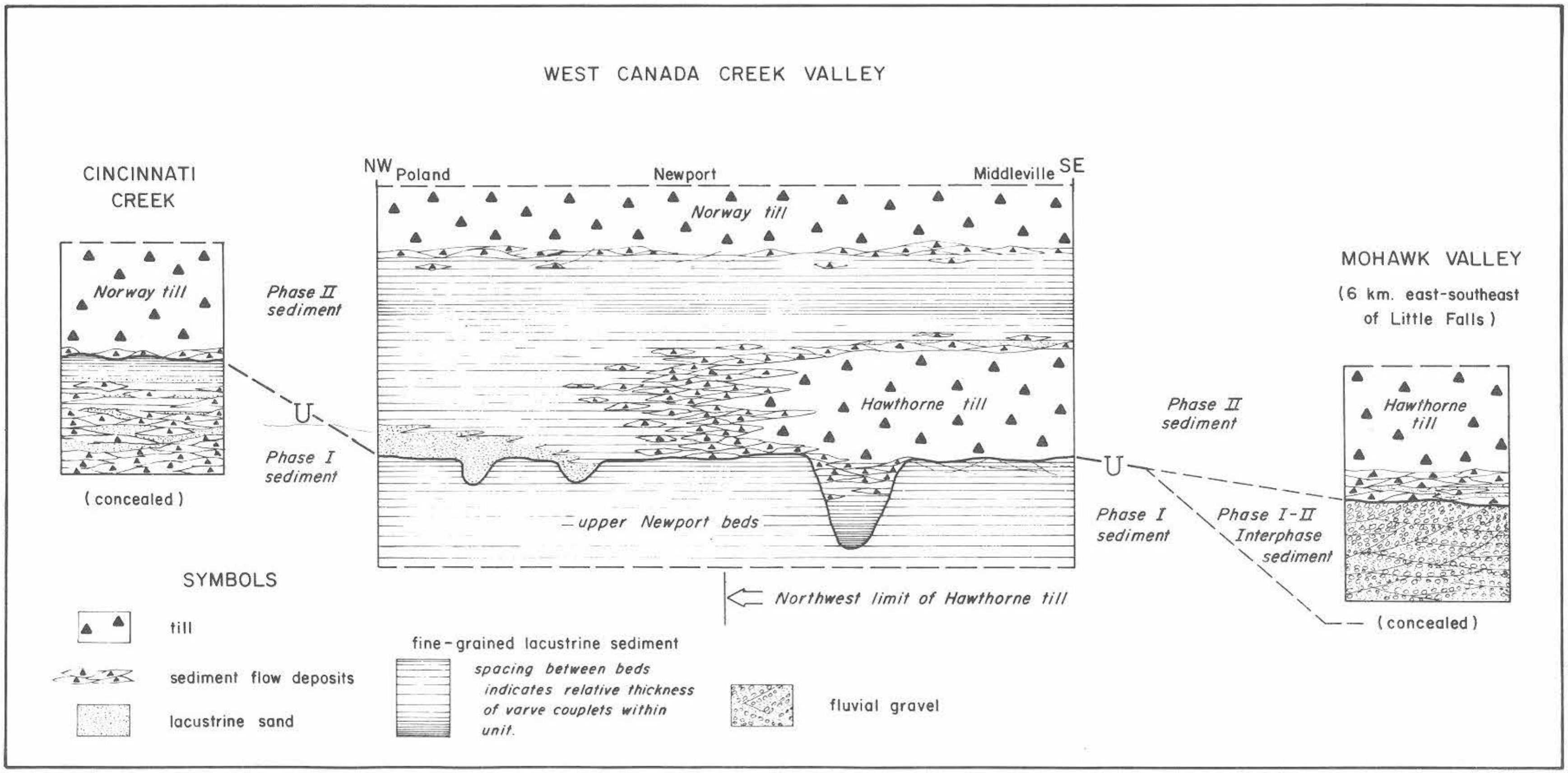


Figure 7. -- Generalized cross sections and stratigraphic relationships of the unconformity and fluvial beds of the Phase I-II Interphase showing the underlying and overlying units in the Cincinnati Creek, West Canada Creek and Mohawk River valleys.

than 122 meters. Subaerial exposure during the Phase I-Phase II Interphase resulted in erosion and dissection of deposits in the West Canada Creek Valley while fluvial gravels were deposited in the Mohawk River Valley.

Phase II

Hawthorne Till

The Hawthorne till is a black to dark gray silty, moderately stony till dominated by black shale clasts and containing rare sandstone clasts (Sections 6, 8 and 9 on Figure 3). To the northwest, the limit of the Hawthorne till is marked by a transition from till to sediment flows (Section 3 on Figure 3). Further up-valley the sediment flows probably correlate with a rhythmite sequence that has a distinctly dark, drab color and northwesterly paleocurrents (STOP 3; Sections 1 and 2 on Figure 3). Laminated and rippled sands below the rhythmites probably represent lacustrine deposition by drainage into the upper West Canada Valley from the north prior to the full impoundment of the valley by the Mohawk Sublobe. A reconstruction of ice deployment and proglacial lakes at the maximum extent of the Mohawk Sublobe during Phase II is shown on Figure 8.

Norway - Hawthorne Interval

Rhythmites deposited between the Hawthorne and Norway tills and widespread at elevations of 210 to 260 meters (STOP 1; Sections 6, 8 and 9 on Figure 3) have been described previously in relation to provenance criteria of the Oneida Sublobe (p. 7). The rhythmites record, through variations in composition and varve couplet thickness, a transition from an eastern source (Mohawk Sublobe) to a western source (Oneida Sublobe). At section 9 (Figure 3) rhythmites between the Hawthorne till and sediment flows related to the Norway till record continuous deposition for about 160 years.

Norway Till

In the West Canada Creek Valley, the Norway till is a clayey to silty, sparsely to moderately stony gray till which contains lithologies from an Oneida Sublobe source. The limit of the Norway till closely coincides with ice-marginal deltas deposited in Lake Miller, an impoundment in the upper West Canada Creek Valley at the former confluence of the Oneida and Black River Sublobes (Figure 9). Lacustrine sands and numerous ice-marginal deltas of the Hinckley Moraine System mark the maximum extent and recession of the Oneida and Black River Sublobes near their suture. To the southeast, the limit of the Norway till is traceable across the East Canada Creek

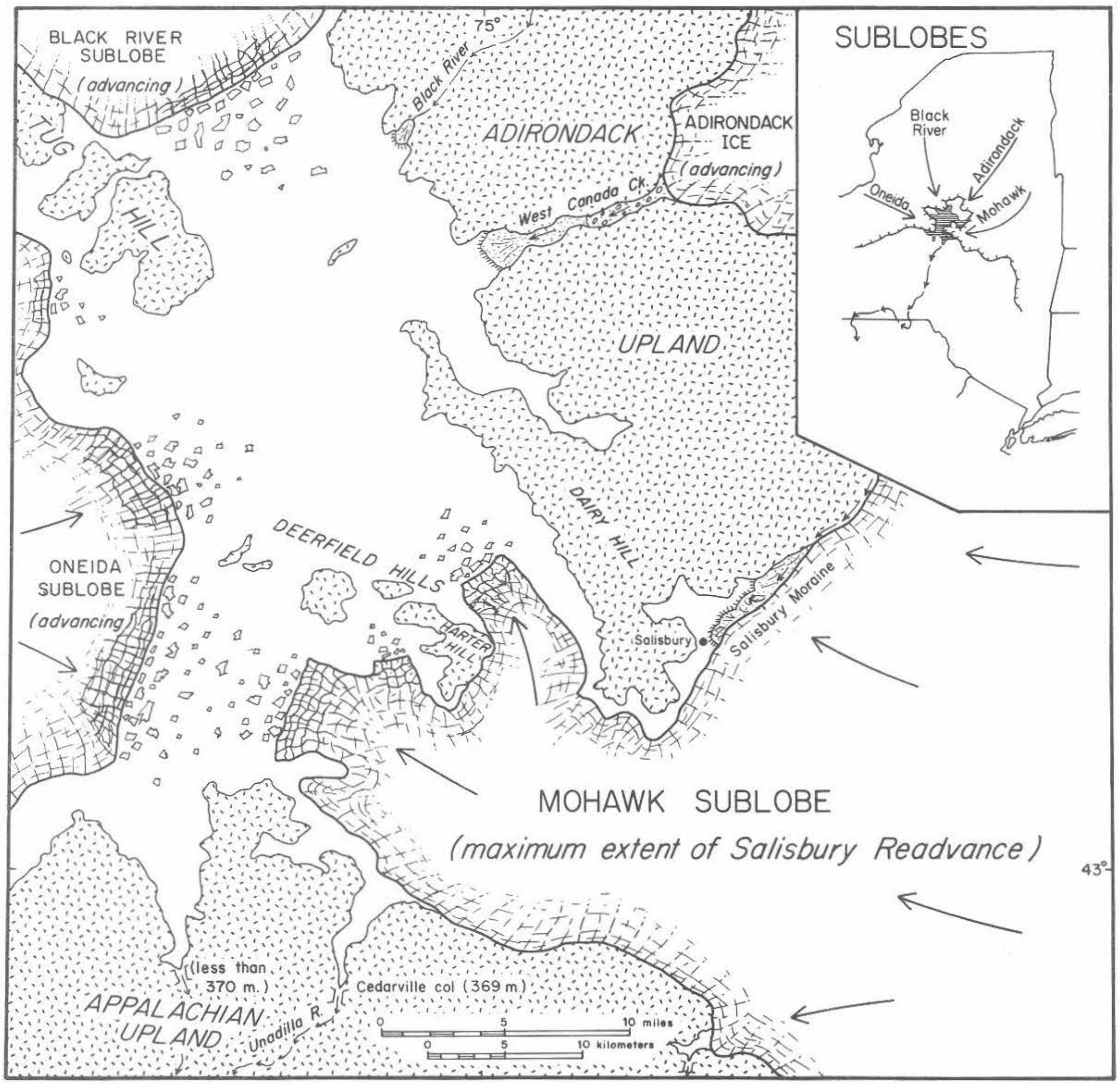


Figure 8. -- The inferred deployment of ice at the time of the maximum extent of the Salisbury Readvance of the Mohawk Sublobe. The Black River, Oneida and Mohawk Sublobes are drawn with calving margins where they terminated in deep water. Cedarville col may have been the outlet for ponded water in the region but the col to the west, although now higher than Cedarville col, may also have served as an outlet. (Base map is Figure 1)

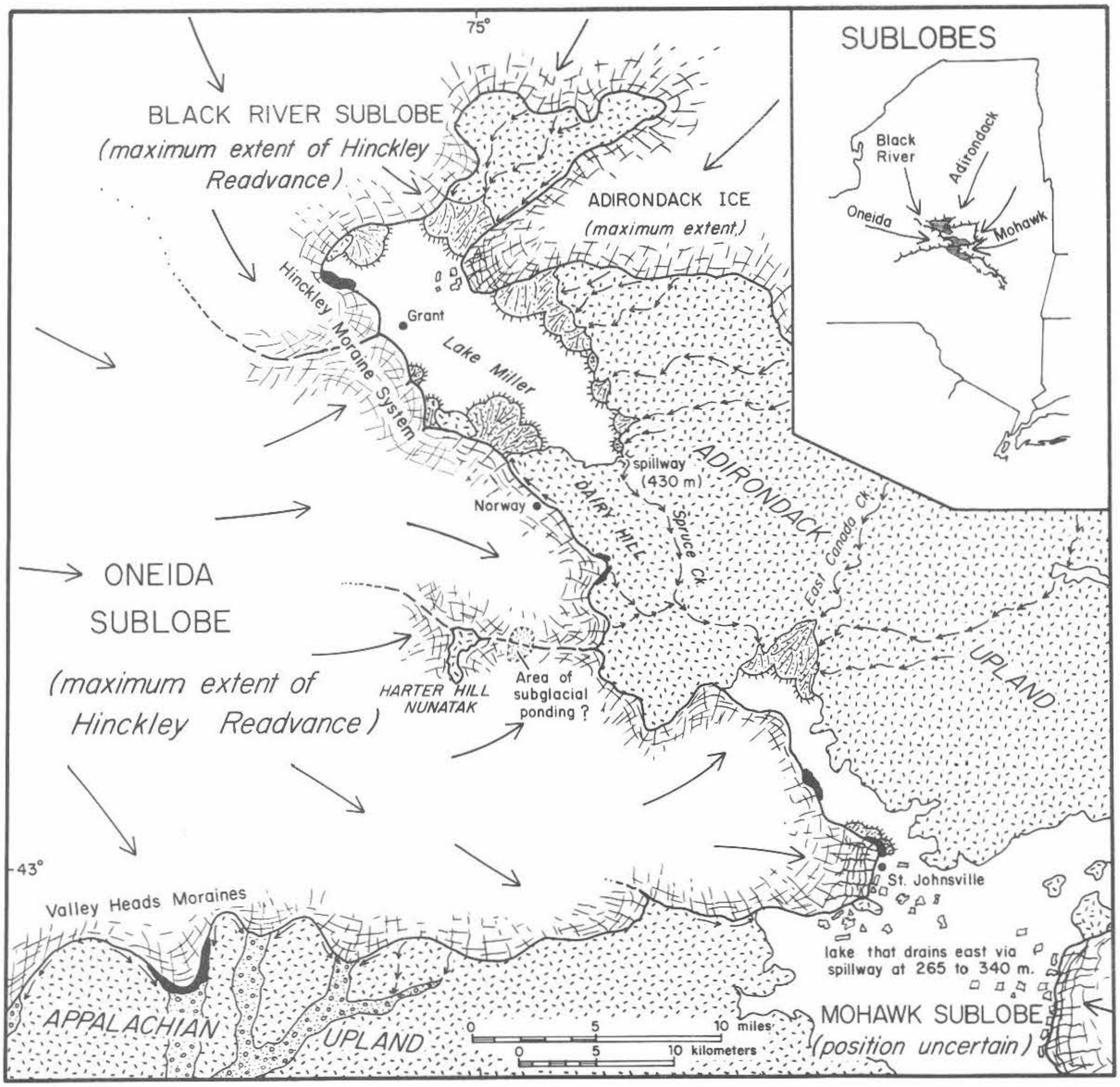


Figure 9. -- The inferred deployment of ice at the time of the maximum extent of the Hinckley Readvance of the Black River and Oneida Sublobes which ponded glacial Lake Miller in the upper West Canada Creek Valley. The exact position of the Mohawk Sublobe is unknown, but it remained far enough west to pond water above an elevation of 265 meters in the Mohawk Valley. Adirondack ice may have reached the maximum position of a readvance into Lake Miller. Till stratigraphy confirming this hypothesis is lacking and this interpretation is based on morphostratigraphic evidence from deltaic deposits at the readvance limit. (Base map is Figure 1; Blackened area at glacier terminus represents either an end moraine or large kame deposit)

Valley to the Mohawk Valley near St. Johnsville (Lykens, 1984).

The sequence of sediment flows at Section 9 (Figure 3; STOP 1) correlate with the Norway till at Section 8 (Figure 3). The close association of the sediment flows and the Norway till suggest that they may have been deposited in an isolated body of water, possibly subglacially ponded, near the confluence of ice tongues of the Oneida Sublobe which flowed around Harter Hill (Figure 9). This configuration of ice masses is consistent with the hypothesis that Harter Hill stood as a nunatak at the Hinckley Readvance maximum. Northeast of Middleville, the Norway till limit occurs between elevations of 400 and 425 meters. On the flank of Harter Hill, an ice margin at this elevation would leave part of the hill uncovered. Coalescing ice in the West Canada Valley may not have developed enough thickness or persisted long enough to fully expel lacustrine waters trapped beneath it in the vicinity of Section 9 (Figure 3; STOP 1).

Alder Till

The Alder till is a Black River Sublobe equivalent of the Norway till and is highly variable in composition and texture. In the southwestern Black River Lowland the till is dark gray, moderately calcareous, and has a silty clay to clay loam matrix. Clasts of dark gray siltstone and shale and gray limestone reflect an eastern Tug Hill provenance. The absence of rounded red and green Medina Sandstone cobbles serves to distinguish the silty facies of the Alder till from the Norway till (Oneida Sublobe) with which it is closely associated. In the eastern Black River Lowland the unit is a light gray to gray-brown, slightly calcareous to noncalcareous till in which subangular to rounded clasts of various metamorphic rock types dominate. Because of its loose and permeable texture, the sandy facies of the Alder till is poorly exposed. Transitional facies of these two lithotypes are uncommon which may reflect lack of mixing of source materials in the Black River Lowland or the burial of transitional facies beneath younger lacustrine sediments.

The maximum southeastern extent of the Black River Sublobe during the Hinckley Readvance (Figure 9) is evidenced by exposures of silty diamicton over lacustrine sediments in bluffs on the shore of Hinckley Reservoir, 3.2 km northeast of Hinckley (STOP 5; Figure 10). The intense intrastratal fluidization and southward-directed shearing observed in the lacustrine unit is interpreted as glaciotectonic in origin, and thus, the diamicton is considered to be a lacustrine facies of the Alder till. The upper sandy facies of the diamicton probably reflects the waning stages of the readvance with much of the deposition occurring beneath an alternately grounded and ungrounded glacier sole. These upper facies were probably subject to a variable amount of resedimentation within the

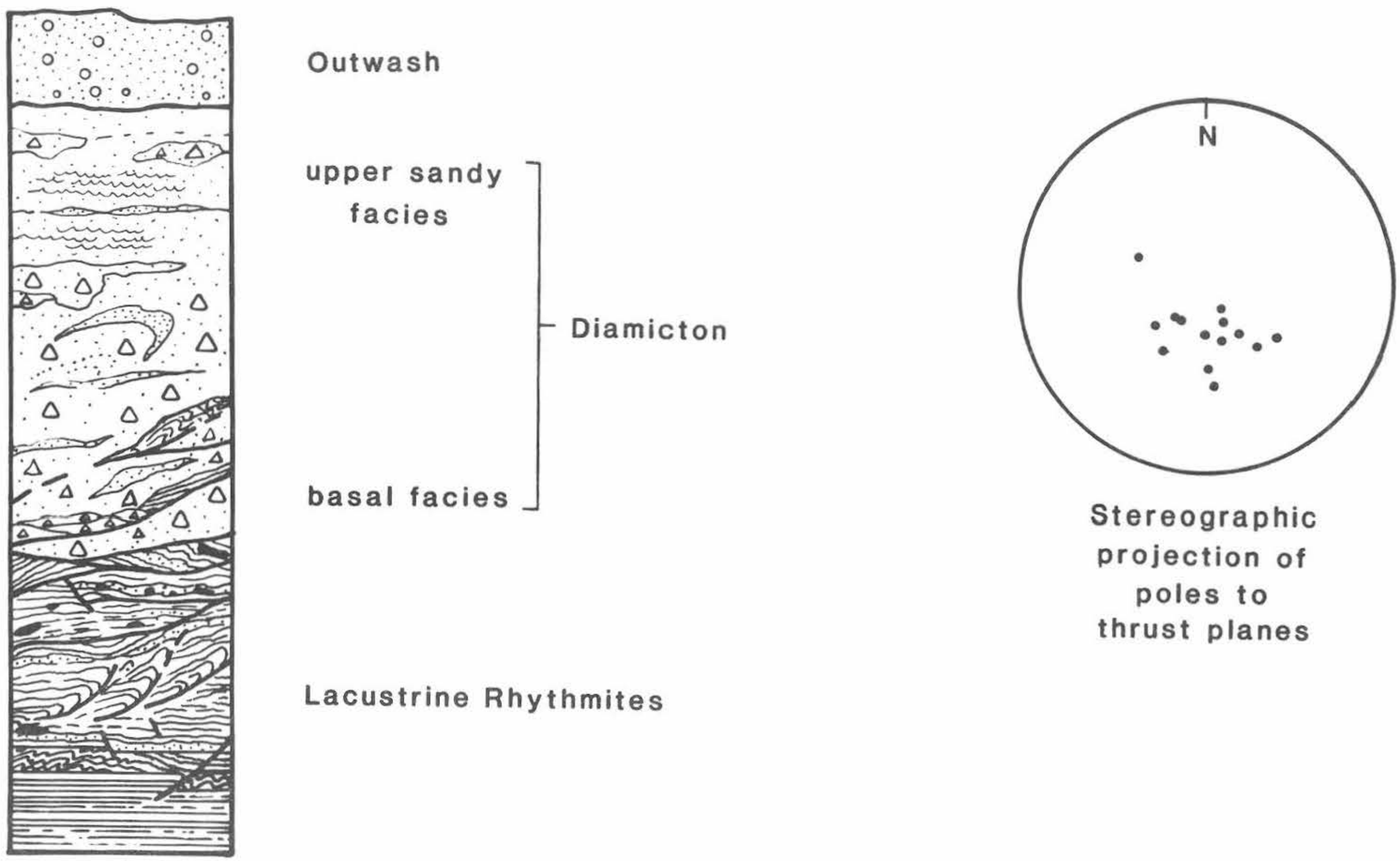


Figure 10. -- Stratigraphic section exposed in bluffs on the south shore of Hinckley Reservoir, 3.2 km northeast of Hinckley (STOP 5). Section shows silty diamicton overlying lacustrine sediments. Stereographic projection is a plot of poles to thrust planes measured in the lacustrine sediments. Symbols used are the same as those in Figure 5.

lake basin.

Norway - Holland Patent Interval

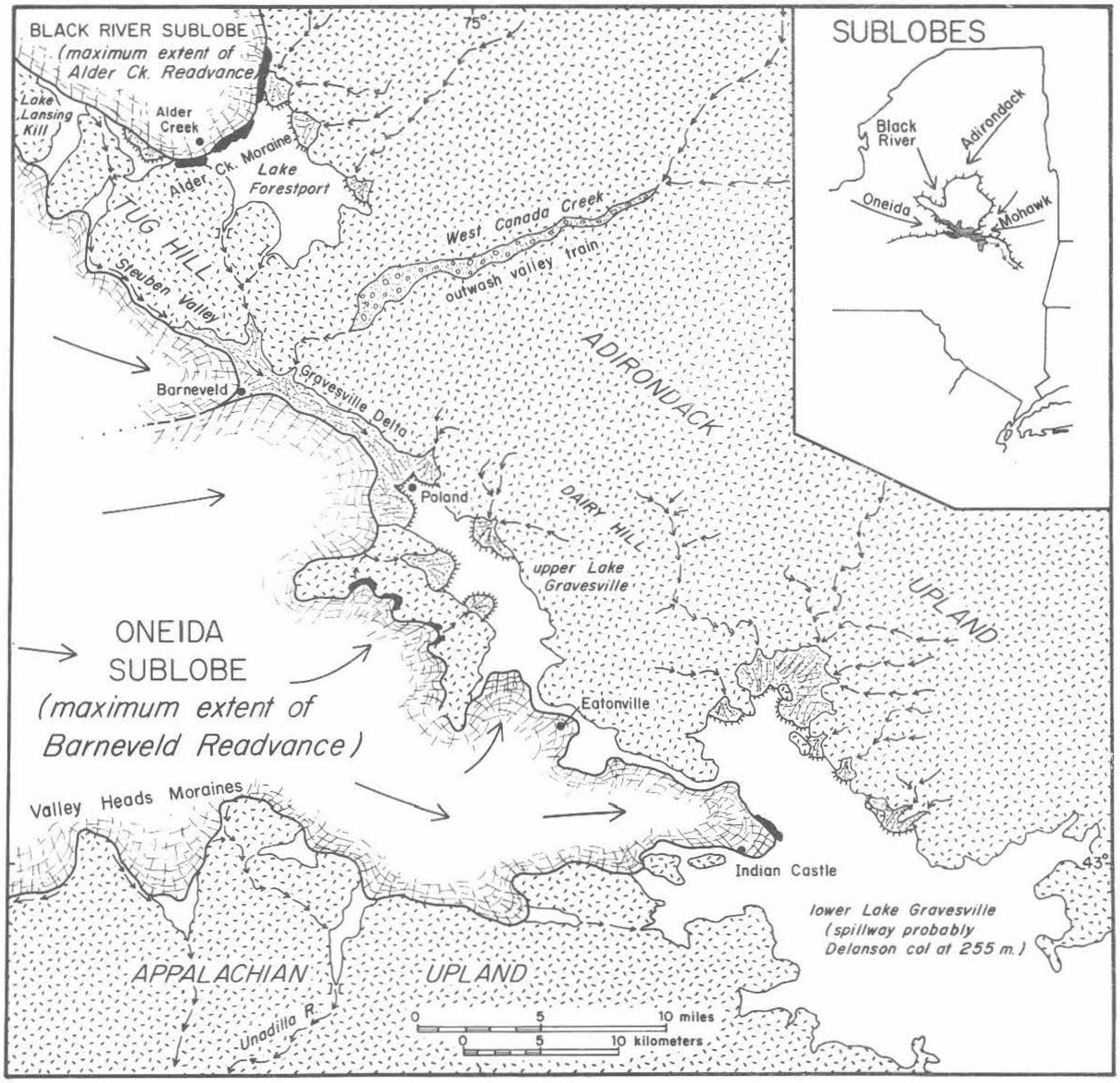
Rhythmites were deposited on top of the Norway till and its associated sediment flows as the Oneida Sublobe receded (STOP 1; section 9 on Figure 3). Varves between the Norway and Holland Patent tills in exposures along North Creek, 0.7 km east of Eatonville, record at least 140 years of continuous lacustrine deposition (Flick, in prep.). Sediment flows occur within the varves at the top of the Norway till and varve couplets thin upward. The varve couplets thicken upward beneath the Holland Patent till and contain sediment flows just below the Holland Patent till. The composition of the varves and sediment flows indicates a western (Oneida Sublobe) provenance.

Holland Patent Till

The Holland Patent till is a clayey, sparsely to moderately stony, gray till containing lithologies of Oneida Sublobe provenance. It is difficult to distinguish the Holland Patent till from the Norway till except where the two are observed in superposition. Commonly, the Holland Patent till is more clayey and has fewer clasts than the Norway till. The limit of the Holland Patent till coincides with end moraines along the southern flank of the Deerfield Hills (Figure 11). In the West Canada Valley ice-marginal deltas were deposited at the limit of the Barneveld Readvance in a lake having an elevation of 290 to 305 meters. This lake may have been continuous with a high-level proglacial impoundment in the eastern Mohawk Valley which drained across a spillway at Delanson col.

"Indian Castle Drift"

Fullerton (1971) applied the term "Indian Castle Readvance" to an expansion of the Ontario Lobe during a period of free eastward drainage in the Mohawk Valley (Figure 12A). Deposits of "Indian Castle Drift" include the Norway and Holland Patent tills described in this report. Furthermore, detailed mapping has shown that the limit of the Indian Castle Readvance corresponds to two separate readvance limits, the Hinckley and Barneveld Readvance limits (Figure 12B). The Indian Castle Readvance corresponds to the Hinckley Readvance from the Hinckley Moraine System southeast to Fairfield. From Fairfield to Indian Castle the Indian Castle Readvance limit corresponds to the Barneveld Readvance limit. The name "Indian Castle" has been abandoned for these reasons (Muller and others, in review).



<u>Figure 11</u>. -- The inferred deployment of ice at the maximum extent of the Barneveld Readvance of the Oneida Sublobe. During the Barneveld Readvance upper Lake Gravesville probably remained connected to the main body of water, lower Lake Gravesville, which may have had a spillway at Delanson east of the area shown here. The Alder Creek Readvance of the Black River Sublobe, which ponded glacial Lake Forestport in the Black River Valley, was probably synchronous with the Barneveld Readvance. Adirondack ice was probably just beyond the northeast corner of the map. The exact position of the Mohawk Sublobe is unknown, although it remained far enough west to pond Lake Gravesville above 255 meters in the Mohawk Valley. (Base map is Figure 1) Blackened areas at the terminus of glacier represent either an end moraine or large kame deposit.

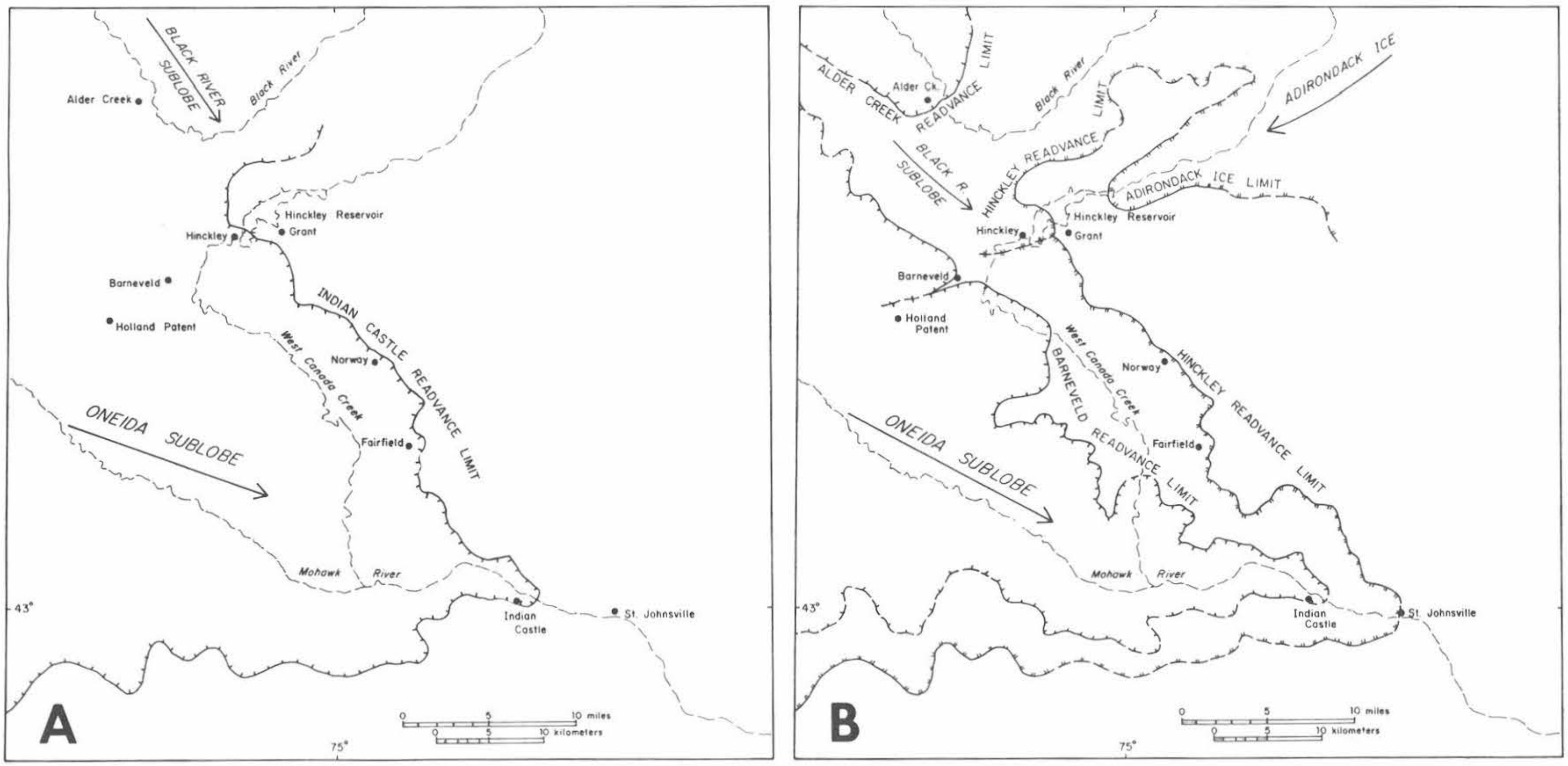


Figure 12. -- A. Maximum eastern extent of the "Indian Castle Readvance" of Fullerton (1971). B. Maximum positions of the Oneida and Black River Sublobes during the Hinckley and Barneveld Readvances and the inferred Alder Creek Readvance. The position of Adirondack ice during the Hinckley Readvance is inferred from ice-marginal deltas deposited in Lake Miller at the eastern end of the Hinckley Reservoir. (Base map for A and B is Figure 1)

Later Deposits

During later phases of deglaciation, readvances were restricted to the extreme western end of the Mohawk Valley and thus did not reach the West Canada Creek Valley. The valley did serve, however, as a major drainage outlet for proglacial impoundments in the upper Mohawk, Ninemile Creek and Black River valleys and from local Adirondack lakes and inwash drainage (Figure 13). At least three levels of valley train deposits, graded to falling lake levels in the Mohawk Valley and local knick points, were formed during ice recession. At Herkimer, West Canada Valley drainage built a delta with topset-foreset elevations of about 140 meters in a Mohawk Valley impoundment.

POINTS FOR DISCUSSION

Correlation to Eastern Great Lakes

Muller and others (in review) proposed correlations of the glacial events in the western Mohawk Valley to those in the Erie and Ontario basins (Dreimanis and Karrow, 1972; Dreimanis and Goldthwait, 1973). These correlations (Figure 2) were based on lithostratigraphic and morphostratigraphic information without supporting radiometric dates. The Erie-Ontario basin nomenclature appears to be the most appropriate in light of its close proximity to the West Canada Valley. This approach is favored over formulating new stratigraphic nomenclature because a record of Ontario Lobe activity is preserved in the West Canada Valley region and reasonable correlations may be formulated.

Two ties may exist between the stratigraphy of the West Canada Valley and the eastern Great Lakes. First, the unconformity and fluvial deposits of the Phase I-Phase II Interphase represent a major interstadial event. They record a period of erosion and fluvial deposition which requires recession of the Mohawk Sublobe to permit eastern fluvial drainage of the Mohawk Valley. The Phase I-Phase II Interphase is thought to represent the Erie Interstadial at which time the Great Lakes drained east across the divide at Rome to the Mohawk and Hudson Valleys (Mörner and Dreimanis, 1973). The unconformity and deposits representing this period further suggest an Erie Interstade equivalence by the distance of ice recession involved, and the fact that they are the only evidence of widespread erosion older than deposits equivalent to the Valley Heads Moraines.

The second tie to the Mohawk Valley stratigraphy is the group of outer moraines of the Valley Heads Moraine System (Muller, 1965) which appear to be correlative with the Norway and Holland Patent tills (formerly Indian Castle Drift of Fullerton, 1971; Muller and others, in review). Lateral tracing of the "Advanced Valley Heads

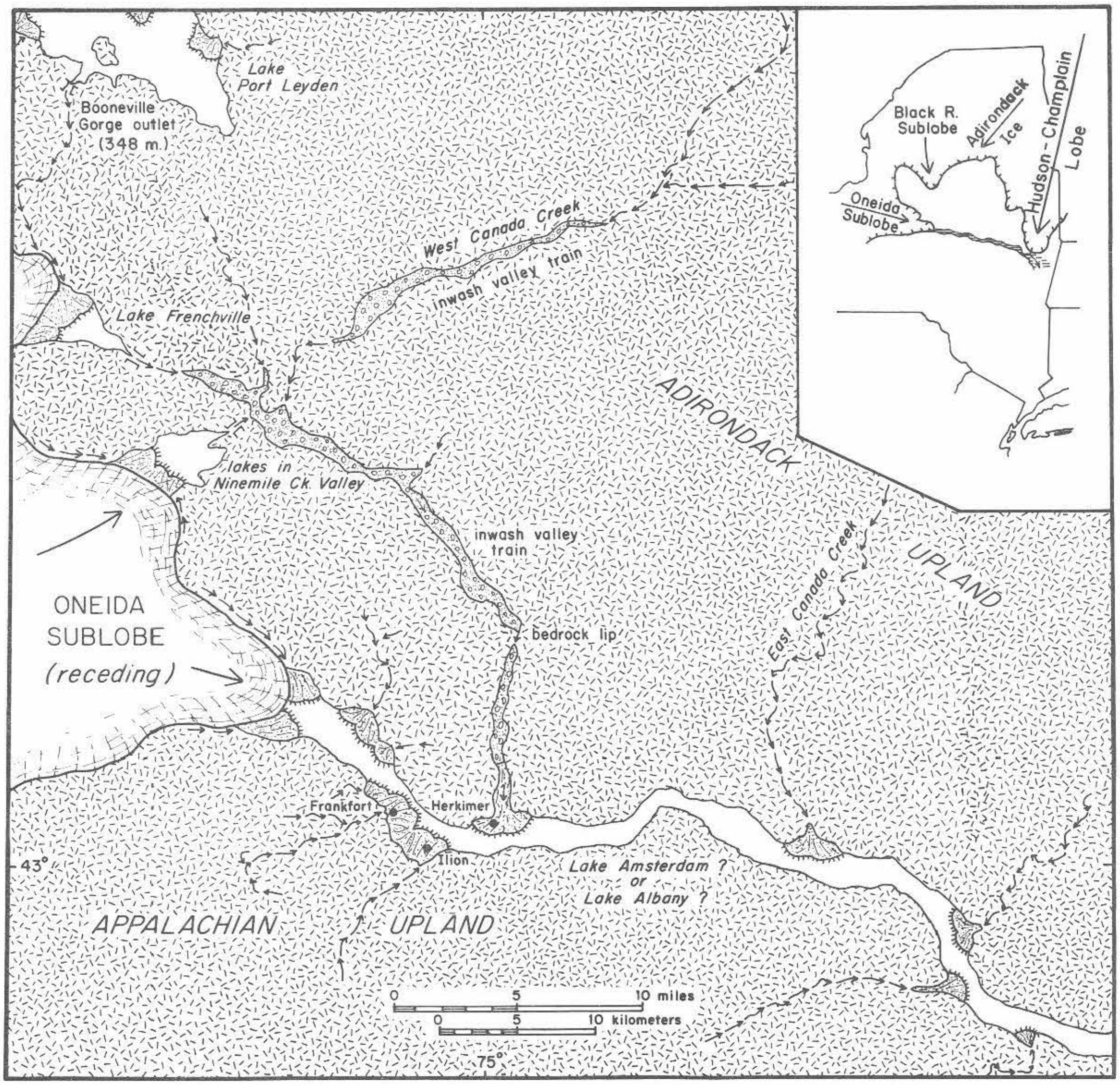


Figure 13. -- The inferred deployment of ice during the final recession of glacial sublobes from the West Canada Creek Valley. The Oneida Sublobe occupied the western end of the Mohawk Valley and ponded lakes in the upper Mohawk and Ninemile Creek valleys which drained into the West Canada Creek Valley. The Black River Sublobe ponded Lake Port Leyden in the Black River Valley which drained into the West Canada Valley via the Steuben Valley. The position of the Mohawk Sublobe is unknown but it may have ponded the Mohawk Valley to form Lake Amsterdam or receded from the eastern Mohawk Valley and allowed Lake Albany to inundate the valley. Adirondack ice receded from the upper West Canada Creek drainage basin but locally ponded glacial lakes in the Adirondacks may have drained into the West Canada Creek Valley. (Base map is Figure 1)

Moraine" of central New York (Muller, 1966) to western New York suggests that it is morphostratigraphically equivalent to the Valley Heads Moraine at Portageville in the Genesee Valley (Wyoming County) and the Lake Escarpment Moraines in the Erie Basin (Muller, 1977). The moraines pre-date the Gowanda Moraine, post-date the Kent Moraine and are probably Port Bruce in age. These correlations, would give a Port Bruce age to the Norway and Holland Patent tills. The Hawthorne till, deposited by ice of an eastern source (Mohawk Sublobe) is closely associated with the Norway till in age (about 200 years older) and for this reason would probably be a Port Bruce deposit.

Correlations to the East

The positions of till limits from the Mohawk Sublobe in the West Canada Valley suggest tentative correlations with ice margins, end moraines and possible readvance limits to the east. Tentative correlations for discussion are show in Table 1.

Parallel Response of the Ontario and Hudson-Champlain Lobes

The West Canada Valley stratigraphy allows the detailed study of readvance synchroneity between the Hudson-Champlain and Ontario Lobes during Phase II of deglaciation. The Black River and Oneida Sublobes of the Ontario Lobe appear to have responded to the same major climatic stimuli as the Mohawk Sublobe of the Hudson-Champlain Lobe. Interesting also is the failure of these sublobes to attain their maximum expansions simultaneously. Both the Hudson-Champlain and Ontario Lobes retreated during the Phase I - Phase II Interphase. Both lobes readvanced at the close of this interphase and reached their maximum extent in the West Canada Valley asynchronously. The Mohawk Sublobe reached its maximum extent at least 160 years prior to the full expansion of the Oneida Sublobe as determined by varve counts between the Hawthorne and Norway tills (Section 9 on Figure 3). Possible explanations may be different response times reflecting different distances from a source of outflow or differing contributions from Adirondack through flow or local Adirondack sources (Muller and others, in press).

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WEST CANADA VALLEY TILL SHEET MARGINS	WESTERN MOHAWK VALLEY	SCHOHARIE AND EASTERN MOHAWK VALLEYS
HAWTHORNE TILL LIMIT	SALISBURY MORAINE (0.8 km N of Salisbury, Salisbury Quad.)	MIDDLEBURG ICE MARGIN
	PINNACLE KAME MORAINE (Cushing, 1905)	
WEST CANADA TILL LIMIT	DIAMOND HILL DELTA Ice marginal delta from Hedgehog Mtn. to Diamond Hill (3.5 km NW or Salisbury, Salisbury Quad.)	WAGON WHEEL GAP MORAINE (Rich, 1935)
	CASSVILLE-COOPERSTOWN MORAINE (Krall, 1977)	

Table 1. -- Tentative correlations of till limits in the West Canada Valley and ice margins, end moraines and possible readvance limits in the Mohawk Valley region.

Scientific Research Society, the Senate Research Fund of Syracuse University, the Geology Department of SUNY at Geneseo, and the New York State Geological Survey.

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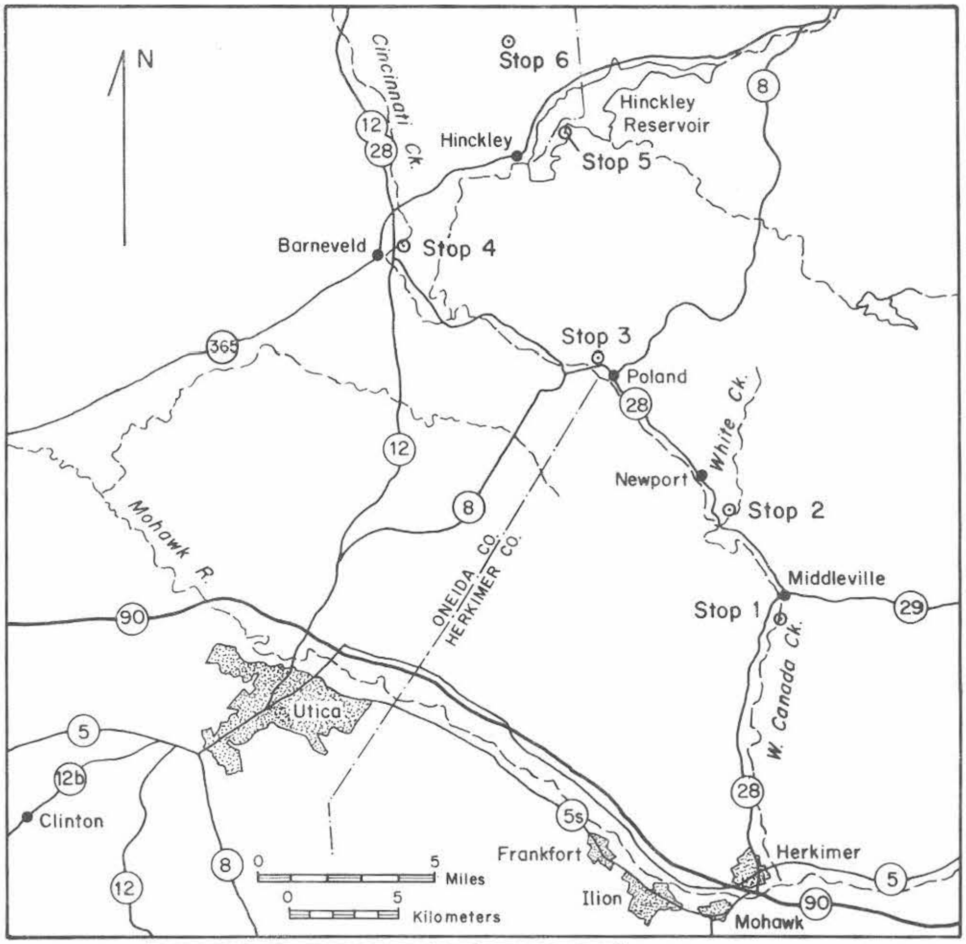
ROAD LOG

The field trip log starts at the intersection of Rt. 28 and Rt. 5 in downtown Herkimer where Rt. 28 heads north to Middleville (see Figure 14). To reach Herkimer head north from Clinton on Rt. 12B to Utica. In Utica, follow Rt. 5 to Herkimer or take the NY State Thruway - Rt. 90 (toll) east to the Herkimer exit. In Herkimer follow signs for Rt. 28.

Assemble at 8:30 A.M. in the parking lot of Chicago Market and Carls Drugs at the intersection of Rts. 5 and 28 in eastern downtown Herkimer

Miles	Route Description
0.0	From intersection of Rts. 5 and 28 in eastern downtown Herkimer (Herkimer quadrangle) head north on Rt. 28 toward Middleville and Newport.
3.0	Kast Bridge. The course of West Canada Creek has been staightened in this section of the valley. Note the flat topped bluff on east bank of West Canada Creek.
3.4	Bluffs on left side of highway (to west) expose Holland Patent and Norway tills. North Creek enters West Canada Creek from the east. West Canada till and striations on bedrock of the Trenton Group (N60W) are exposed along the last mile of North Creek before it enters West Canada Creek.
3.7	Bluffs on east bank of West Canada Creek expose units from Holland Patent till at the top through Norway and Hawthorne tills to West Canada till at the base. The bluffs are typical or many high banks along West Canada Creek between Kast Bridge and Middleville. These bluffs do not exist on the west bank of West Canada Creek because of the asymmetry of the bedrock profile of the valley. The apparent flat-topped surfaces of the bluffs are capped by till and lacustrine sediments. The flat benches may be the result of initial fluvial downcutting in the valley or the floors of lakes that existed during recession of the Barneveld Readvance.
7.1	Enter Middleville quadrangle.
7.3	view of STOP 1 from KOA campground. Bluff on east bank of West Canada Creek (Section 9 on Figure 3). Ahead on left are commercial Herkimer "diamond" (more appropriately



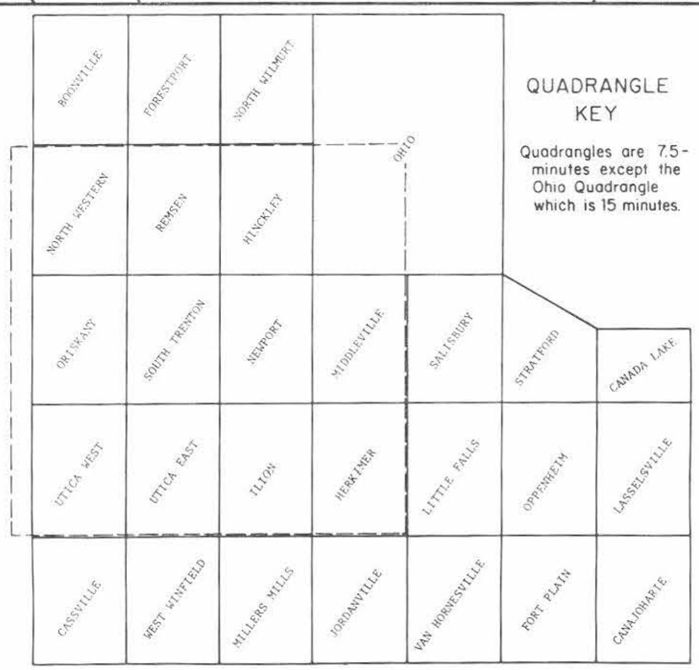


Figure 14. -- Highway map showing field trip stops. Key to USGS quadrangle maps covered by Figures 1 and 8, 9 and 11 thru 13. Dashed line shows area covered by highway map.

Middleville "diamond") prospecting grounds.

7.7 Exposures of Little Falls Dolostone on left while entering town of Middleville.

8.1 Turn right (east) staying on Rt. 28 in Middleville and proceed across bridge over West Canada Creek.

8.3 Traffic light at intersection of Rts. 28, 29 and 169 in Middleville. Turn right at light onto Rt. 169.

8.75 STOP 1. Park vehicles on shoulder of Rt. 169 next to entrance to pasture on right. Walk on dirt road to right to top of hill and across pastures for 0.5 miles to West Canada Creek bluff exposure.

Features and topics for discussion: (Section 9 on Figure 3)

1) Lithology and genesis of unit equivalent to the Norway till.

2) Lithologic and provenance changes within lacustrine sediment between the Norway till equivalent and the Hawthorne till.

3) Nature of contact of Hawthorne till and upper Newport beds.

4) Transition from upper Newport beds to West Canada till.

5) Compact and stony character of West Canada till.

6) Paleomagnetic samples were taken from the upper Newport beds and from three levels within the lacustrine sediments between the Norway till equivalent and the Hawthorne till.

Return to vehicles and head back to Rt. 28 via Rt. 169.

9.2 Junction of Rts. 28, 29 and 169. Continue north (straight) through Middleville on Rt. 28.

9.8 Terrace along West Canada Creek. Many terraces in the valley are graded to lake levels in the lower West Canada and Mohawk valleys (Figure 13). They contain alluvial fan sediment and slope away from large gullys cut in the glacial deposits along the valley sides.

10.2 Right side or Rt. 28 and continuing for the next 1.5 miles are outcrops of Little Falls Dolostone. Behind the exposure on the right is a large quarry noted for its Herkimer "diamonds".

11.6 Junction of Rt. 28 and White Creek Rd. on right. On left is West Canada Valley Central School. Turn right onto White Creek Rd. which parallels White Creek.

12.0 STOP 2. Park vehicles and walk across pasture on west side of White Creek Road.

Features and topics for discussion: (Section 5 on Figure 3)

1) Upper Newport beds appear at top of bluffs in soil profile.

2) Clayey and silty character of West Canada till.

3) Deformation of lower Newport beds beneath West Canada till. Nappe-like folds and thrusts with northwest directed displacement.

4) Transition from lower Newport beds to White Creek till and associated sediment flows.

5) Topography on the upper surface of the White Creek till (morainic or drumlinoid features or neither?).

6) Very stony character of the White Creek till which contains a high proportion of metamorphic clasts.

7) Subglacial grooving on the surface of the White Creek till has been exposed here in the past.

Return to vehicles and head back to Rt. 28 via White Creek Rd.

12.4 Junction of White Creek Road and Rt. 28. In view to the west-southwest from this intersection is a large bluff on the west bank of West Canada Creek (Section 6 on Figure 3). This section contains all the units from the lower Newport beds to the Norway till on top. Turn right (west) onto Rt. 28 toward Newport.

12.45 Enter Newport quadrangle.

12.7 Cross White Creek.

12.9 Right side of highway. Bluff exposure of Hawthorne till overlying upper Newport beds.

13.5 Right side of highway. Old meander scar of West Canada Creek. Across pasture are bluff and gully exposures (Sections 3 and 4 on Figure 3) which contain all units from basal White Creek till up thru Norway till.

13.8 Enter town of Newport and continue northwest on Rt. 28.

15.0 Leave Newport and head west toward Poland on Rt. 28.

16.3 Right side of highway. Lacustrine sands which lie stratigraphically below the Norway till and may predate or be time-equivalent to the Hawthorne till and Salisbury Readvance.

17.2 Fluvial terrace on right is inset in an older alluvial fan which slopes toward the center of the valley from large gullys to the northeast.

17.4 Enter Poland.

18.1 Center of Poland. Junction of Rts. 28 and 8. Stay on Rt. 28 to the northwest.

18.6 Turn off Rt. 28 onto dirt road leading into gravel pits.

STOP 3. Poland sand and clay pits.

Two exposures:

1) Pit exposing outwash sands. Fluvial sand infilling early incision of the valley cut during fall in lake levels proceeding the Barneveld Readvance.

2) Clay pit exposure of Phase II sediments.

Section: (similar to Sections 1 and 2 on Figure 3 and northwest end of West Canada Valley profile on Figure 7)

(top)

6 ft. oxidized varves (soil profile)

1 ft. sediment flows

6 ft.	clayey dark gray till with high proportions of sandstone and Trenton Group limestone clasts (Norway till)
1 ft.	deformed lacustrine beds
4 ft.	thin varve couplets (less than 2 inches) containing a pink and tan clay component and red rain-out sediment (Oneida Sublobe provenance); flutes indicate paleocurrent flow to the southeast.
12 ft.	thick varve couplets (greater than 2 inches) with dark gray and drab colors and black rain-out sediment near base (Mohawk Sublobe provenance); couplets thicken downward; rib and furrow structures indicate paleocurrent flow to the northwest.
2 ft.	lacustrine sand and silt, oxidized at top and contain calcareous cement.
25 ft.	rippled lacustrine sands with calcareous cemented layers and lenses; paleocurrent flow to the east.

(limit of exposure)

Note especially change from Mohawk to Oneida Sublobe provenance within the varve section accompanied by a change in paleocurrent directions from northwest to southeast.

Return to Rt. 28 and head north.

18.8	Cross West Canada Creek.
19.0	Flat-topped hill to right with gravel pit excavations is capped by an ice-marginal delta at the maximum position of the Barneveld Readvance.
19.5	Intersection of Rts. 8 and 28. Stay on Rt. 28 to the right. Cross bridge over West Canada Creek.
21.6	Cross Mill Creek on Rt. 28.
21.8	Enter South Trenton quadrangle.
22.4	Cross West Canada Creek.
22.6	Enter Remsen quadrangle.

23.3 Cross Cincinnati Creek.

24.5 Main road bears to the left. Continue on Rt. 28 to the left.

25.35 Intersection of Routes 12 and 28. Proceed north on Rt. 12 toward Barneveld.

25.7 Railroad underpass.

25.8 Turn left off Rt. 12, follow signs to Barneveld.

26.5 Intersection with Rt. 365 in Barneveld. Proceed east (right turn).

26.9 Rt. 12 underpass.

27.4 STOP 4. Park vehicles at side of road. Cincinnati Creek sections (Figure 5).

Return to vehicles and continue northeast from STOP 4.

28.2 Intersection with Rt. 365. Proceed northeast on Rt. 365 (right turn) through railroad underpass.

30.9 The highway crosses an incised portion of a delta deposited during recession from the Hinckley Readvance.

31.3 Enter Hinckley quadrangle.

31.6 Hills on both sides of the highway are a part of the Hinckley Moraine System.

31.9 Enter village of Hinckley. Turn right (south) onto bridge crossing West Canada Creek.

32.3 Main road bears left.

32.7 Hinckley Dam

35.0 Main road bears right. Leave main road proceeding north (straight ahead) and follow signs to Trail's End Park. DO NOT ENTER PARK. Exposure on the east side of the road contains calcareous western provenance gravel overlying older non-calcareous gravel of Adirondack provenance.

35.1 STOP 5. Dirt driveway leading to Trail's End bluff exposure on Hinckley Reservoir (Figure 10). NOTE: This is private property.

Follow route of field trip back to Hinckley.

38.2 Intersection with Rt. 365 at Hinckley. Follow one of two options.

Option 1. Follow Rt. 365 west (turn left) to Rt. 12 south and terminate field trip. Distance to Rt. 12 from Hinckley is 6.1 miles.

Option 2. Follow Rt. 365 east (turn right) and proceed to STOP 6.

38.5 Hinckley Dam

40.8 Village of Ninety Six Corners. Turn left off Rt. 365 and proceed northwest toward Bailey Hills.

42.4 STOP 6. Bailey Hills delta. Follow dirt road into gravel pit.

Return to vehicles and continue west.

42.5 Enter Remsen quadrangle. On both sides of the road is hummocky topography resulting from stagnation of the Black River Sublobe as it thinned over the Black River-West Canada Creek divide during recession from the Hinckley Readvance. The drift is composed of both stratified deposits and ablation till.

44.6 Intersection near Fairchild Cemetary. Continue west.

46.0 Road ends at "T" intersection. Turn left and proceed south into Remsen.

47.1 Intersection in village of Remsen. Turn right and proceed west to Rt. 12.

47.5 Junction with Rt. 12. Proceed south and follow signs to Utica and Clinton.

Structure and Rock Fabric Within the Central and Southern Adirondacks

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INTRODUCTION

The area referred to as the southern Adirondacks is shown in Figure 1. Within this region, the Precambrian is bounded approximately by the towns of Lowville and Little Falls on the west and Saratoga Springs and Glens Falls on the east (Fig. 2).

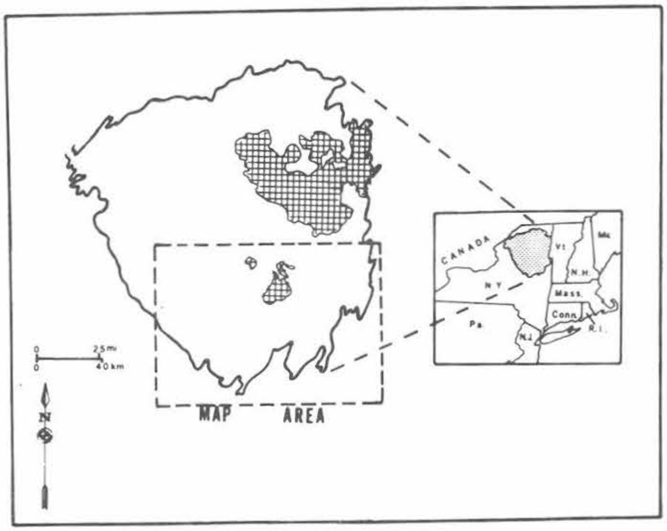

Fig. 1. Location map of the Adirondack Mts. Major anorthosite massifs are represented by the grid pattern. The central and southern Adirondacks lie within the dashed rectangle labeled "Map Area".

Mapping in the southern Adirondacks was done first by Miller (1911, 1916, 1920, 1923), Cushing and Ruedemann (1914), Krieger (1937), and Cannon (1937); more recent investigations were undertaken by Bartholome (1956), Thompson (1959), Nelson (1968), and Lettney (1969). At approximately the same time Walton (1961) began extensive field studies in the eastern portion of the area (Paradox Lake, etc.), de Waard (1962) began his studies in the west (Little Moose Mt. syncline). Subsequently de Waard was joined by Romey (de Waard and Romey, 1969).

Separately and together, Walton and de Waard (1963) demonstrated that the Adirondacks are made up of polydeformational structures, the earliest of which consist of isoclinal, recumbent folds. Their elucidation of Adirondack geology set the tone for future workers in the area. In this regard one of their most important contributions to the regional picture was that the lithologic sequence of the west-central Adirondacks is similar to that of the eastern Adirondacks.

Beginning in 1967 McLelland (1969, 1972) initiated mapping in the southernmost Adirondacks just to the west of Sacandaga Reservoir subsequently this work was extended north and east to connect with that of Walton and de Waard.

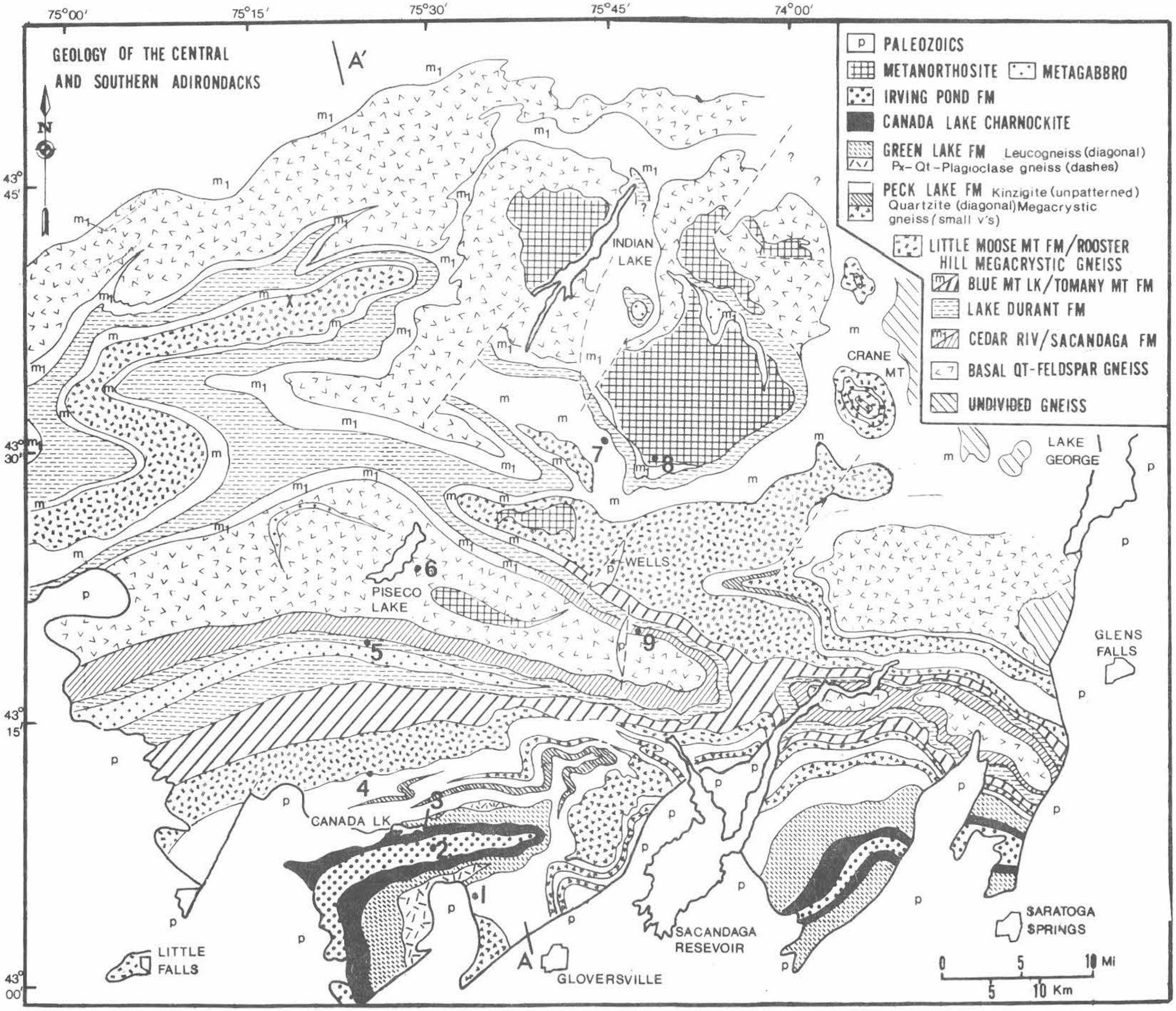

Fig. 2. Formational map and stop localities for the central and southern Adirondacks (from McLelland and Isachsen, 1980)

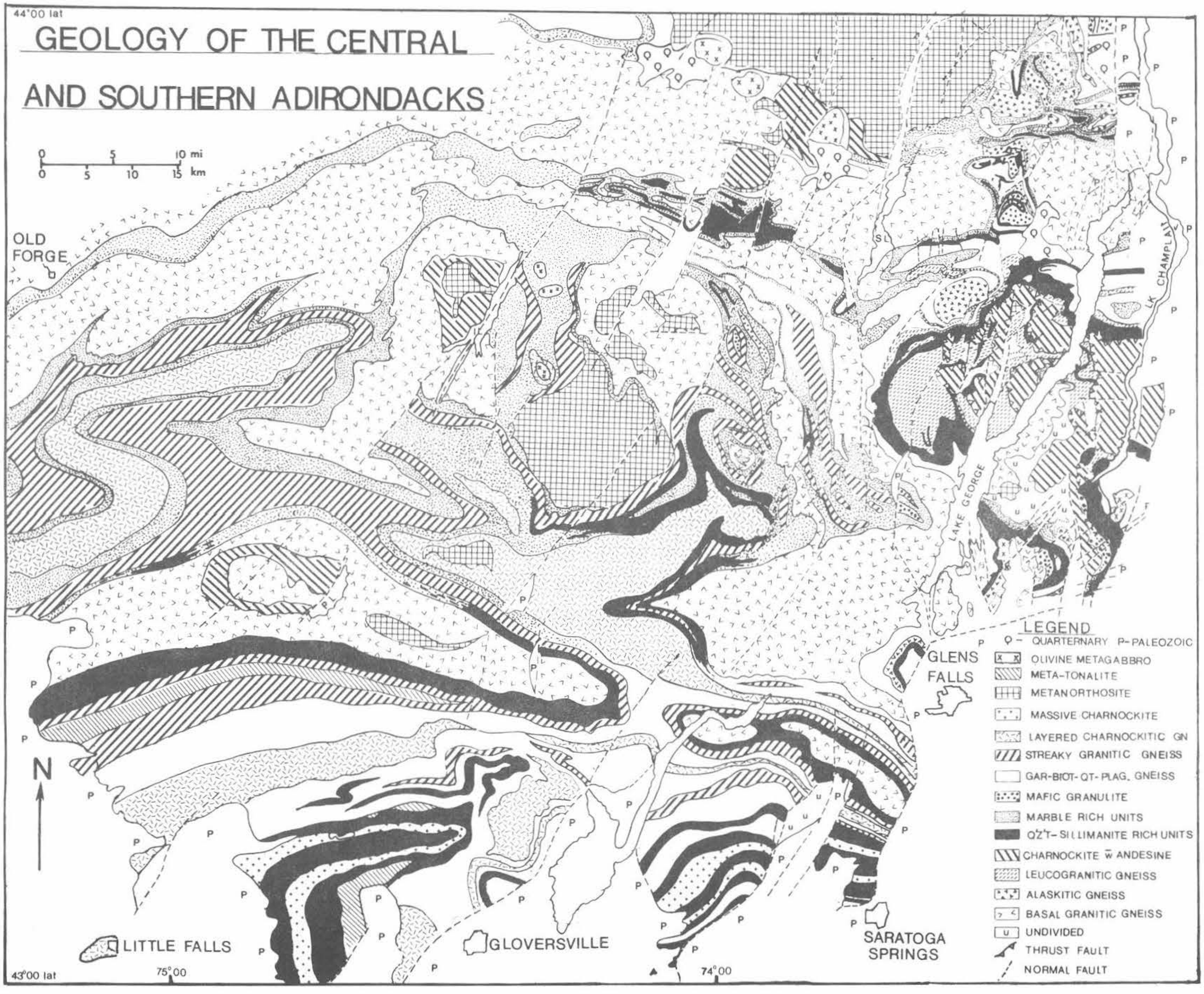

Fig. 3. Lithological map of the central and southern Adirondacks. Note that only a few high angle faults are shown.

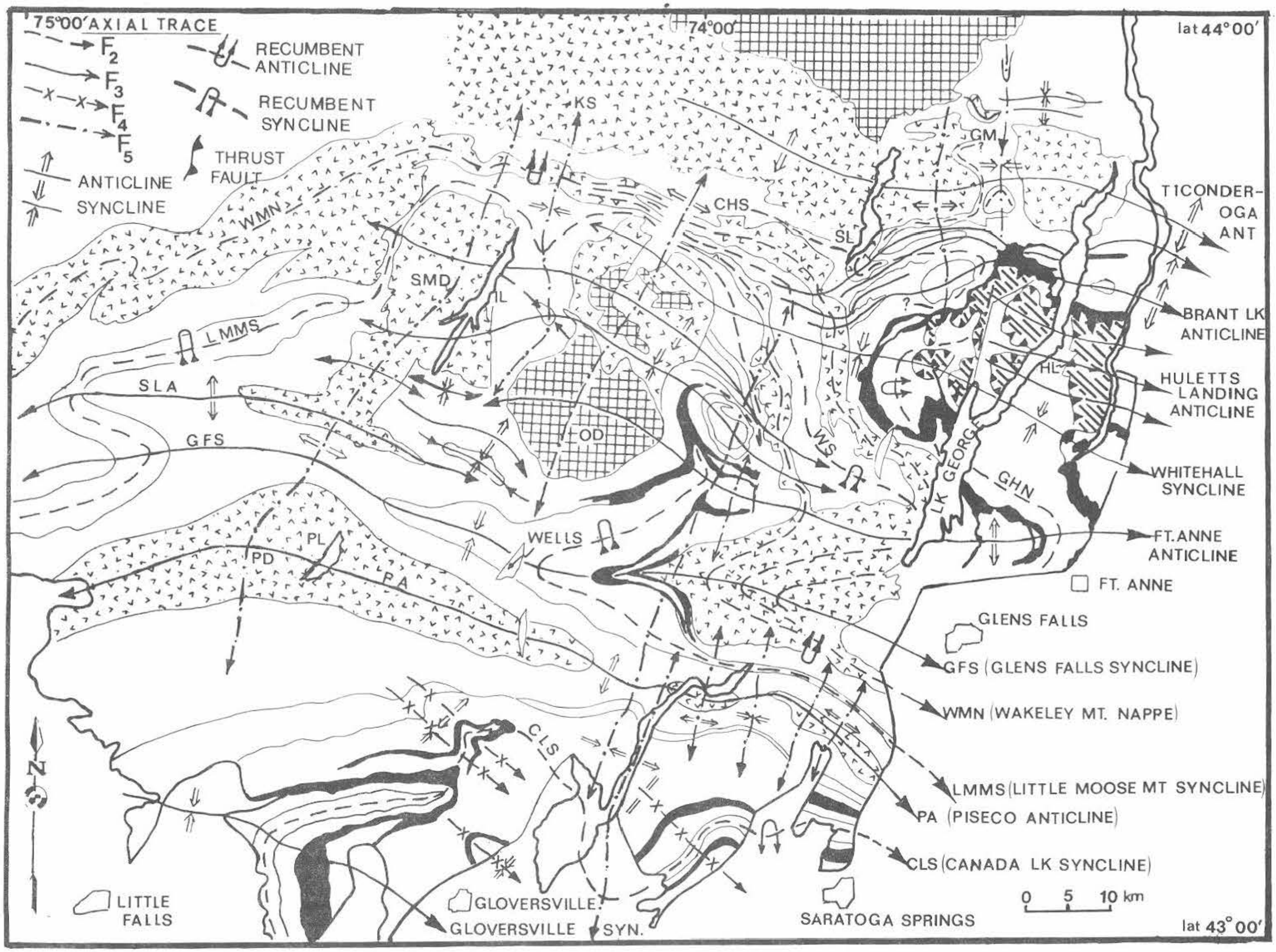

Fig. 4. Axial trace map of the central and southern Adirondacks. For lithological map symbols see Fig. 3. Abbreviations not on map: CHS - Canada Hill syncline; GHN - Green Hill nappe; GM - Glidden Marsh syncline; HL - Hullett's Landing; OD - Oregon Dome; PD - Piseco Dome; PL - Piseco Lake; SL - Schroon Lake; SLA - Spruce Lake anticline; SMD - Snowy Mt. Dome; WS - Warrensburg syncline

Geraghty (1973) and Farrar (1976) undertook detailed mapping in the eastern half of the North Creek 15' quadrangle, and tied into investigations in the Brandt Lake region by Turner (1971). Recently, Geraghty (1978) completed a detailed study of the structure and petrology in the Blue Mt. Lake area. Current investigations by McLelland and by the N.Y. Geological Survey are going forward in the general region surrounding Lake George.

The foregoing investigations have increased our knowledge of the southern Adirondacks, and this fieldtrip is designed to show as many examples of the region's structure, lithology, and petrology as time permits.

STRUCTURAL FRAMEWORK OF THE SOUTHERN ADIRONDACKS

The southern Adirondacks (Figs. 2-5) are underlain by multiply deformed rocks which have been metamorphosed to the granulite facies. The structural framework of the region consists of four unusually large fold sets, F_2 - F_5 together with an early set of isoclines represented solely by intrafolial minor folds with associated axial planar foliation (Figs. 2-4). Relative ages have been assigned to these fold sets, but no information exists concerning actual time intervals involved in any phase of the deformation. It is possible that several, or all of the fold sets, are manifestations of a single deformational continuum.

The earliest and largest of the map-scale folds are recumbent, isoclinal structures (F_2) -- for example the Little Moose Mt. syncline (de Waard, 1962) and Canada Lake nappe (McLelland, 1969) (Figs. 2 and 5). These isoclines have axes that trend approximately E-W and plunge within 20° of the horizontal. As seen in Figures 4 and 5 the axial traces of each of the F_2 folds exceeds 100 km. They are believed to extend across the entire southern Adirondacks. Subsequent useage of the terms "anticline" and "syncline," rather than "antiform" and "synform," is based on correlations with rocks in the Little Moose Mt. syncline where the stratigraphic sequence is thought to be known (de Waard, 1962).

Close examination reveals that the F_2 folds rotate an earlier foliation defined principally by platets of quartz and feldspar and axial planar to minor intrafolial isoclines. Although this foliation is suggestive of pre-F_2 folding, such an event does not seem to be reflected in the regional map patterns (Fig. 3). However, it is possible that major pre-F_1 folds exist but are of dimensions exceeding the area bounded by Figure 3. If this is the situation, their presence may be revealed by continued mapping. The existence of such folds is suggested by the work of Geraghty (1978) in the Blue Mt. area. In the vicinity of Stark Hills charnockites of the Little Moose Mt. Fm. appear to be identical to supposedly older quartzo-feldspathic gneisses (basal) which lie at the base of the lithologic sequence. Given this situation, then the Cedar River and Blue Mt. Lake Fms. are identical, and there emerges a pre-F_2 fold cored by the Lake Durant Formation. However, careful examination of the Lake Durant Formation has failed to reveal the internal symmetry implied by this pre-F_2 fold model. It is possible, of course, that the pre-F_1 foliation may not be related directly to folding (e.g. formed in response to thrusting, gravity sliding, etc.; Mattauer, 1975). Currently the origin of the pre-F_2 foliation remains unresolved. In most outcrops the pre-F_2 foliation cannot be distinguished from that associated with the F_2 folding.

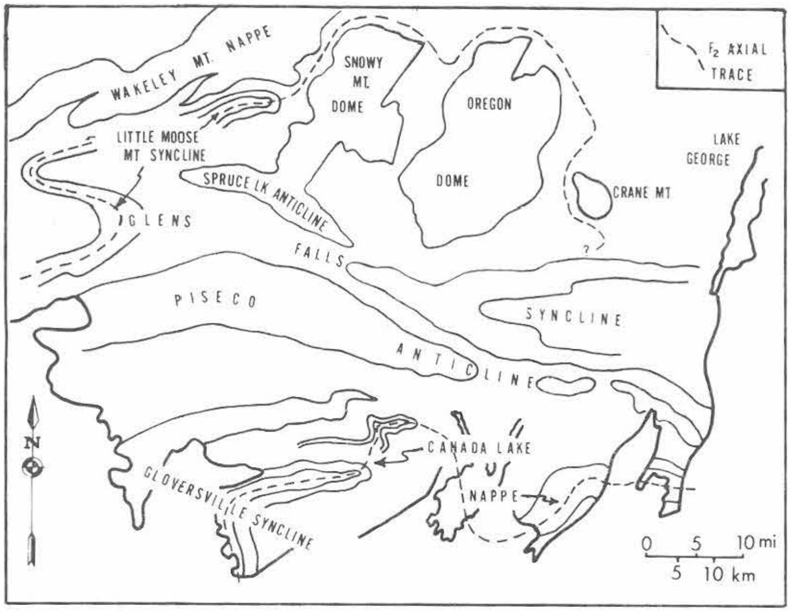

Fig. 5. Blocked out major folds of the central and southern Adirondacks (from McLelland and Isachsen, 1980).

Following the F_2 folding, there developed a relatively open and approximately upright set of F_3 folds (Figs. 2-5). These are coaxial with F_2. In general the F_3 folds are overturned slightly to the north, the exception being the Gloversville syncline with an axial plane that dips 45°N. The F_3 folds have axial traces comparable in length with those of the F_2 set. The Piseco anticline and Glens Falls syncline can be followed along their axial traces for distances exceeding 100 km until they disappear to the east and west beneath Paleozoic cover. The similarity in size and orientation of F_2 and F_3 suggests that both fold sets formed in response to the same force and field.

The fourth fold set (F_4) is open, upright, and trends NW. Within the area these folds are less prevalent than the earlier sets. However, Foose and Carl (1977) have shown that within the NW Adirondacks, northwest-trending folds are widespread and play an important role in the development of basin and dome patterns.

The fourth regional fold set (F_5) consists of large, upright NNE folds having plunges which differ depending upon the orientation of earlier fold surfaces. The F_5 folds are observed to tighten as one proceeds towards the northeast.

The regional outcrop pattern is distinctive because of the interference between members of these four fold sets (Figs. 2-5). For example, the "bent-finger" pattern of the Canada Lake nappe west of Sacandaga Reservoir is due to the superposition of the F_3 Gloversville syncline on the F_2 fold geometry (Fig. 4). East of the reservoir the reemergence of the core rocks of the Canada Lake nappe is due to the superposition on F_2 of a large F_5 anticline whose axis passes along the east arm of the reservoir (Fig. 4). The culmination-depression pattern along the Piseco anticline results from the superposition of F_3 and F_5 folds. The structure of the Piseco dome is due to the intersection of the Piseco anticline (F_3) with the Snowy Mt. anticline (F_5). Farther to the north, Crane Mt. is a classic example of a structural basin formed by the interference of F_3 and F_5 synclines (Figs. 2 and 6).

DISCUSSION AND SYNTHESIS OF STRUCTURAL RELATIONSHIPS

Over a decade ago Walton and de Waard (1963) proposed that rocks of the anorthosite-charnockite suite comprise a pre-Grenvillian basement on which a coherent "supracrustal" sequence was deposited unconformably. Rocks which would be assigned a basement status in this model are designated as basal quartzo-feldspathic gneiss in Figure 3. The basal Cedar River Fm. of the overlying "supracrustal" sequence consists of marbles, quartzites, garnet-sillimanitic gneisses, and various calc-silicates. This lowermost unit is followed upward by various quartzo-feldspathic gneisses, marbles, and other metasedimentary sequences shown in Figure 2. Although our own research agrees with the generalized lithologic sequences of de Waard and Walton, two major provisos are necessary and are given here.

(1) Anorthositic rocks intrude the so-called supracrustal sequence, and therefore the anorthosites post-date these units and cannot be part of an older basement complex (Isachsen, McLelland, and Whitney, 1976; Husch, Kleinspehn, and McLelland, 1976). Isotopic evidence (Valley and O'Neill's (1983); Ashwal and Wooden, 1984) suggests that the anorthosites intruded prior to the 1.1 Ma Grenvillian metamorphism probably during a non-compressional stage (Emslie, 1978, Whitney, 1983). Angular, rotated xenoliths within the anorthosites exhibit pre-intrusion foliation and imply an earlier orogenic event(s).

(2) Within the metastratified units of the region, there exists field evidence for primary facies changes. For example, the well-layered sillimanite-garnet-quartz-feldspar gneisses of the Sacandaga Formation grade laterally into marble-rich units of the Cedar River Fm. exposed north of the Piseco anticline (Figs. 2,3). This transition along strike can be observed just south of the town of Wells, and its recognition is critical to the interpretation of the regional structure. Thus the great thickness of kinzigites (granulite-facies metapelites) south of the Piseco anticline gives way to the north to thinner units marked by marbles, calcsilicates, and quartzites. We interpret this lithologic change as due to a transition from a locally deep basin in which pelitic rocks were accumulating to a shallow-water shelf sequence dominated by carbonates and quartz sands.

Given the foregoing information, it has been possible to map and correlate structures and lithologies on either side of the Piseco anticline. In the northwest the sequence on the northern flank proceeds without structural discontinuity into the core of the Little Moose Mt. syncline. There occurs on the southern flank a mirror image of the northwestern lithologic sequence as units are traced towards the core of the Canada Lake nappe. It follows that the Canada Lake nappe and Little Moose Mt. syncline are parts of the same fold (Fig. 6). The amplitude of this fold exceeds 70 km. and it can be followed for at least 150 km along its axial trace. The major F_2 and F_3 folds of the area are exposed through distances of similar magnitude, but their amplitudes are less than those of the F_1 isoclines. The structural framework that emerges is one dominated by exceptionally large folds.

Accepting that the Little Moose Mt. syncline and Canada Lake nappe are the same fold, and noting that the fold axis is not horizontal, it follows that the axial trace of the fold must close in space. The axial trace of the Canada Lake nappe portion of the structure can be followed from west of Gloversville to Saratoga Springs. Therefore, the axial trace of the Little Moose Mt. syncline also must traverse the Adirondacks to the north. Mapping strongly suggests that the hinge lines of this fold passes through North Creek and south of Crane Mt. (Fig. 6). From here the axial trace swings westward along the north limb of the Glens Falls syncline to a point north of Wells and thence eastward to a point south of Glens Falls. This model is depicted schematically in Figure 6 where the southern Adirondacks are shown as underlain largely by the Canada Lake-Little Moose Mt. syncline. Later folding by F_3 and F_5 events has resulted in regional doming of the F_2 axial surface and erosion has provided a window through the core of this dome. Note the western extension of the Piseco anticline beneath the Paleozoic cover. This extension is consistent with aeromagnetics of the area.

Currently attempts are underway to synthesize the structural framework of the entire Adirondacks by extending the elements of the present model to other areas. A preliminary version is shown in Figure 7 and suggests that most Adirondack structure is explicable in terms of the four regional fold sets described here. Thrust faulting has been recognized in the eastern Adirondacks (Berry, 1961) and high strain zones exist in many other areas of the Adirondacks (McLelland, 1984). Associated with these are distinctive ribbon gneisses (Fig. 8) and sheath folds (Fig. 9). These are further discussed in Stop 6, Road Log.

CONCLUDING SPECULATIONS

The ultimate origin of the structural and petrologic features of the Adirondacks remains obscure. A possible clue to the mechanisms involved is Katz's (1955) determination of 36 km as the present depth to the M-discontinuity beneath the Adirondacks. Because geothermometry-geobarometry place the peak of the Grenville metamorphism at 8-9 kb (24-36 km), a double continental thickness is suggested. Such thicknesses presently exist in two types of sites, both plate-tectonic related. The first is beneath the Andes and seems related to magmatic underplating of the South American plate (James, 1971). The second is beneath the Himalayas and Tibet and is due to thickening in response to collision (Dewey and Burke, 1973) or continental underthrusting (Powell and Conaghan, 1973). The presence of ribbon lineation, sheath folds, and subhorizontal mylonitic foliation within the region strongly suggests regional rotational strain with a dominant component of simple shear (McLelland, 1984). Rotated K-feldspar augen exhibit tails asymmetric to foliation suggesting an east side up and to the west sense of tectonic transport.

Southeastward directed subduction would be consistent with this model. The relevant plate margin presumably lies buried beneath the present day Appalachians.

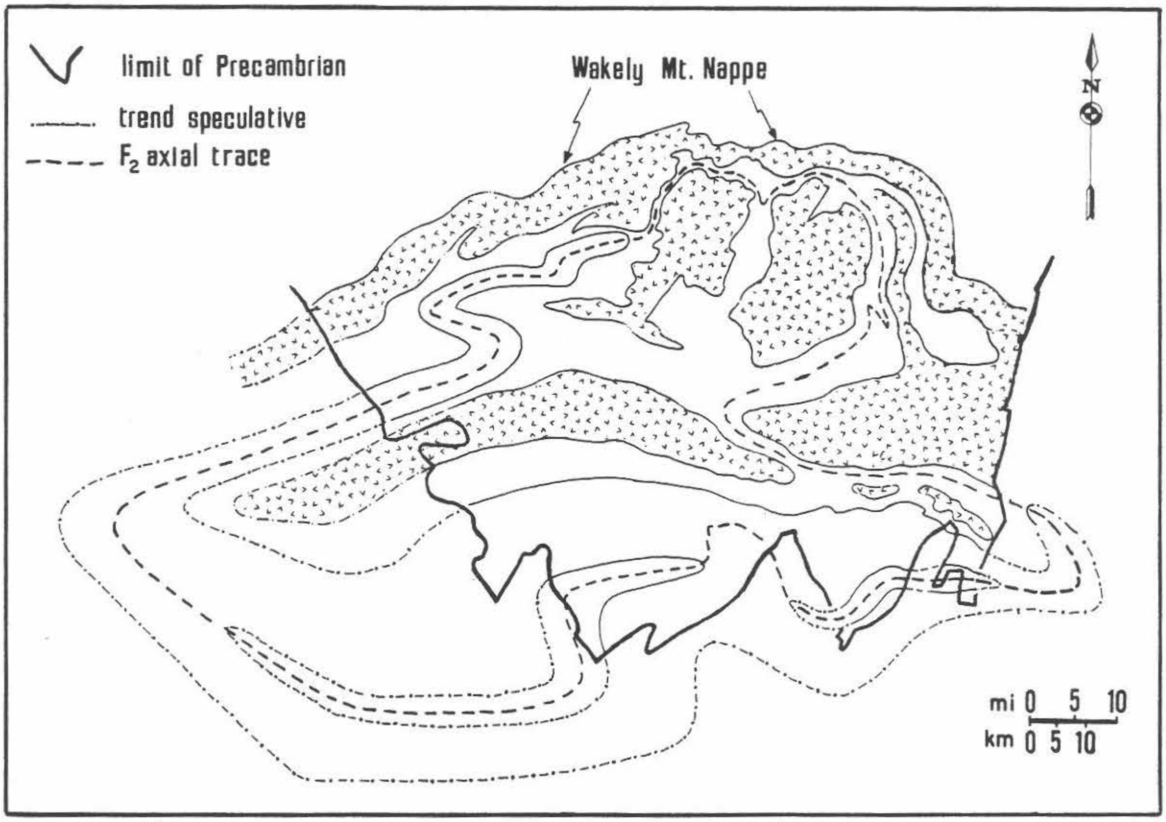

Fig. 6. Geologic sketch map showing the proposed axial trace of the F_2 Little Moose Mt. - Canada Lake syncline. The western extension of the Piseco anticline is inferred from aeromagnetic data (from McLelland Isachsen, 1980).

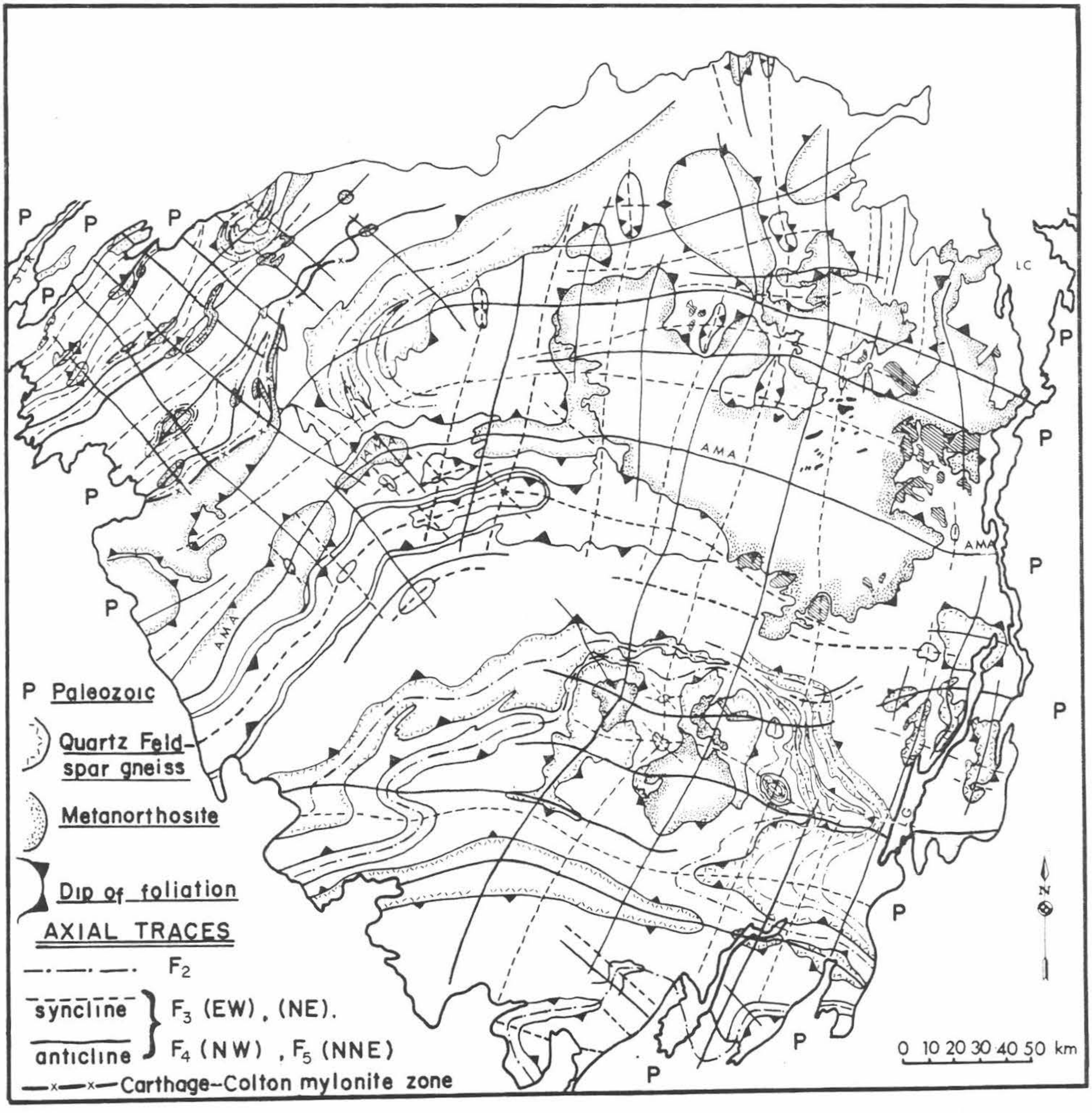

Fig. 7. Axial trace map of the Adirondack Mts. (from McLelland and Isachsen, 1980).

ROAD LOG

Mileage

0 Junction of Willie Road, Peck Hill Road, and NY Rt. 29A

1.3 Mud Lake to northeast of NY Rt. 29A

2.8 Peck Lake to northeast of NY Rt. 29A

3.6 **Stop #1**. Peck Lake Fm.

This exposure along Rt. 29A just north of Peck Lake is the type locality of the sillimanite-garnet-biotite-quartz-feldspar gneisses (kinzigites) of the Peck Lake Fm. In addition, there are exposed excellent minor folds of several generations. Note that the F_2 folds rotate an earlier foliation.

The white quartzo-feldspathic layers in the kinzigites consist of quartz, two feldspars, and garnet and are believed to be anatectic. Note that fish-hook terminations on some of these suggest that they have been transposed. It is also clear that these anatectites have been folded by F_2 indicating a pre-F_2 metamorphic event(s). In a similar fashion some garnets in the rock appear to be flattened while others do not. It is believed that the anatectites formed at the muscovite-quartz reaction and are essentially in situ melts. Further anatexis did not take place due to absence of vapor.

6.1 Junction NY Rt. 29A and NY Rt. 10

8.0 Nick Stoner's Inn on west side of NY Rt. 29A-10

8.6 **Stop #2.** Irving Pond Fm., .5 mile north of Nick Stoner's Inn, Canada Lake.

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

At the southern end of the cut typical, massive quartzites of the Irving Pond are seen. Proceeding north the quartzites become "dirtier" until they are essentially quartzose sillimanite-garnet-biotite-feldspar gneisses (kinzigites).

At the northern end of the cut, and approximately on the Irving Pond/Canada Lake Fm. contact there occurs an excellent set of F_2 minor folds. Polished slabs and thin sections demonstrate that these fold an earlier foliation defined by biotite flakes and flattened quartz grains.

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

Recently Eckelmann (pers. comm.) has studied zircon population morphologies in the Rooster Hill and similar lithologies. His results strongly suggest an igneous plutonic origin. This would be consistent with the igneous origin assigned the Hermon granite of the northwest Adirondacks - a rock that is markedly similar to the Rooster Hill.

Mileage	
20.0	Low roadcut in kinzigites of Tomany Mt. Fm.
21.4	Avery's Hotel on west side of NY Rt. 10
22.5	Long roadcuts of quartzofeldspathic gneisses and metasediments of Lake Durant Fm. intruded by metagabbro and anorthositic metagabbro.
23.6	Roadcut of anorthositic metagabbro and metanorite.
23.9	Roadcut on west side of highway shows excellent examples of anorthositic gabbros intrusive into layered pink and light green quartzo-feldspathic gneisses. The presence of pegmatites and cross-cutting granitic veins is attributed to anatexis of the quartzo-feldspathic gneisses by the anorthositic rocks.
24.0	**Stop #5**. Lake Durant and Sacandaga Fms. intruded by anorthositic gabbros and gabbroic anorthosites.

These roadcuts are located on Rt. NY 10 just south of Shaker Place.

The northernmost roadcut consists of a variety of metasedimentary rocks. These lie directly above the Piseco anticline and are believed to be stratigraphically equivalent to the Sacandaga Formation. The outcrop displays at least two phases of folding and their related fabric elements. These are believed to be F_2 and F_3. A pre-F_2 foliation is thought to be present. Both axial plane foliations are well developed here. Several examples of folded F_2 closures are present and F_2 foliations (parallel to layering) can be seen being folded about upright F_3 axial planes.

Farther to the south, and overlooking a bend in the west branch of the Sacandaga River, there occurs a long roadcut consisting principally of pink and light green quartz-perthite gneiss belonging to the Lake Durant Fm. About half-way down this roadcut there occurs a large boudin of actinolitic and diopsidic gneiss. To the north of the boudin the quartzo-feldspathic gneisses are intruded pervasively by anorthositic gabbros, gabbroic anorthosites, and various other related igneous varieties. At the north end of the cut and prior to the metastratified sequences these intrusives can be seen folded by upright fold axes. They are crosscut by quartzo-feldspathic material.

Within this general region the Lake Durant Fm. and other quartzo-feldspathic gneisses seem to have undergone substantial anatexis. This is suggested by the "nebular" aspect of the rocks. Good examples of this are seen in the manner in which green and pink portions of the quartzo-feldspathic gneisses mix. Note also the clearly cross-cutting relationships between quartzofeldspathic gneiss and mafic layers at the south end of the roadcut. Here it seems that mobilized Lake Durant is cross-cutting its own internal stratigraphy. Also note that the quantity of pegmatitic material is greater than usual. This increase in anatectic phenomena correlates closely with the appearance of extensive metagabbroic and metanorthositic rocks in this area. It is believed that these provided a substantial portion of the heat that resulted in partial fusion of the quartzo-feldspathic country rock.

Mileage

31.0 Red-stained basal quartzofeldspathic gneisses that have been faulted along NNE fractures.

31.5 Junction NY Rt. 10 and NY Rt. 8. End Rt. 10. Turn east on NY Rt. 8.

33.0 **Stop #6.** Core rocks of the Piseco anticline.

Hinge line of Piseco anticline near domical culmination at Piseco Lake. The rocks here are typical basal quartzo-feldspathic gneisses such as occur in the Piseco anticline and in other large anticlinal structures, for example Snowy Mt. dome, Oregon dome.

The pink "granitic" gneisses of the Piseco anticline do not exhibit marked lithologic variation. Locally grain size is variable and in places megacrysts seem to have been grain size reduced and only a few small remnants of cores are seen. The open folds at this locality are minor folds of the F_3 event. Their axes trend N70W and plunge 10-15° SE parallel to the Piseco anticline. Co-axial F_2 isoclines are also present.

The most striking aspect of the gneisses in the Piseco anticline is their well-developed lineation. This is expressed by rodding and ribbon fabrics. These may consist of alternating ribbons of quartz, quartzo-feldspathic gneiss, and biotite-rich layers. In some instances the rods represent transposed layering on the highly attenuated limbs of early, isoclinal minor folds. Near the northeast end of the roadcut such minor folds are easily seen due to the presence of more massive layers in the rock. Slabbed and polished specimens from this and similar outcrops demonstrates that these early folds are exceedingly abundant in the Piseco anticline. Examination of these folds shows that the dominant foliation in the rock is axial planar to them.

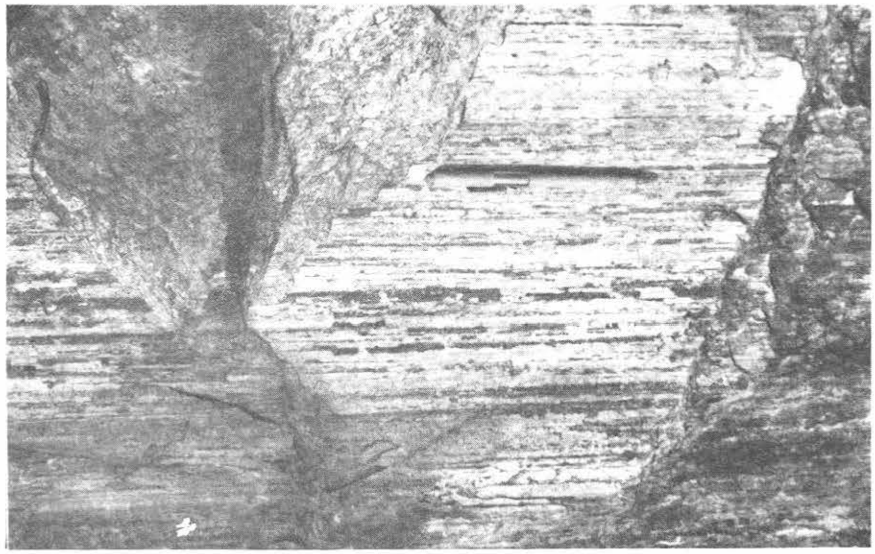

Fig. 8a. Example of ribbon lineation on a foliation surface of quartzofeldspathic gneiss of the Piseco anticline. The dark ribbons are quartz and the light ones K-feldspar.

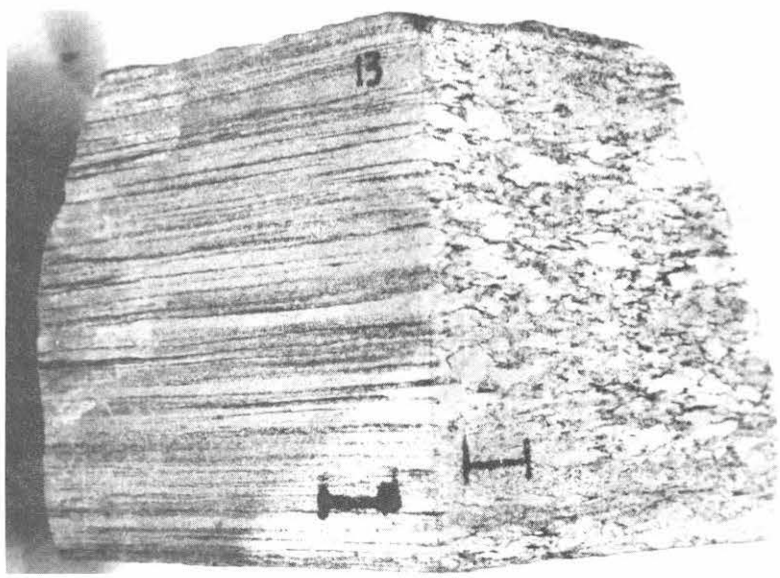

Fig. 8b. Quartzo-feldspathic gneiss of Piseco anticline cut perpendicular to foliation and parallel to lineation (left face) as well as perpendicular to lineation (right face). Note the elongation of the light colored K-feldspar augen in the direction of lineation. Bar markers are 1 cm long.

Fig. 9. The development of tails on K-feldspar augen in a gneiss less deformed than shown in Fig. 8. The face shown is perpendicular to foliation but parallel to lineation. The tails suggest a sinistral shear sense. The bar marker is 1 cm long.

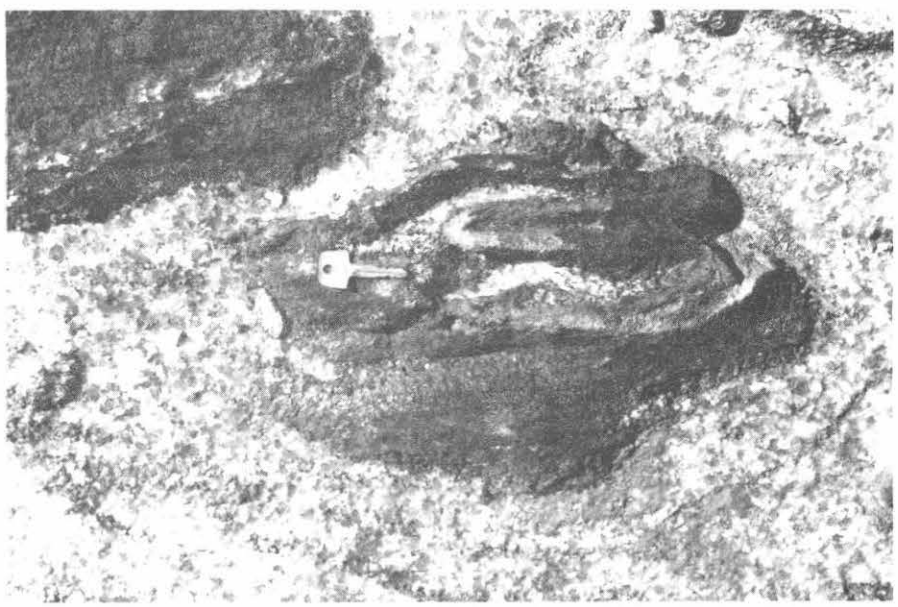

Fig. 10. Sheath fold developed in calcsilicate band in marble. Steep, down dip lineation visible on right hand side. Key is 5 cm long.

The lineation in these outcrops is shown in Fig. 8a while slabbed sections are shown in Figs. 8b,c. Fig. 8b is cut perpendicular to foliation and both parallel to (left face) and perpendicular to (right face) lineation. As can be seen K-feldspar augen have been elongated in the direction of lineation. Together with elongated quartz aggregates, these grain size reduced minerals form the prominent mineral lineation that characterizes foliation surfaces (Fig. 8a). Note that the K-feldspar augen exhibit shapes closer to equant on the right hand face at perpendicular to lineation. This strongly suggests that the rock fabric is the result of rotational strain in the direction parallel to lineation, i.e., the lineation is an elongation, or stretching type. Fig. 9 shows a less deformed sample slabbed perpendicular to foliation and parallel to lineation. The development of tails on K-feldspar augen are clearly visible. These are the result of grain size reduction during ductile rotational strain. A sinistral (east over west) sense of motion is indicated. At more extreme conditions of strain the K-feldspar augen and quartz aggregates are drawn into ribbons as in the present outcrop. Long dimensions of 40-60 cm are common along with thicknesses of a millimeter, or less. Clearly strain has been extreme and elongations of 30-40 times are not unusual.

Ribbon gneiss origin by rotational strain is also suggested by the parallelism between lineation and F_2, F_3 fold axes. It is believed that these fold axes were drawn into parallelism with the lineation by ductile, rotational strain directed from east to west. The most satisfactory mechanism for this configuration is the stacking of thrust sheets and thrust nappes during plate collision. Thrusts have been recognized to the east of Lake George (Fig. 4) and others probably exist although the intense, ductile nature of the deformation has resulted in extremely subtle truncations that are difficult to recognize. Sheath folds (Fig. 10) with tube axes parallel to lineation are consistent with this model. Presumably crustal thickening during the Grenville Orogeny was caused by the stacking of these thrusts.

Mileage

43.5 Junction NY Rt. 8 and NY Rt. 30 in Speculator. Head southeast on NY Rt. 8-30.

47 **Stop #7.** Northern intersection of old Rt. NY 30 and new Rt. NY 30, 3.3 miles east of Speculator, New York.

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

Near the west end of the outcrop a deformed layer of charnockite is well exposed. In other places the charnockite-marble interlayering occurs on the scale of one to two inches.

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

Commonly included in the Blue Mt. Lake Fm., but not exposed here, are quartzites, kinzigites; sillimanite rich, garnetiferous, quartz-microcline gneisses; and fine grained garnetiferous leucogneisses identical to those characterizing the Sacandaga Fm. These lithologies may be seen in roadcuts .5 mile to the south.

Almost certainly these marbles are of inorganic origin. No calcium carbonate secreting organisms appear to have existed during the time in which these carbonates were deposited (> 1 b.y. ago). Presumably the graphite represents remains of stromatolite-like binding algae that operated in shallow water, intertidal zones. If so, the other roadcut lithologies formed in this environment as well. This seems reasonable enough for the clearly metasedimentary units such as the quartzites and kinzigites. The shallow water environment is much more interesting when applied to the charnockitic and amphibolite layers. The fine scale layering, and ubiquitous conformity of these, strongly suggests that they do not have an intrusive origin. Perhaps they represent the metamorphosed products of volcanic material in a shelf like environment. Such intercalation is now occurring in many island arc areas where shallow water sediments cover, and in turn are covered by, ash and lava. Alternatively they may represent metasediments. A large number of minerals are developed within these outcrops. Both calcite and minor dolomite are present in the carbonate horizons. These are accompanied by green diopside and serpentinized forsterite as well as by tourmaline, grpahite, and variosu sulfides. In calcsilicate horizons phlogopite, diopside (white), and tremolite occur. Wollastonite is locally present. The presence of tremolite and wollastonite is believed to be a function of the relative concentration of CO_2 and H_2O in the vapor phase (Valley et al., 1983).

Mileage

47.5 Extensive roadcuts in lower part of Blue Mt. Lake Fm. Quartzites, kinzigites, and leucogneisses dominate. Minor marble and calcsilicate rock is present.

47.9 Large roadcuts in lower Lake Durant Fm. Pink, well-layered quartzo-feldspathic gneisses with subordinate amphibolite and calcsilicate rock.

49.0 **Stop #8.** One half mile south of southern intersection of old Rt. 30 and with new Rt. 30.

On the west side of the road small roadcut exposes an excellent example of Adirondack anorthositic gneiss intermediate in character between the so-called Marcy type (coarse) and the Whiteface type (fine grained). About 50% of the rock consists of fine grained crystals of andesine plagioclase. Some of these crystals appear to have measured from 6-8" prior to grain size reduction. Excellent moonstone sheen can be seen in most crystals. In places ophitic to subophitic texture has been preserved with the mafic phase being represented by orthopyroxene.

In addition to the anorthosite there exists a clearly cross-cutting set of late iron-rich orthopyroxene rich dikes containing xenoliths of coarse grained anorthosite. The latter may represent a late mafic differentiate related to cotetic liquids responsible for the ophitic intracrystalline rest magma. This would be consistent with the iron enrichment trend characteristic of Adirondack igneous differentiation.

Near road level there can be found several inclusions of calcsilicate within the anorthositic rocks. These are believed to have been derived from the Cedar River Fm. and are consistent with a non-basement status for the anorthosite.

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

Orthopyroxene + Plagioclase + Fe-bearing oxide + quartz =
garnet + clinopyroxene.

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

Mileage	
51.0	Cedar River Fm. Minor marble, amphibolite, and calcsilicate rock. Predominantly very light colored sillimanite-garnet-quartz-K-feldspar leucogneisses.
52.0	Junction NY Rt. 8 and NY Rt. 30. Continue south on NY Rt. 30. To the west of the intersection are roadcuts in leucogneisses of the Blue Mt. Lake Fm. A large NNE normal fault passes through here and fault breccias may be found in the roadcut and the woods beyond.
52.5	Entering Little Moose Mt. Fm. on northern limb of the Glens Falls syncline. Note that dips of foliation are to the south.
54.8	Entering town of Wells which is situated on a downdropped block of lower Paleozoic sediments. The minimum displacement along the NNE border faults has been determined to be at least 1000 meters.
58.3	Silver Bells ski area to the east. The slopes of the ski hill are underlain by coarse anorthositic gabbro intrusive into the Blue Mt. Lake Fm.
60.3	Entrance to Sacandaga public campsite. On the north side of NY Rt. 30 are quartzo-feldspathic gneisses and calcsilicate rocks of the Lake Durant Fm. An F_1 recumbent fold trends subparallel to the outcrop and along its hinge line dips become vertical.
60.8	Gabbro and anorthositic gabbro.
62.0	**Stop #9.** Pumpkin Hollow.

Large roadcuts on the east side of Rt. 30 expose excellent examples of the Sacandaga Fm. At the northern end of the outcrop typical two pyroxene-plagioclase granulites can be seen. The central part of the outcrop contains good light colored sillimanite-garnet-microcline-quartz gneisses (leucogneisses). Although the weathered surface of these rocks are often dark due to staining, fresh samples display the typical light color of the Sacandaga Fm. The characteristic excellent layering of the Sacandaga Fm. is clearly developed. Note the strong flattening parallel to layering and the lineation developed on many foliation surfaces. These gneisses are similar to so-called straight gneisses found in proximity to ductile shear zones. The Sacandaga Fm. may represent mylonitic rocks of this type and its layering may, in fact, be tectonic.

Towards the southern end of the outcrop calc-silicates and marbles make their entrance into the section. At one fresh surface a thin layer of diopsidic marble is exposed. NO HAMMERING PLEASE. Many "punky" weathering layers in the outcrop contain calc-silicates and carbonates.

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

No angular discordance or other indications of unconformity can be discerned at the base of the Sacandaga Fm. However, this does not preclude the prior existence of an angular discordance which may have been swept into pseudoconformity by tectonism.

Along most of the roadcut there can be found excellent examples of faults and associated pegmatite veins. Note that the drag on several of the faults gives conflicting senses of displace-ment. The cuase of this is not known to the author. Also note the drag folds which indicate tectonic transport towards the hinge line of the Piseco anticline.

Mileage	
62.5-67.0	All exposures are within the basal quartzo-feldspathic gneisses at the core of the Piseco anticline.
67.0	Re-enter the Sacandaga Fm. Dips are now southerly.
68.0	In long roadcuts of southerly dipping quartzo-feldspathic gneisses of Lake Durant Fm.
70.4	Cross bridge over Sacandaga River.
74.4	Bridge crossing east corner of Sacandaga Reservoir into Northville, New York.

END LOG

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Trip C-12

GROUNDWATER CONTAMINATION FROM A SANITARY LANDFILL: INVESTIGATIONS, REMEDIAL ACTION AND ENERGY RECOVERY AS A POSSIBLE SOLUTION

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Purpose

The purpose of this field trip is to view the adverse impacts resulting from the improper disposal of solid and hazardous wastes. A tour of the Rome Sanitary Landfill will reveal the impacts on the environment resulting from hazardous waste landfill leachate. Leachate can be seen at the landfill and in nearby streams. The participants will be able to view the remedial measures being conducted at the site, including, the construction of a barrier cap, gas vents and monitoring wells. A discussion on the elements of the site investigation and remedial actions utilized will be conducted while touring the facility. Questions are encouraged! Later in the day the participants will tour the Oneida County Energy Recovery Facility in Rome, N.Y. This facility, which will not be operational at the time of the field trip, is one of Oneida County's solutions to the land burial of industrial and municipal solid waste. Although the facility will not be operational, all equipment will be on-line and will be reviewed.

INTRODUCTION

According to USEPA estimates (1979) over one million tons of municipal and industrial solid wastes is generated and disposed of daily. The disposal of these wastes is becoming an increasing problem. As new materials are developed and marketed yearly, the safe handling and disposing of them is an ever increasing dilemma.

Improper disposal of industrial and municipal wastes has led to numerous pollution problems. In the past, much attention has been focused upon the visible consequences of the dump, the destruction of natural resources and wildlife habitats, or the contamination of air and surface water. Only recently has much attention been focused on the seriousness of groundwater contamination resulting from improper landfilling.

Dealing with the problem of safely transporting and disposing of massive quantities of solid waste is a large task. This problem has received much attention at all levels of government within the last decade.

Solid waste and hazardous waste issues are a high priority for the Oneida County Department of Planning and Environmental Management Council (EMC). The EMC staff has devoted much time and effort to the problems and possible solutions to the solid waste problem in Oneida County. The Council has reviewed conditions at landfills in the County. Staff has reviewed both in-house and DEC files on selected landfills in the County. At times, staff conducted site investigations, and provided technical assistance to DOH, DEC, and local bodies of government.

This report presents the results of a detailed hydrogeologic and water quality investigation of a sanitary landfill in Rome, New York.

In July of 1979, Dunn Geoscience Corp. entered into a contract with the NYS Dept. of Environmental Conservation (DEC) Division of Solid Waste, to determine what impacts, if any, selected municipal landfills across New York State were having on groundwater. This is a report on the elements of a site investigation and remedial actions utilized at the Rome SLF. This hydrogeologic investigation is typical of investigations being carried out at numerous selected landfills located throughout New York State.

Site Description

The landfill is located on Tannery Road just north of Route 49 (figure 1) in the City of Rome, Oneida County, N.Y. This landfill is located within a swampy area contained within a region known as the Rome Sand Plains. Several sand dunes which are characteristic of the topography of the Sand Plains are located at the landfill. The land surrounding the landfill is a rural residential area.

Elements of a Site Investigation

Investigative techniques utilized in this hydrogeologic study are typical of investigations conducted at landfill sites throughout New York State. These work items include the following:

A) Background Investigation: Review of file data and published information;
B) Non-Drilling Activities: Initial site inspection, existing on-site and off-site sampling, geophysical surveys, and remote sensing;
C) Drilling/Sampling: Monitor well installation, soil borings, water level measurements, and groundwater sample collection and analysis;
D) Site mapping;
E) Interpretation and evaluation of data;
F) Conclusions and report preparation; and
G) Site remedial actions.

I. Review of Site File Data and Published Information

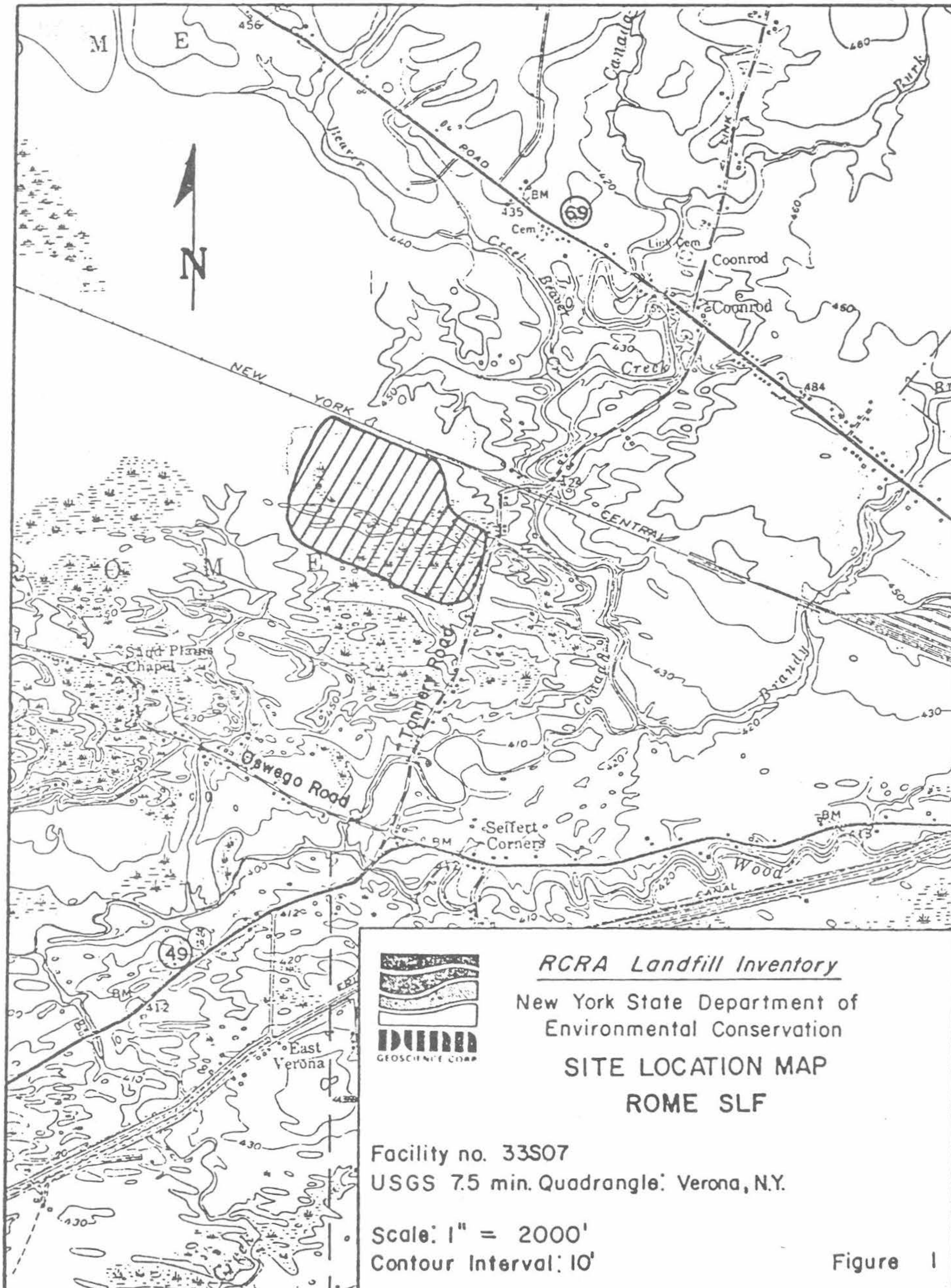
This included an examination of Department of Environmental Conservation (DEC) files and any published information on the Rome SLF and the surrounding area. These sources were examined in order to provide information on past waste disposal practices at the site, and types of wastes received. Published information on geology of the area was examined, where possible, in order to aid in the interpretation of site hydrology and geology.

II. Results of Non-drilling Activities

A. Site Geology

Bedrock:

The bedrock underlying the Rome SLF is the Ordovicean Utica shale. The bedrock is generally flat lying.



RCRA Landfill Inventory

New York State Department of Environmental Conservation

SITE LOCATION MAP
ROME SLF

Facility no. 33S07
USGS 7.5 min. Quadrangle: Verona, N.Y.

Scale: 1" = 2000'
Contour Interval: 10'

Figure 1

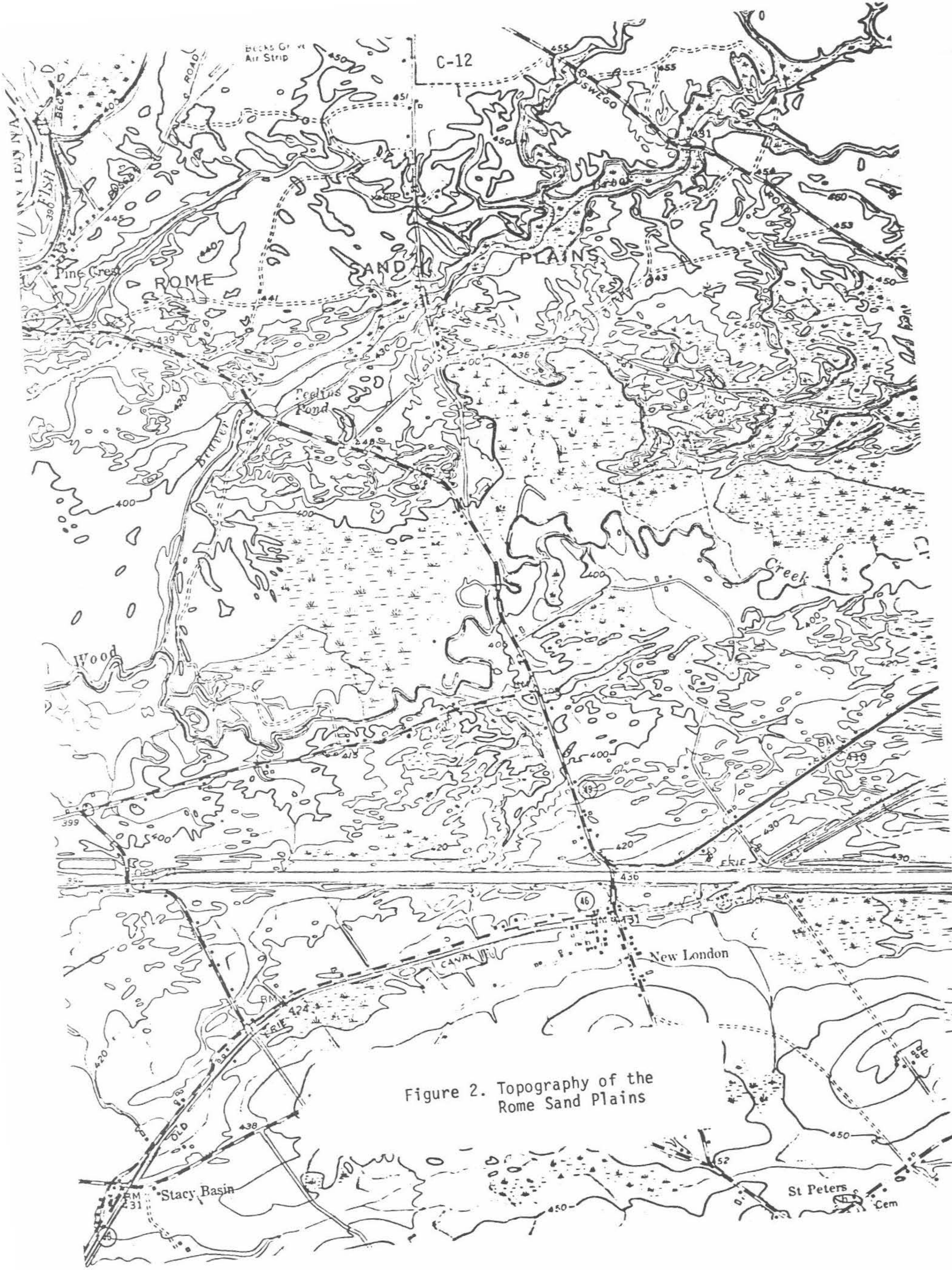


Figure 2. Topography of the Rome Sand Plains

Surficial Geology:

Overlying the bedrock is a series of glacial sediments of late Pleistocene age. The first unconsolidated unit to be deposited was a thin layer of glacial till which mantles the bedrock. Overlying the glacial till is a layer of lacustrine sand and clay which was deposited in glacial Lake Iroquois (fig. 2). When the lake drained, it left large sand deposits which were wind-blown into large crescent shaped dunes. Depressions on the leeward side of the dunes accumulated water and wetlands were formed.

Hydrology

Surface Water:

The regional surface drainage patterns are not well defined due to the override of the last major glacier. Due to the presence of very porous sandy soil at the surface, precipatation infiltrates the ground rapidly resulting in minimum run-off.

Canada Creek is located adjacent to the site. The Creek flows north to south in close proximity to the east boundary at the site. To the south, the Canada Creek joins Wood Creek before the Creek enters Oneida Lake at Sylvan Beach.

Groundwater:

The regional groundwater flow pattern parallels the general flow direction of surface drainage. Groundwater flow is controlled both regionally and locally by the existing topography and the distribution of the unconsolidated aquifer deposits.

Underlying the site area is an aquifer comprised of lacustrine sands. The lacustrine sand can produce fair quantities of groundwater whenever the sands are below the water table. Due to the presence of fine grained sands, the transmissibility of the aquifer is limited.

Little is known about the shale aquifer in the area. The groundwater flow within the shale is controlled by fractures and joints and yields are very low. An analysis of the well logs indicates that the bedrock aquifer is not influenced by the landfill because two aquicludes of lacustrine clay and till are present between the landfill and the bedrock.

Site Groundwater Hydrology

The groundwater flow direction is a radial pattern from the landfill with flow towards the Canada Creek and towards the south and west (figure 3).

The groundwater gradient is gentle approximately one foot drop per 250 feet horizontal distance. This gradient is due, in part, to the flat terraine and the presence of the underlying clay aquiclude.

Determination of the groundwater gradient and construction of the water table contour map were based upon the evaluation of water levels measured in the monitoring/observation wells, and results of the resistivity survey.

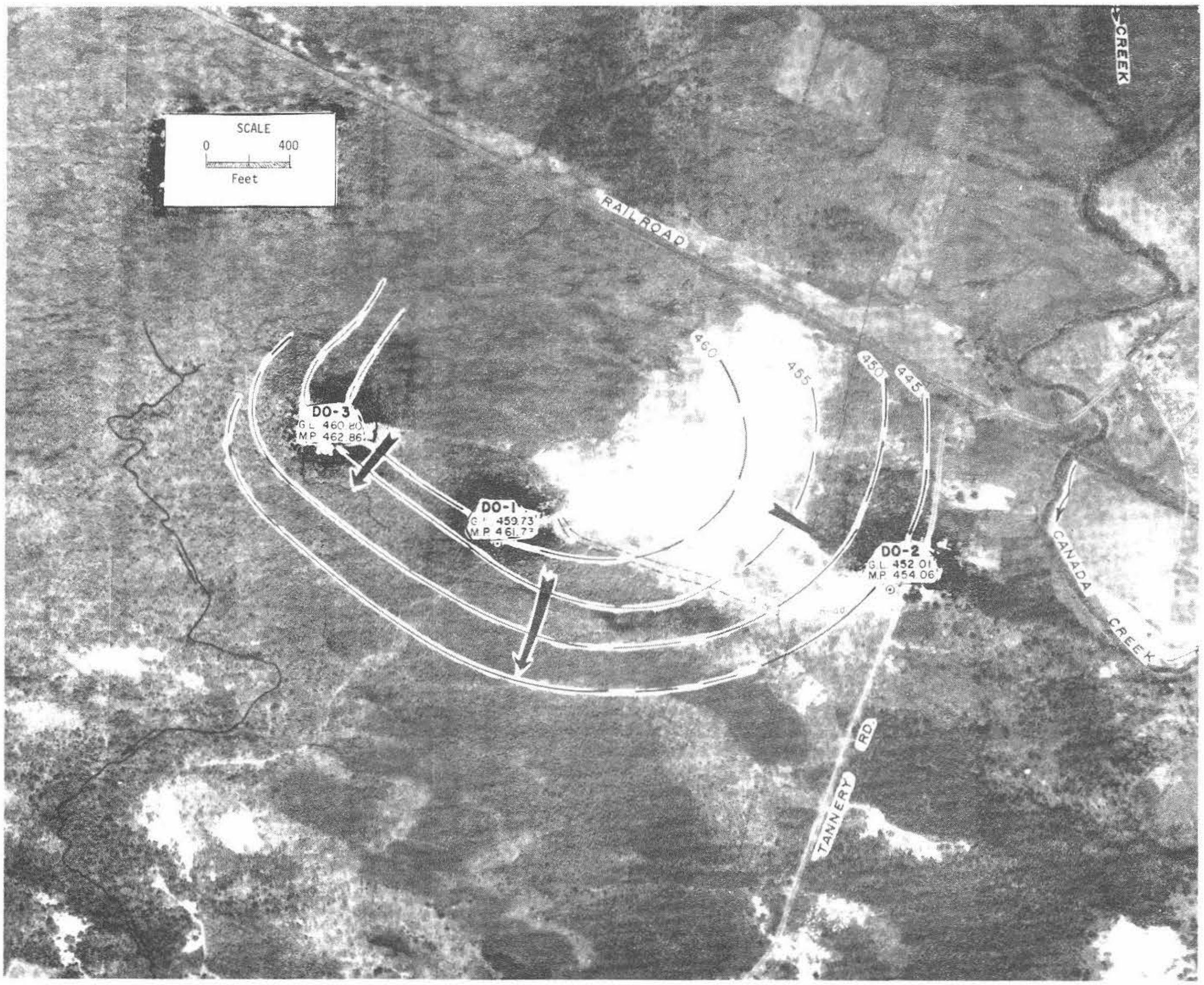


Figure 3. Water table contour map, Landfill Facility No. 33S07, Rome SLF. (Numbered lines give elevation of water table in feet; heavy arrows show direction of groundwater flow;circled dots show positions of monitoring wells) Map by Dunn Geoscience Corp, June 1980.

The actual design and construction of the wells will be reviewed under section heading Drilling/Sampling Program.

Geophysical Resistivity Survey:
An electrical earth resistivity survey (EERS) of the immediate area surrounding the landfill was conducted to determine if there were any areas of leachate contamination of groundwater. Landfill leachate increases the conductance of groundwater and, hence, lowers the bulk resistivity of the aquifer in a contaminated zone. Electrical earth resistivity surveys can also provide a good indication of water table elevation.

Results of the resistivity survey indicated that the groundwater is contaminated. Downgradient well samples had a conductivity that indicated groundwater contamination from landfill leachate. In addition, a small stream that runs adjacent to the landfill appeared to be contaminated by leachate.

III. Drilling/Sampling Program

A.) Monitoring Well Installation:
Based upon the results of the site inspection and resistivity survey, it was determined that three monitoring wells would be needed (1 upgradient, 3 downgradient) around the landfill.

Three monitoring wells were installed by a private drilling outfit (subcontracted by Dunn Geoscience). Figure 4, A, B, C, shows the construction details, depth, screen position, and major subsurface units. A rotary drilling rig was mobilized at the site in order to obtain soil borings (samples of subsurface materials). The subsurface materials were sampled by a split-spoon sampler two out of every five feet according to ASTM standard sampling procedures. All samples were logged (fig. 5) as to sample interval, blow counts, and material type.

When the bottom of the boring was reached, the hole was cleaned out with clean drilling water, removing any excess material. It was then converted to a monitoring well by installing a two-inch OD PVC pipe and well screen in the water-producing zone. Figure 6 shows the construction of a typical well.

The monitoring wells provide a means for measuring the groundwater level at each of the well sites. An electric tape is utilized to measure water levels to an accuracy of 0.01 feet. These level readings provide the basis for the evaluation and determination of groundwater conditions over the site. They were used to construct the groundwater table contour map and to determine the direction of groundwater flow and the groundwater gradient (see figure 3).

B.) Groundwater Sample Collection and Analysis:
Sample withdrawal, transportation, preservation, and analysis must be conducted with extreme care in order to maintain sample integrity. Groundwater sample collection must be in accordance with the standard procedures of the United States Environmental Protection Agency and the American Public Health Association.

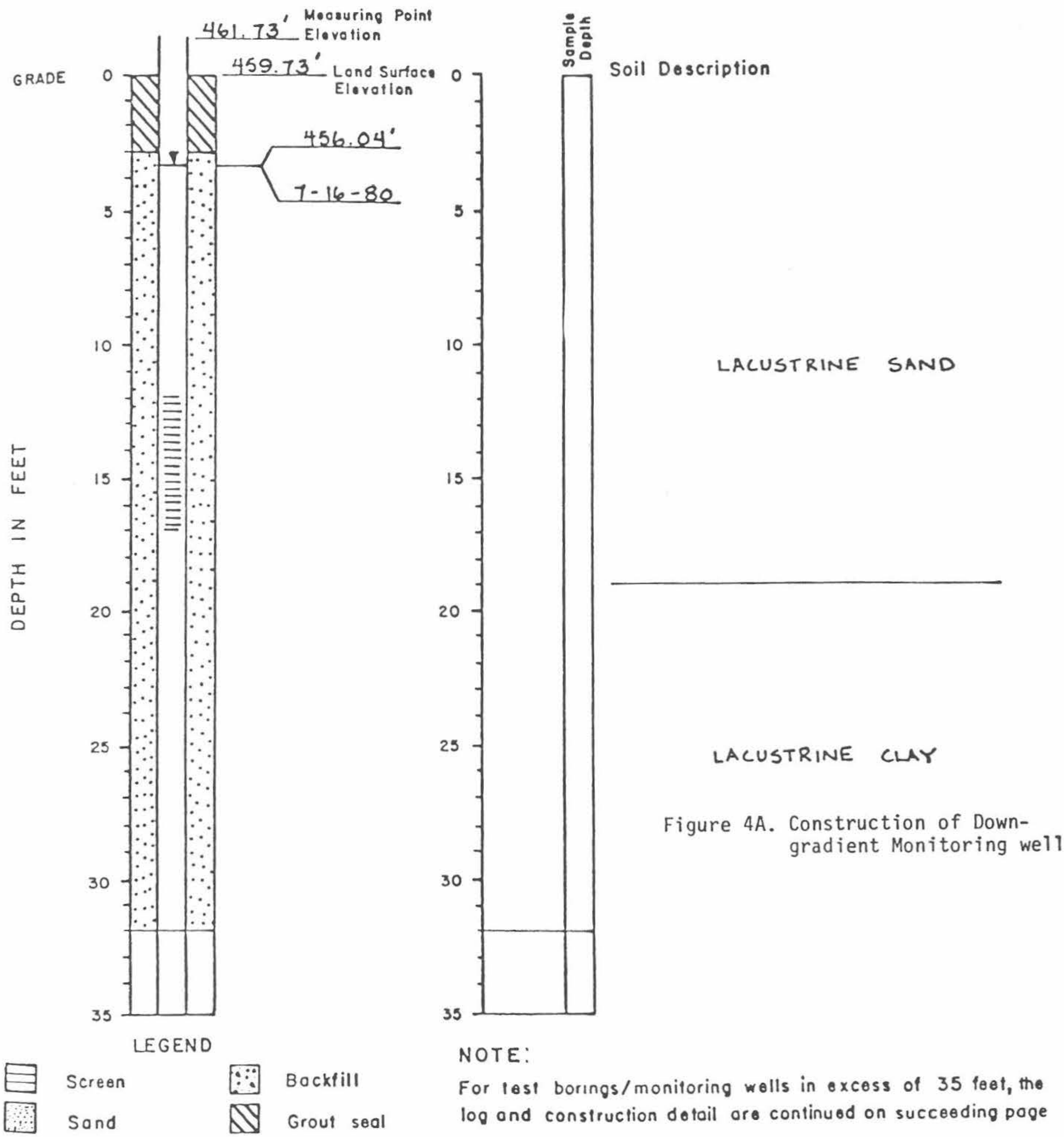


Figure 4A. Construction of Down-gradient Monitoring well.

NOTE:

For test borings/monitoring wells in excess of 35 feet, the log and construction detail are continued on succeeding page

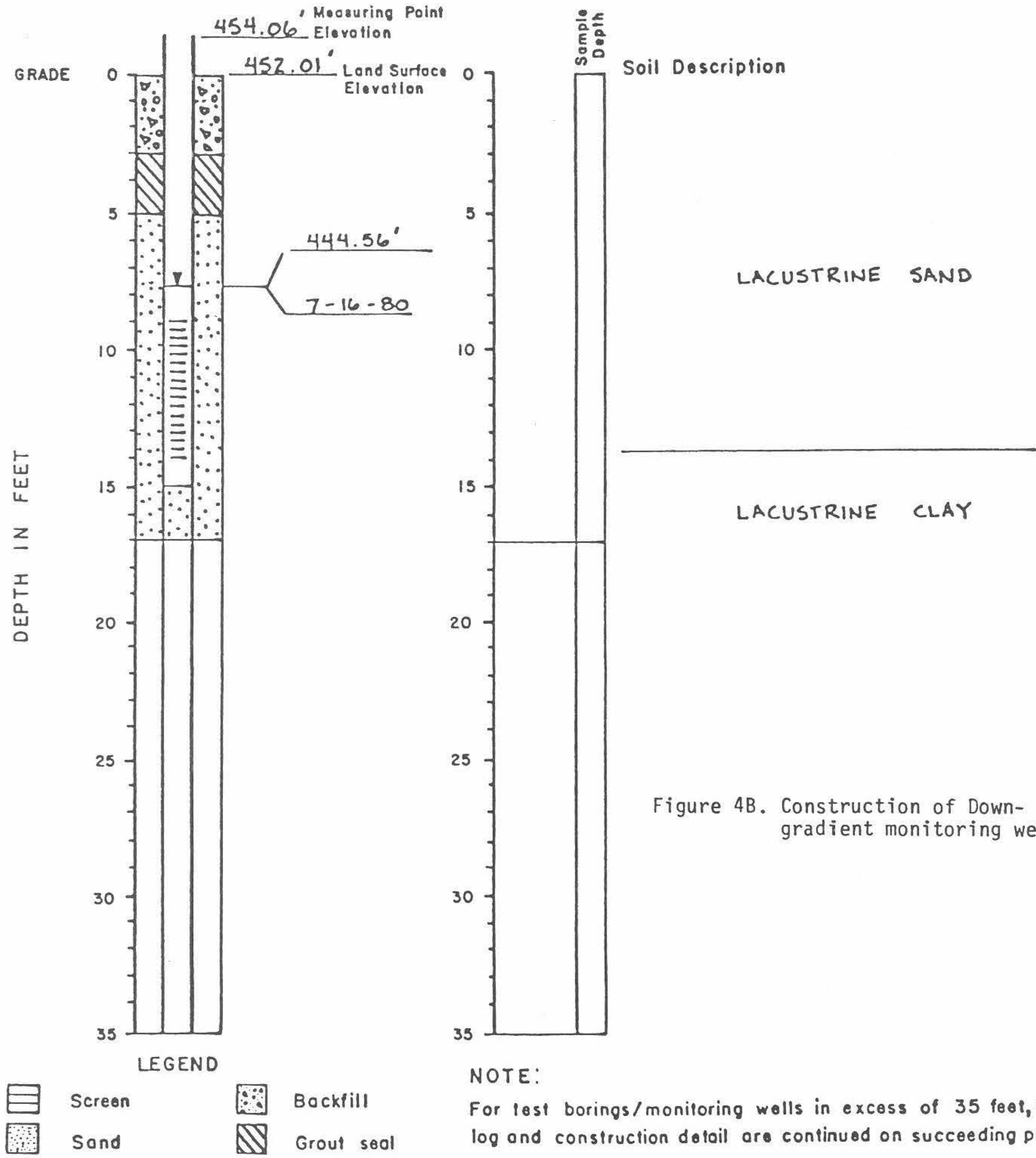


Figure 4B. Construction of Down-gradient monitoring well.

NOTE:

For test borings/monitoring wells in excess of 35 feet, the log and construction detail are continued on succeeding pages.

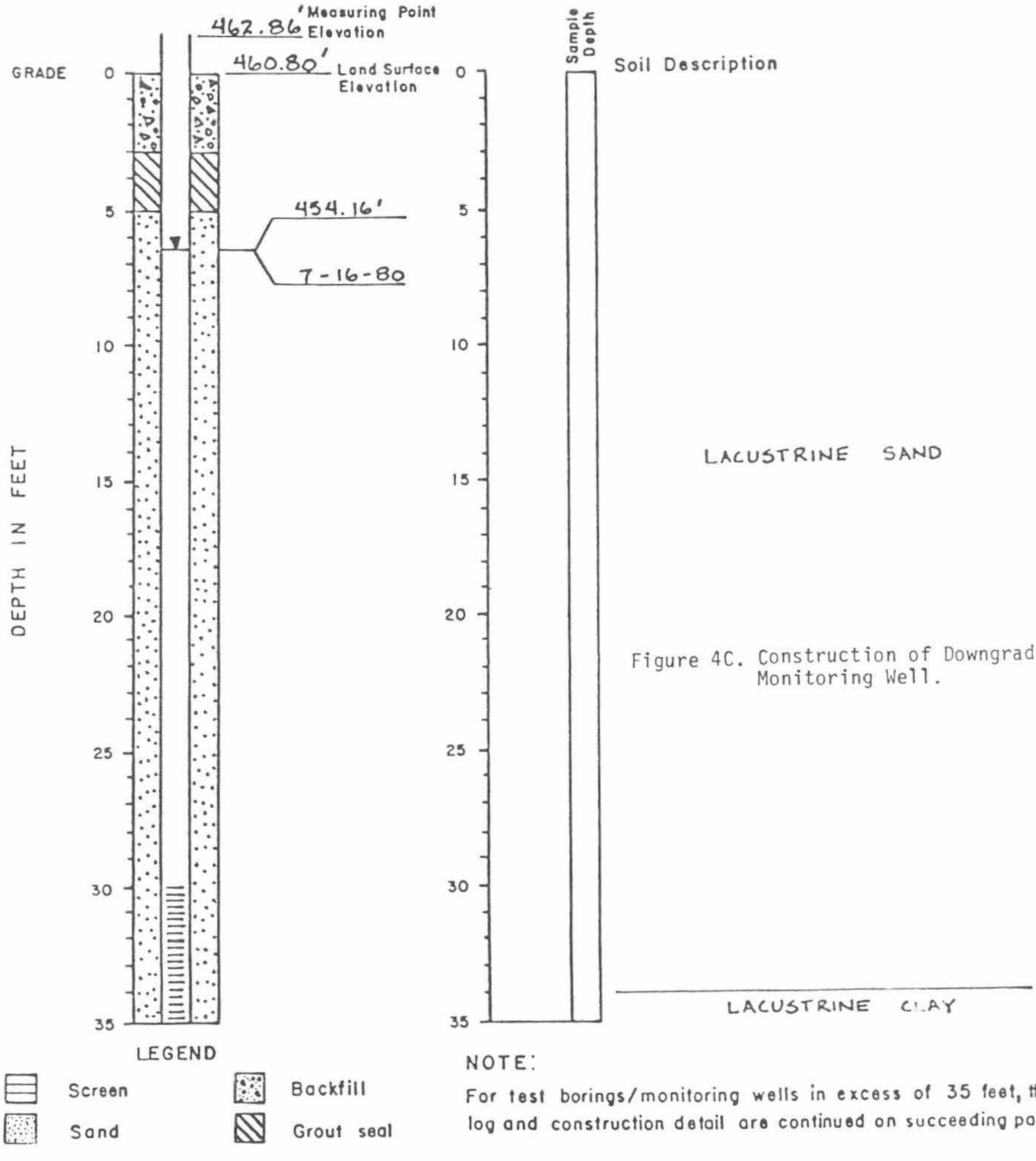


Figure 4C. Construction of Downgradient Monitoring Well.

NOTE:

For test borings/monitoring wells in excess of 35 feet, the log and construction detail are continued on succeeding pages

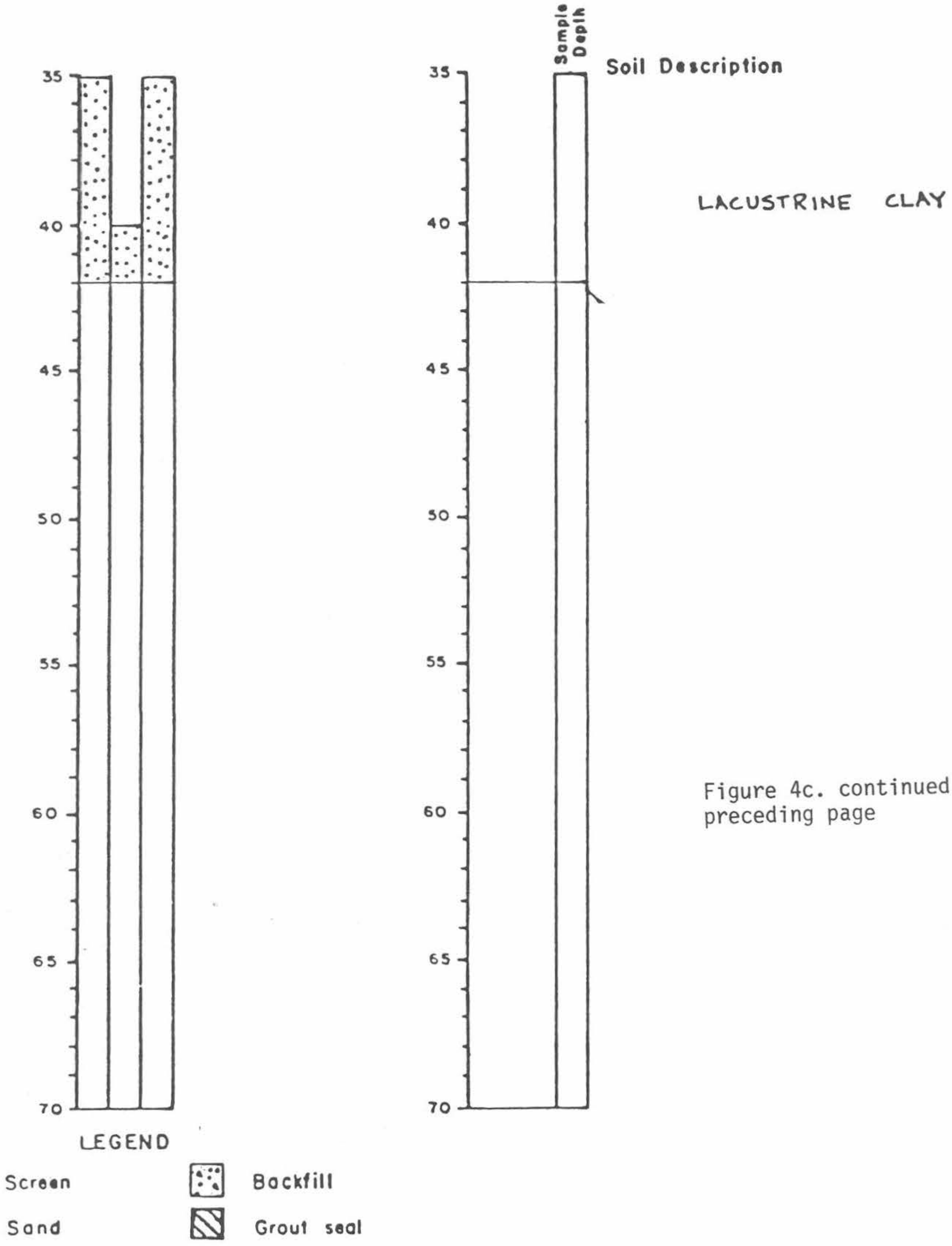


Figure 4c. continued from preceding page

...GR Chisick	SURFACE ELEV.	SAMPLING	WATER LEVEL OBSERVATION
DRILLER Rappold	BORING STARTED 5/6/80	SS SIZE 1 3/8" ID 2" OD	WL: ____ WS CR WD
HELPER Lodge - George	BORING COMPLETED 5/6	HAMMER 140 DROP: 30"	WL: ____ BCR ____ ACR
RIG NO. J-2	STATION On	ST SIZE ____ ST SIZE ____	WL: ____ AB ____ Hr. AB
	OFF SET	CASING USED 17.0' SIZE 4"	WL: ____ 24 Hr. AB

JOB NO. 72040 F BORING NO. RA-1 CLIENT Dunn Geo. WEATHER Rain Cloudy TEMP. 60

Boring Location: Rome N.Y. landfill

Sample No.	Depth or Elevation From	Depth or Elevation To	Sampling Method	Hydraulic Pressure PSI: Time Sec. Hour	Pressure While Sampling	Pressure While Coring	Split Spoon Blows 6"	6"	6"	6"	Casing Blows Per Foot	R Length Recovered in Feet	Qp Penetrometer Test in TSF	Sample Description
1	0.0	2.0	SS				2	5	9	11		1.5		Brown, fine sand
	2.0	5.0	SS											Wet @ 4.5'
2	5.0	7.0	SS				18	19	20	22		1.7		Brown, fine sand (wet)
	0.0	12.0	RB				Casing to 10.0'							
3	10.0	12.0	SS				11	15	11	11		1.2		Brown, fine sand (wet)
	10.0	15.0	RB				Casing to 15.0							
4	15.0	17.0	SS				8	14	27	31		1.6		Brown,fine sand (wet) change @ 16' to
4A							Casing to 17.0							Grey clayey silt
	15.0	20.0	RB											
5	20.0	22.0	SS				19	15	17	22		1.5		Grey clay, sile silt
	20.0	25.0	RB											
6	25.0	27.0	SS				4	7	9	10		1.7		Grey clay, some silt
	25.0	30.0	RB											
	30.0	32.0	SS				2	5	7	7		1.7		Grey clay, some silt

End of boring 32.0'

Monitoring well - set with tip @ 32' B.G.

Wellscreen from 12' - 17' B.G. Silica sand

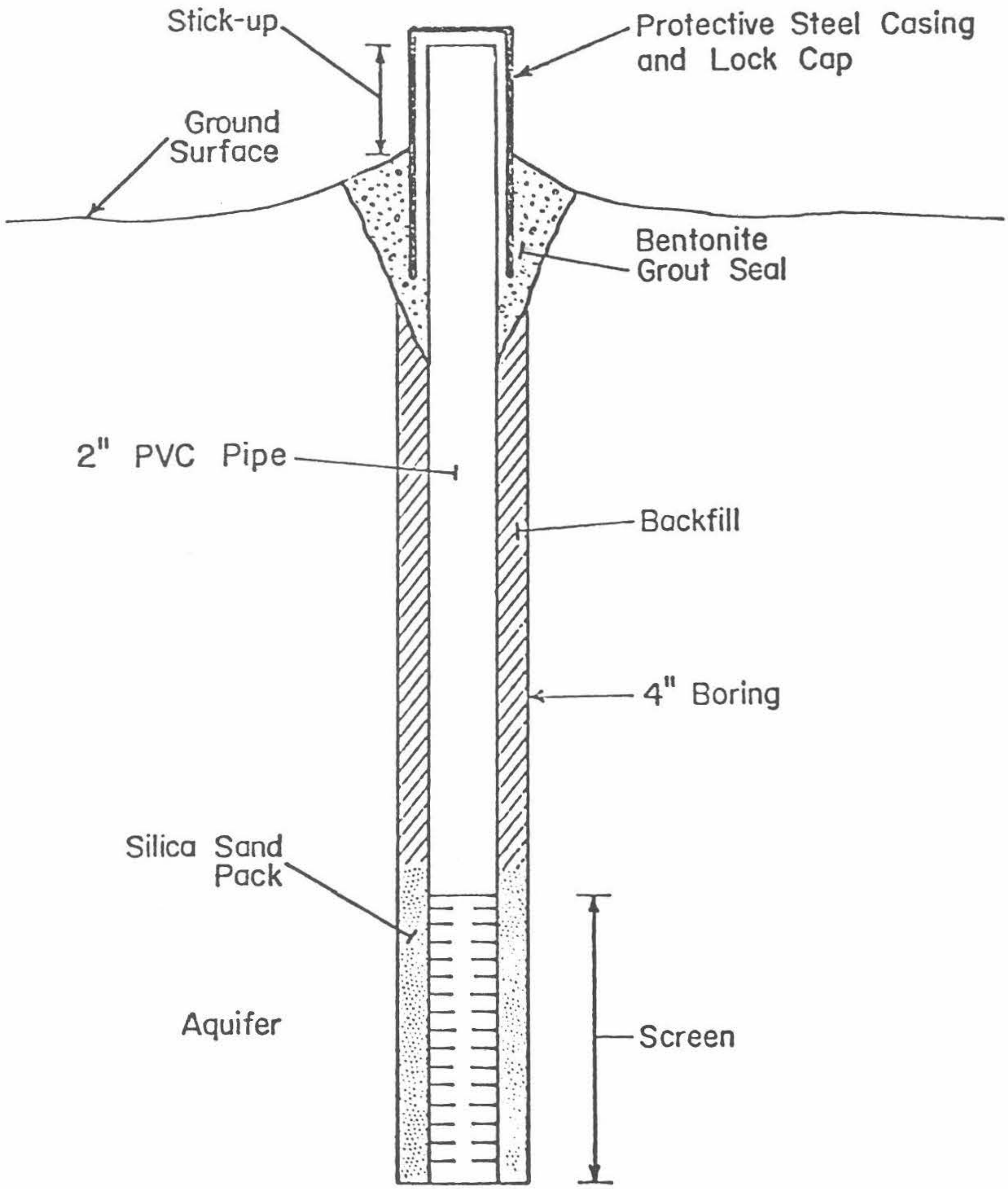
backfill with steel casing grouted @ surface

ABBREVIATIONS
F.T. - Fish Tail
W.O. - Wash Out
S.S. - Split Spoon
D.B. - Diamond Bit
P.A. - Power Auger
R.B. - Rock Bit
W.S. - While Sampling
W.D. - While Drilling
B.C.R. - Before Casing Removal
A.C.R. - After Casing Removal
A.B. - After Boring

Figure 5. Typical Drillers Log.

Figure 6.

TYPICAL MONITORING WELL CONSTRUCTION



In order to insure that the samples being withdrawn are uncontaminated from equipment use or from other unsuspected materials, all the wells are bailed, at least one column of water, with stainless steel bailers. To avoid cross-contamination, a different stainless steel bailer is used for each well.

C.) Sampling Protocol

The sampling protocol required an indicator scan of 8 parameters and a baseline scan of 23 parameters. The indicator scan included the following 8 parameters: Conductivity, Eh, PH, Temperature, Chloride (Cl), Iron (Fc), Total Dissolved Solids (TDS), and Total Organic Carbon (TOC). Baseline scan included the eight indicator parameters plus Cadmium, Chromium, Lead, Selenium, Mercury, Copper, Zinc, Aluminum, Manganese, Sulfate, Nitrate, Color, Odor, Hardness, Alkalinity, and Phenols.

The baseline series was run in order to determine if the groundwater had been contaminated by leachate. Baseline scans, in general, are conducted when the results of the indicator parameters suggest groundwater contamination. In this investigation the indicator parameters as well as DEC data on the landfill revealed the potential for contamination.

TOC, nitrate, and phenols are indicators of organic contamination. The analysis indicated that TOC and phenols are elevated in 2 downgradient monitoring wells. Thus, there is evidence of some organic contamination.

IV. Interpretation and Evaluation of Data

Analysis of resistivity data, well logs and water quality analyses indicates that there are zones of low, medium and high quality groundwater around the landfill. These zones are representative of varying degrees of leachate contaminated groundwater. The data suggests some organic contamination and that the leachate is "weak."

Based on the results of this study, the landfill complied with the RCRA, Part 257, Groundwater Quality Criteria at this time. However, these data suggest that leachate from the landfill is entering the groundwater system.

V. A.) Site Remediation

After the site investigation is completed, and the analysis of the data indicates potential pollution of surface water, groundwater, air and/or all three major environmental pathways, it is necessary to select the proper site remediation necessary to maintain a level of control which would prevent further environmental degradation. The necessary site remediation chosen will depend on such factors as: 1) economic feasibility; 2) nature of contamination; 3) specific site conditions; and 4) level of environmental risk.

There are a myriad of existing remediation techniques available to abate environmental degradation from landfill leachates. The major remedial categories include: 1) surface controls; 2) groundwater controls; 3) leachate collection and treatment; 4) gas control systems; and 5) direct waste treatment.

A discussion of all the major remedial categories would be exhaustive and lengthy. In addition, it is not the intent of this report to discuss in detail the various remedial actions. Therefore, a list of the major remedial actions is provided in Appendix A. Illustrative diagrams are also provided in Appendix A for several of the major control techniques.

B.) Remedial Action at the Rome SLF

The remedial action utilized at the Rome SLF was specified by the NYSDEC. On June 2, 1982, the Mayor of the City of Rome signed a Department of Environmental Conservation Consent Order, thereby agreeing to upgrade the Rome Landfill according to a specified schedule of remedial action.

The Consent Order requires the site be capped in three stages in compliance with 6NYCRR Part 360 (Solid Waste Management Facilities) which specifies the minimum requirements for a landfill cap (see fig. 1, appendix A). Construction of the cap will be from off-site materials placed in layers and compacted to predetermined densities and permeabilities (coefficient of permeability of 1 x 10 -5 cm/sec).

When the barrier layer has been constructed and its permeability certified, a one-foot minimum layer of sandy material will be placed on top of the barrier cap. Overlying the sandy layer will be a layer of topsoil which will be seeded.

Landfill gas vents are installed on a 200-foot grid around the landfill. Water samples will be taken quarterly from the three on-site monitoring wells, and any other wells DEC indicates.

Parameters to be measured for include: Chlorides, Specific Conductivity, Total Organic Carbon, PH, Total Iron, Total Dissolved Solids.

Closure plans do not require the construction and placement of a leachate collection system, liners, or subsurface drainage ditches.

Oneida County Energy Recovery Facility An Alternative to the Land Disposal of Solid Wastes

I. INTRODUCTION

BACKGROUND

Landfill disposal in Oneida County has and currently is a responsibility of the towns, cities and villages. In the late 1960's Oneida County recognized the growing solid waste disposal problem was an intermunicipal problem and together with Herkimer County retained a consulting firm to prepare a comprehensive solid waste plan. That plan, completed in 1969, called for the creation of two regional landfills to serve Oneida County; one in Rome and one north of Utica in Trenton. The plan met with vocal opposition from those opposed to the philosophy of burying garbage (wasting a natural resource); those from Rome and Trenton who didn't want waste from "the outside" brought into their community; and those individuals living near the selected landfill site.

As a result of the opposition, the County created the Solid Waste Agency and asked this group of citizens to study resource recovery as an alternative to landfill. The Agency's studies led them to pursue energy recovery from municipal waste as the most appropriate technology for Oneida County. Talks began with Griffiss AFB in 1973 following the fuel shortage of that year and developed over the years to full fledged contract negotiations during 1978-79. The negotiations culminated on December 13, 1979 when Griffiss forwarded steam purchase and waste disposal contracts to the County for signature.

Prior to County authorization each participating municipality signed contracts during February and March 1980 to insure a waste supply. In May, the County Board of Legislators authorized the County Executive to sign the Air Force and municipal contracts, which he did in June. In July, the Board approved the necessary bonding resolution to finance the project.

II. PROJECT DESCRIPTION

The facility will have a design capacity of 200 tons of waste per day using four Modular Incineration Units (MCU) with waste heat recovery. Waste will be accepted from the City of Rome, Town of Floyd, Griffiss Air Force Base, and from NOCO and SWOCO landfill service areas. The steam produced by the plant will be purchased by Griffiss Air Force Base year round. Access to the plant will be directly off State Route 365 (River Road). The plant will have a baghouse on the stacks to clean air emissions. A steam condenser will allow the facility to incinerate waste and pass the cooled exhaust gases through the baghouses when the steam demand is lower than the steam production during the summer months. Wastewater generated at the facility will be treated at the Rome sewage treatment plant. The ash residue will be milled to reduce the metal content and leaching potential prior to burial.

PLANT SITE

The plant will be located (Appendix B, figure 1) on a parcel of land containing approximately 75 acres. The property adjoins Griffiss AFB next to the B-52 bomber alert area. The energy recovery plant will be located about 1,500 feet from the B-52 bombers, about 1,600 feet west of Rickmeyer Road and 2,000 feet north of State Route 365 (River Road).

HAZARDOUS WASTE

Hazardous waste will not be accepted at the plant. Waste coming into the plant will be screened while it is dumped on the tipping floor and when the incinerator hopper is being charged. Any waste that is hazardous or in some way might cause damage will be rejected and removed from the incinerator waste stream. This can be easily accomplished with the relatively small handling equipment (skid steer loader) and the ability of the operator to view the waste being placed into the charging hopper.

The EPA maintains a list of hazardous substances which if generated with other listed hazardous wastes in excess of 2,200 pounds per month must be reported to EPA. This number invokes the EPA hazardous waste manifest (cradle to grave) reporting system. Those handling any mixture of hazardous wastes comprising the minimum 2,200 pounds or more are required to have a manifest accompany it at all times. The New York State Department of Environmental Conservation has proposed reducing that reporting threshold to 220 pounds per month in 1981 for New York State generators of a hazardous waste.

Generators and disposers of hazardous waste who violate the manifest system or improperly dispose of hazardous waste are subject to both civil and criminal penalities. The manifest system is designed to direct hazardous wastes to facilities that are licensed to accept and dispose of such wastes. The Oneida County facility will not accept such hazardous wastes, nor should any such wastes be accepted at the landfills operating in Oneida County.

Most of the substances on the EPA hazardous waste list are liquids or gases and as such would not be accepted at the energy recovery facility in any event. Before the plant is operational, a survey of industries in the area to be serviced by the energy recovery facility will be made. This "one to one" meeting with local industry will identify wastes which are considered to be unacceptable (i.e. hazardous) or incompatible with the incineration equipment.

SOLID AND HAZARDOUS WASTE PROBLEMS AND RECOMMENDED SOLUTIONS

The Oneida County EMC has devoted much time and effort to the problem and possible solutions to the solid waste problem in Oneida County. In August of 1983, staff drafted a statement presented to the NYS Assembly Standing Committee on Environmental Conservation. This statement outlined a basic philosophy that efforts should be made both to remove existing impediments to reducing reliance on landfilling and to increase utilization of proven solid waste management technologies. This effort was accomplished by EMC staff working with the Solid Waste Agency (subcommittee of the EMC), the County Planning Department, and the Department of Public Works Division of Solid Waste. It was pointed out that a direct result of lack of enforcement of the Part 360 Regulations (Solid Waste Management Facilities) could be threats to public health and the environment from waste leachate. In the absence of equal treatment and enforcement by the State, the economic advantage weighs heavily in favor of existing facilities often run in violation of Part 360. The EMC has made specific recommendations to improve solid waste management at the State level as follows:

1. There should be full and equal enforcement of Part 360 NYCRR for all solid waste facilities, either existing or proposed.
2. There should be a clear and consistent state policy on solid waste which emphasizes resource recovery, source separation and recycling and discourages new and continued landfilling. This policy should be translated into positive identifiable actions which promote new technologies instead of stifling such advancement.

3. There should be technical assistance available to evaluate appropriate solid waste technology and assist local governments with its implementation.
4. There should be direct financial aid to solid waste facilities which employ systems that alleviate dependence on landfills.
5. There should be a coordinated review within a pre-set time for new solid waste facility proposals to promote their development.

Appendix A

List of Major Remedial Actions
and
Representitive Diagrams

I. Surface Controls

  a. Caps (see figure 1 appendix A)

    1. Purpose

    2. Types of materials

      a. clay

      b. clay, sand, and gravel

      c. clay and mix (bentonite, lime, etc.)

      d. sprayed bituminous membranes

      e. synthetic membrances

      f. industrial residues (flyash, slag)

  b. Run-on/run-off Controls

    1. Berms

    2. Ditches

    3. Benches

    4. Basins/lagoons

II. Groundwater Controls

  a. Impermeable Barriers

    1. Slurry walls (see figure 2 appendix A)

      a. bentonite

      b. cement-bentonite

      c. vibrated beam

    2. Grout curtains

      a. suspension (bentonite, cement)

      b. chemical (silicates, lignochrome, acrylamide)

    3. Sheet piles (various arrangements)

  b. Permeable treatment beds

    1. Limestone

    2. Activated carbon

    3. Ion exchange (zeolites, synthetic)

c. Pumping

1. To lower water table (see figure 3 appendix A)

2. To prevent underlying aquifer contamination (see figure 4 appendix A)

3. To contain plume (see figure 5 appendix A)

a. extraction

b. injection

d. Treatment

1. Air stripping of VOC packed tower

2. Carbon adsorption

3. Ion exchange

4. Pretreatment requirements

e. Alternate water supply

1. Local household treatment

2. New wells

3. Extension of municipal service

4. Well-head treatment

5. Blending

6. Capital improvements

III. Leachate Collection and Treatment

a. Drains

1. Subsurface

2. Ditches

b. Liners, as

1. Collection aids

2. Interceptors

c. Treatment

1. Activated sludge

2. Activated carbon

3. Air stripping

4. Physical/chemical

IV. Gas Control Systems

a. Pipe vents

1. Atmospheric

2. Forced ventilation

b. Barriers

c. Collection - Recovery

d. Treatment

1. Carbon adsorption

2. Burn-off

V Direct Waste Treatment

a. Siting

1. Secure landfill

2. Incorporate into site closure

b. Excavation

1. Equipment

a. drag line

b. clam-shell

c. backhoe

2. Waste inventory - segregation

3. Handling

a. safety

b. drums

c. over-pack drums

c. Dredging

1. Hydraulic

a. cutter head

b. mud cat

2. Pneumatic dredge

3. Mechanical dredge

d. In-site treatment

1. Solidification

2. Encapsulation

3. Neutralization

4. Microbial degradation

e. Incineration

1. Applicability

2. Mobile units

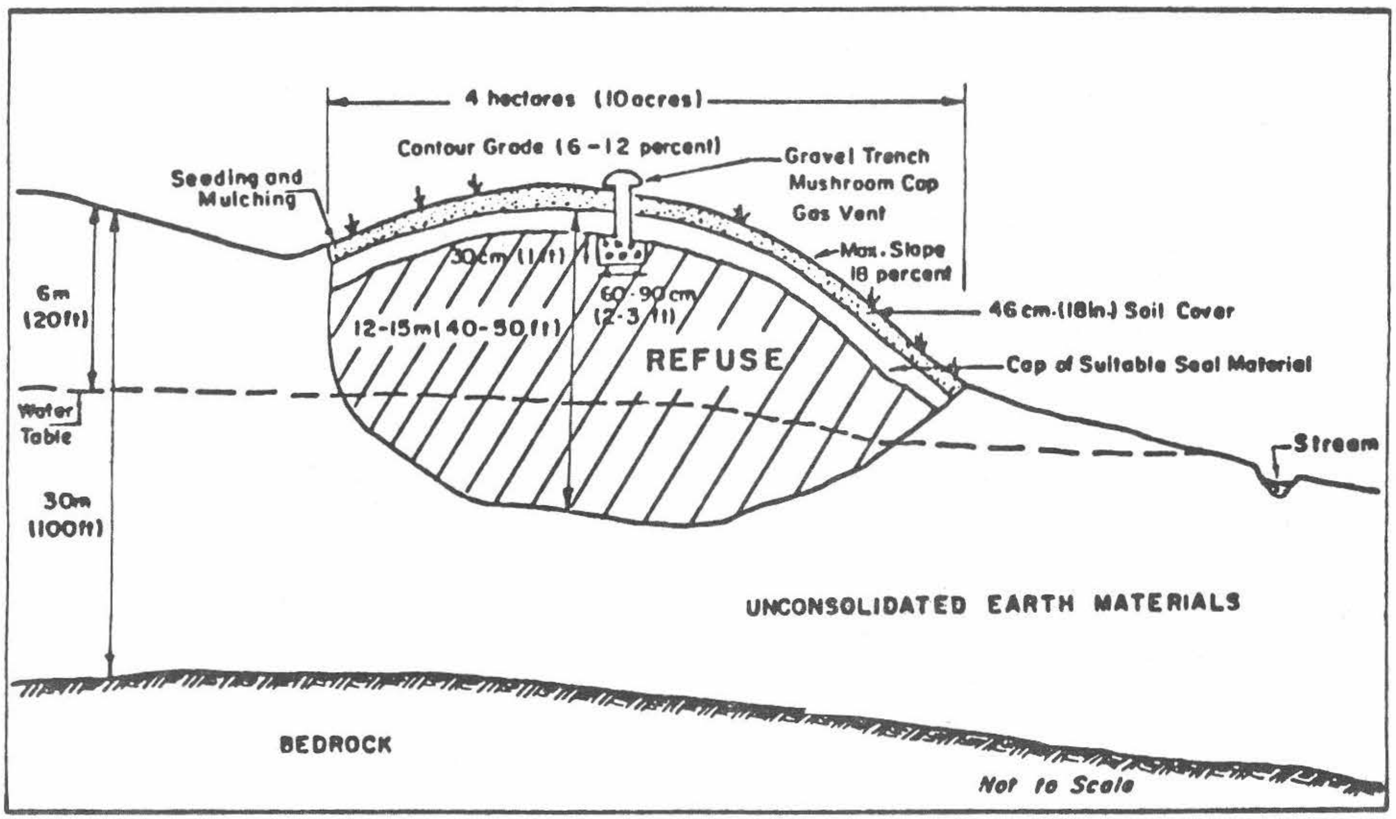
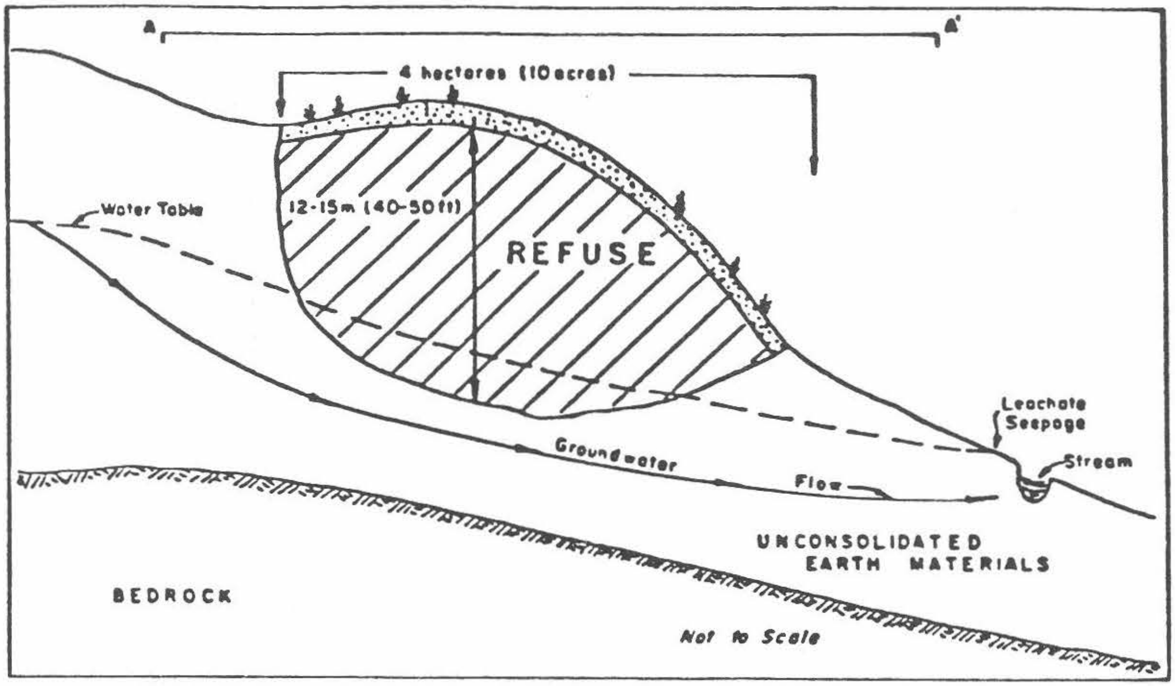
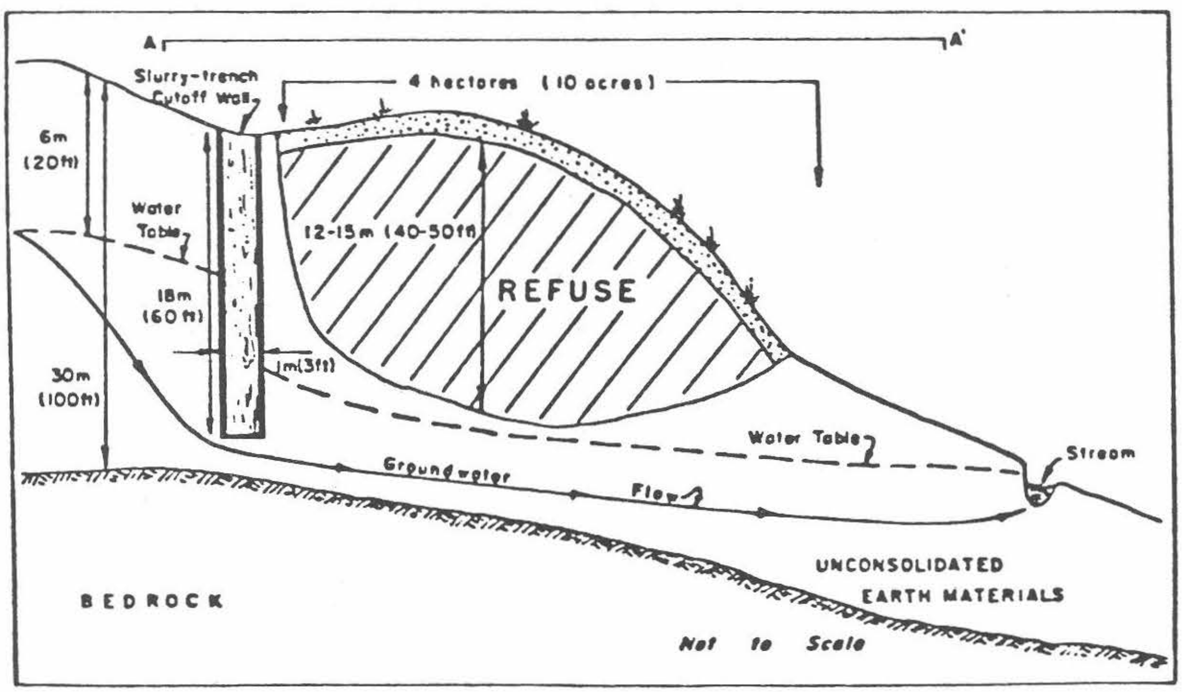


Figure 1. Cross section of capped landfill.

CUT-OFF WALL EFFECTS ON THE WATER TABLE



(a)



(b)

Figure 2. Cross section of landfill before (a) and after (b) slurry-trench cutoff wall installation.

EPA-600/2-78-142

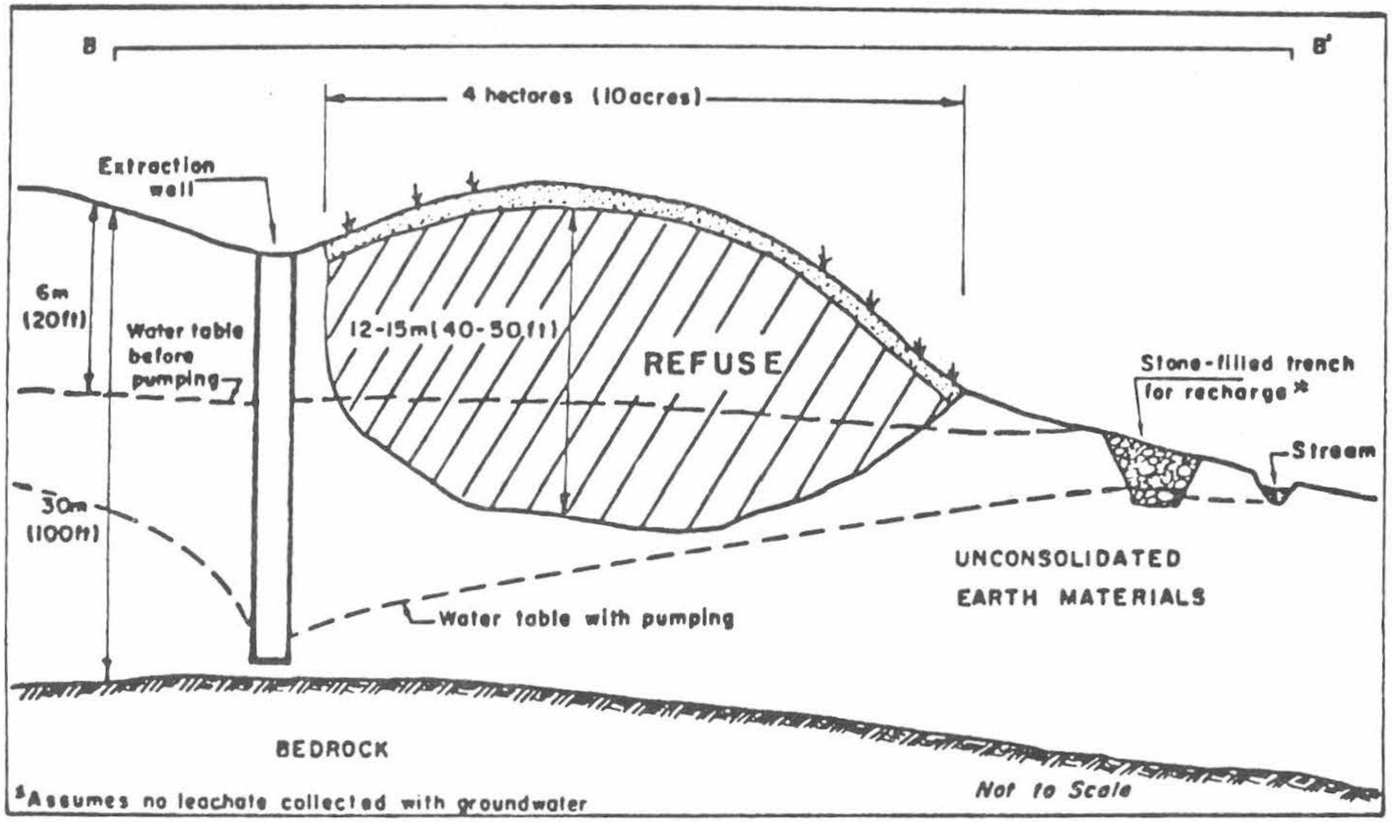


Figure 3. Cross section of extraction well at landfill.

PREVENT UNDERLYING AQUIFER CONTAMINATION

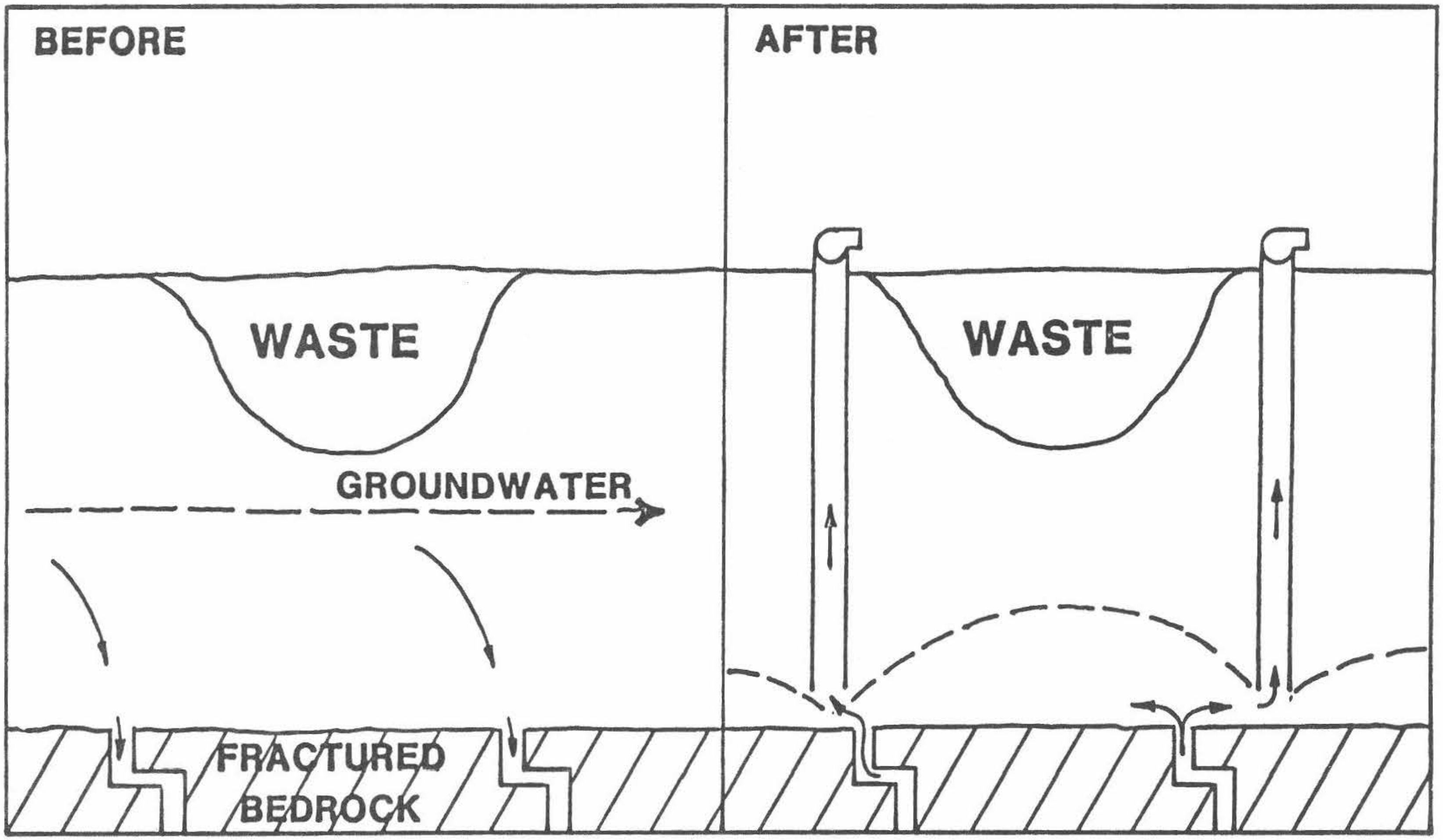


Figure 4.

CONTAIN PLUME

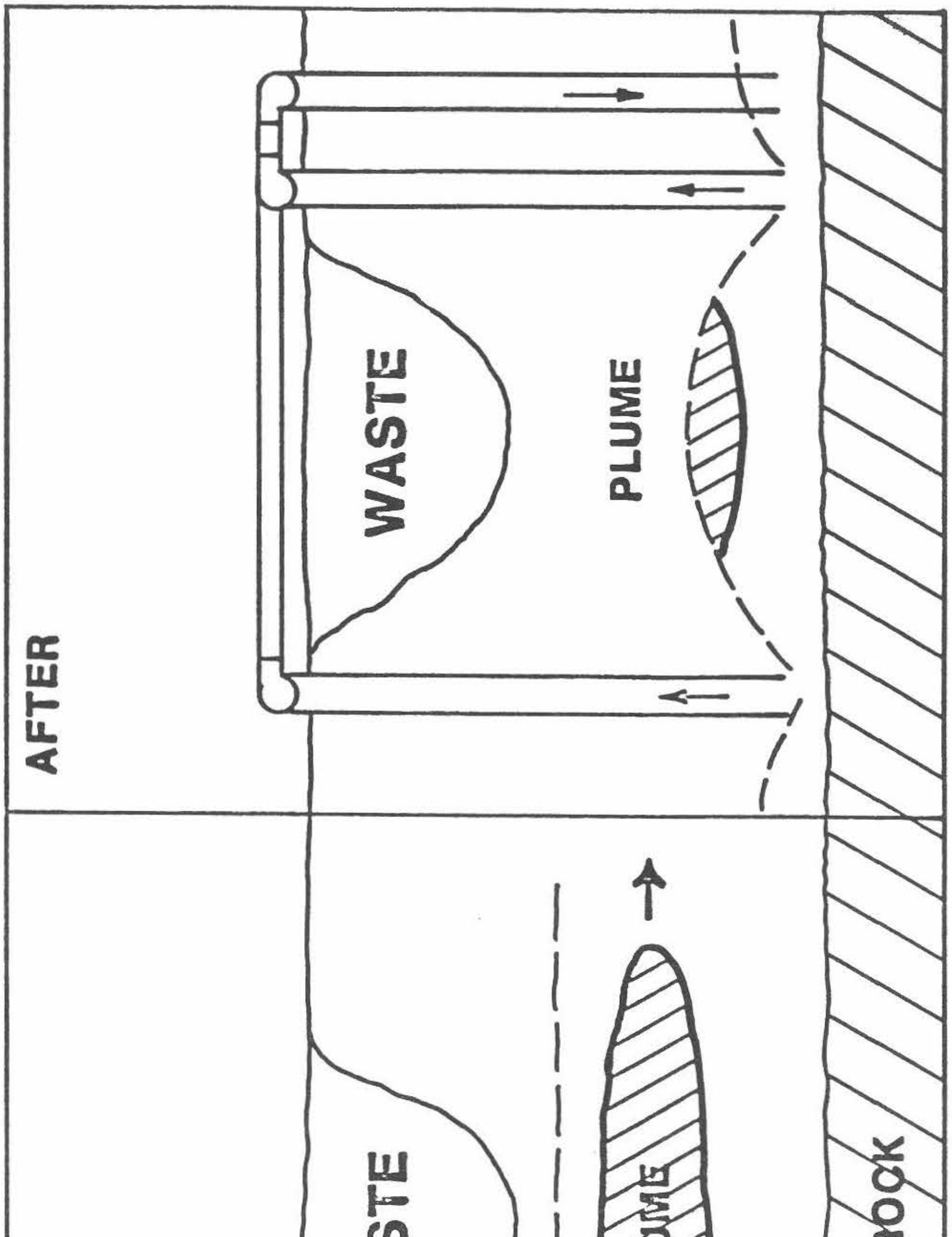
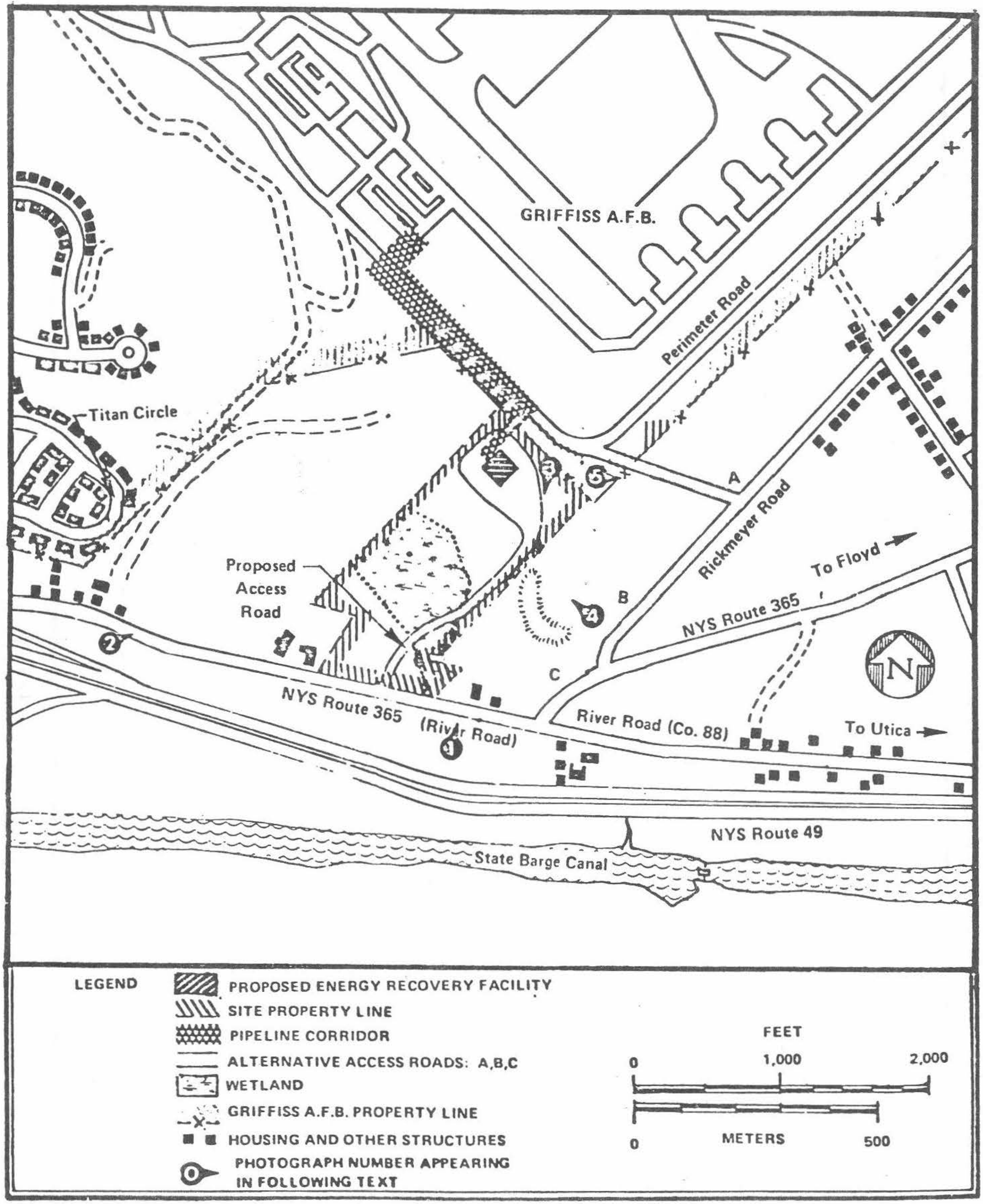


Figure 5.

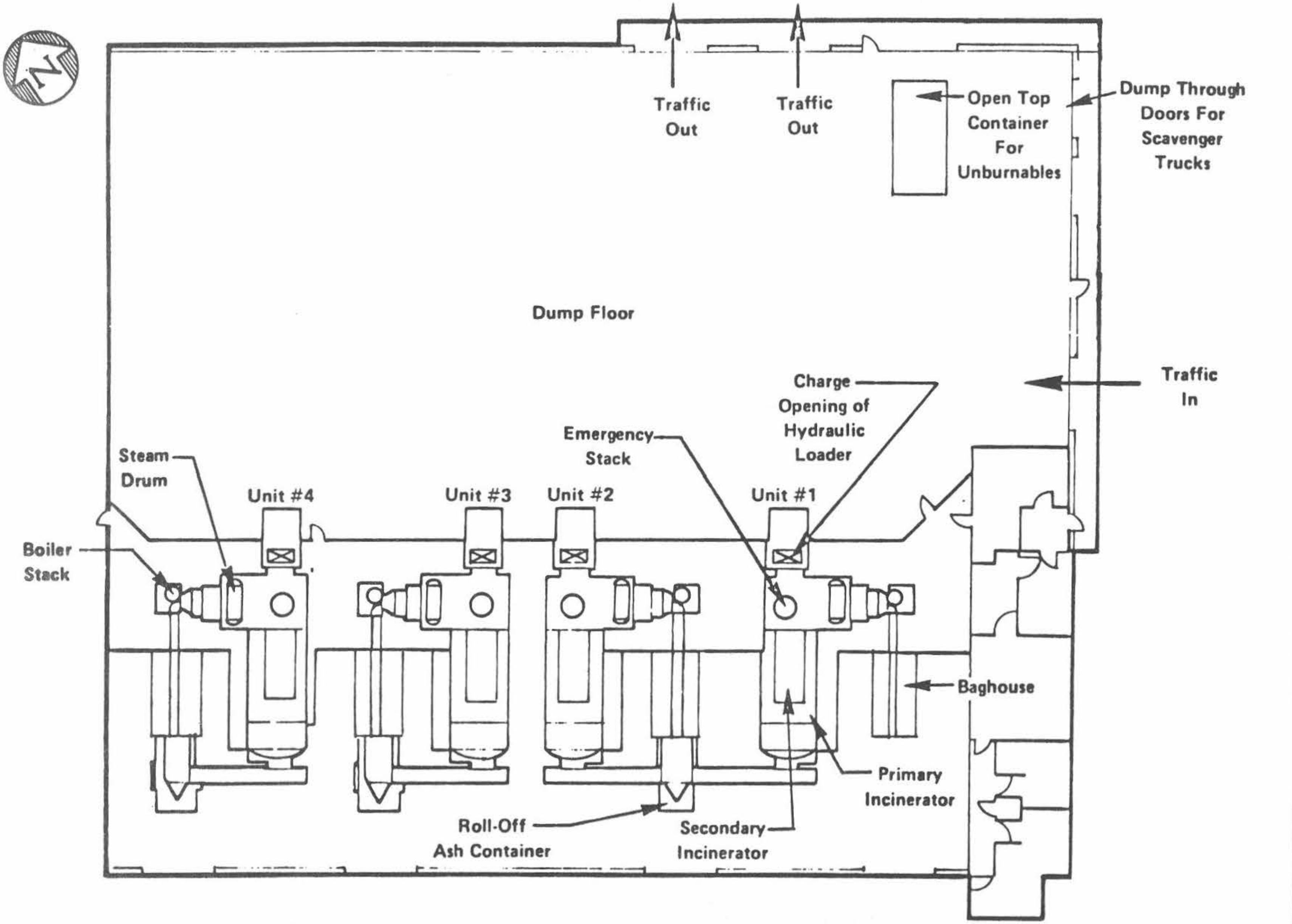
Appendix B

Energy Recovery Facility



SOURCE: BATTELLE, COMPILED FROM OCSWA, 1981; BARTON AND LOGUIDICE, 1981a; 1981b.

FIGURE 1. SITE LOCATION MAP



SOURCE: BARTON AND LOGUIDICE, 1981b; 1982a.

PLAN VIEW OF PROPOSED BUILDING AND EQUIPMENT

Seismicity in the Central Adirondacks with emphasis on the Goodnow, October 7, 1984 Epicentral Zone and its Geology

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Palisades, New York 10964

Seismicity and deformation world wide are concentrated along plate boundaries, as the theory of plate tectonics predicts. Tectonic activity is also present in the interior of plates. No theory, however, can yet successfully predict the occurrence of this intraplate tectonism. One of the manifestations of intraplate tectonism is intraplate seismicity with the accompanying consequence of earthquake hazard. The potential damage from large earthquakes is of concern in the eastern U.S. because very large earthquakes are known from the pre-instrumental historic record. The concern stems not only from the potential destruction from a repeat of these events at the same location, but also from the possibility that similar earthquakes may occur elsewhere in the eastern U.S.

INTRODUCTION

Earthquake hazard in the eastern U.S. has been estimated by relying heavily on the assumption that the seismicity detected during historic time is representative of future occurrence. Recent studies, however, on the 1886 Charleston, S.C. earthquake raise doubts on the validity of this assumption. Some of the results make it possible, if not more likely, that future large damaging earthquakes may occur at locations other than the epicentral zones of the large historic earthquakes (e.g., Hayes and Gori, 1983). This uncertainty about one of the fundamental premises for earthquake hazard analysis is symptomatic of the poor understanding we have of intraplate seismicity and tectonics in general. Clearly, a reliable estimate of earthquake hazard in the eastern U.S. and in other intraplate regions, depends on an improved understanding of the tectonic processes active in these regions.

The earthquakes themselves are one of the most important sources of information on intraplate tectonics. During the past 15 years the Lamont-Doherty Geological Observatory of Columbia University as been monitoring seismicity in New York and surrounding states from a telemetered seismic network and from temporary networks deployed in aftershock zones. It has been clear from the beginning that a key to an improved understanding of intraplate neotectonics was to combine information on seismicity, crustal structure, and stress (e.g., Sbar and Sykes, 1973; Aggarwal and Sykes, 1978; Yang and Aggarwal, 1981). The fault movements manifested by the seismicity must be the consequence of a stress field acting on a geologic environment characterized by fault zones that would tend to distort the stress field and cause stress concentrations.

[1] also at Department of Geological Sciences, Columbia University

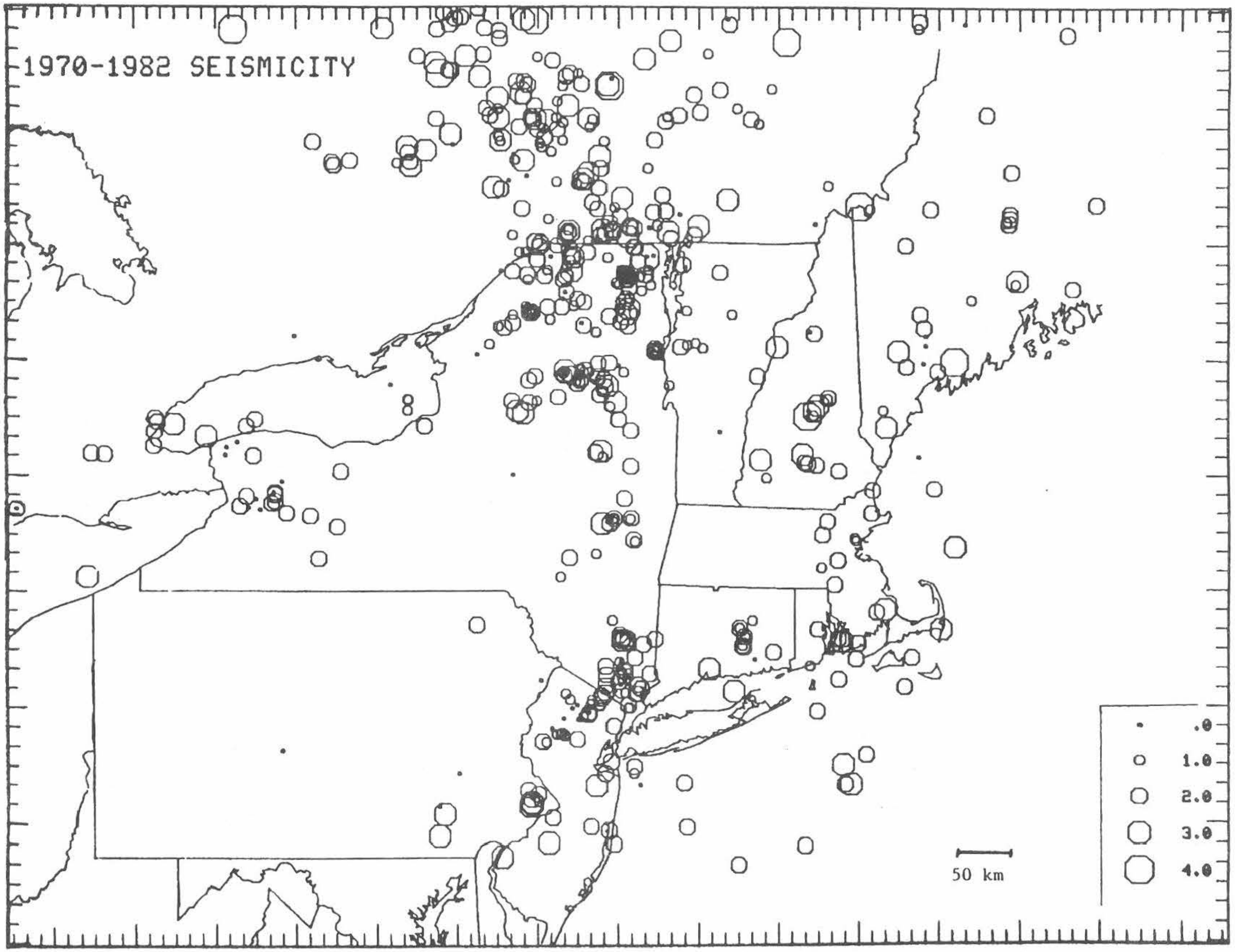

Figure 1. Twelve years of earthquake data from the New York-New Jersey Seismic Network.

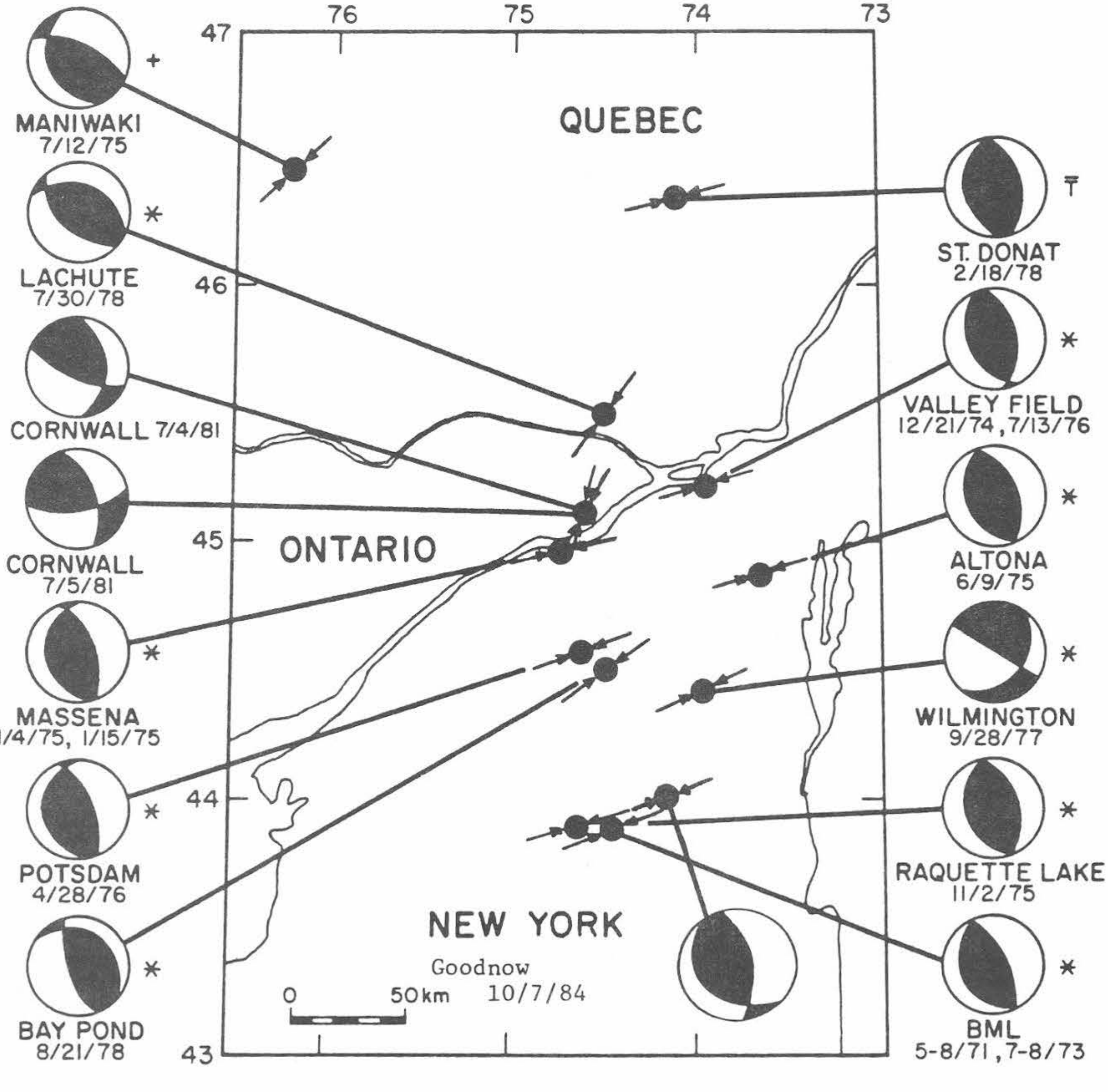

Figure 2. Fault-plane solutions from the Adirondack-Ontario seismic zone. Reverse faulting predominates and the P axes are predominantly ENE. The fault-plane solution of the Goodnow earthquake is shown at bottom center and fits this pattern.

The Goodnow, 7 October, 1983 magnitude 5.1 (M_S) earthquake in the Central Adirondacks offered a new opportunity to study an earthquake in one of the prominent areas of seismicity in the eastern U.S. (Seeber et. al. 1984; Suarez et. al., 1984; Figure 1). A previous earthquake of similar size in the same area occurred in the northern Adirondacks near Messina, N.Y. in 1944 (M_S=5.6). The study of the Goodnow earthquake has been directed at three main goals: 1) constraining the source parameters of the main shock primarily from body and surface waves recorded teleseismically; 2) resolving the characteristics of the aftershock sequence and other related seismicity from data of both the fixed stations of the telemetered network and from data of the portable network; and 3) understanding the relationship between seismicity and structural features that characterize the Grenville basement in the seismogenic zone. This last item will be the primary concern here.

MAIN SHOCK-AFTERSHOCK SEQUENCE

Main Shock-Focal Parameters

The characteristics of the Goodnow, 7 October 1983 main shock (Suarez, et al., 1984) are: moment is $M_O = 2.5 \times 10^{23}$ dyne cm (from long period Raleigh waves); source radius is $r \approx 0.5$ to 1.2 km (from body wave modeling and assuming a circular rupture with a rupture velocity = 0.9 V_S); static stress drop is then $\Delta\sigma$ = 870 bars to 120 bars, respectively (assuming a Brune's model); the average fault displacement is w = 80 cm to 14 cm (assuming rigidity $\mu = 4 \times 10^{11}$ dynes/cm^2); focal depth is h = 7.5 $\pm$ 0.5 km; rupture plane strikes N to N 15° west. The fault plane solution of the Goodnow main shock is well constrained and is consistent with an E to ENE P axis, in remarkable agreement with previous fault plane solutions in the Adirondacks (Figure 2).

Main Shock - Intensity

The intensity data from the Goodnow main shock are still being compiled. Near field data was collected locally shortly after the event. These data are scanty and non-uniformly distributed, reflecting the population distribution in this remote area. Nevertheless some interesting patterns emerge (Figure 3). An area of maximum intensity can be recognized within the valley containing Catlin Lake and the epicenter. Masonry structures tended to be slightly damaged (cracked walls and broken chimneys) and several landslides were reported in this area. These are indicators of Modified Mercalli intensity VII. At a distance somewhat less than the hypocentral depth of the mainshock ($\approx$ 8 km) from the epicenter the intensity level drops off significantly. The 1944 Messina earthquake destroyed more than a thousand chimneys in that town. Considering that the majority of the chimneys in the Goodnow meizoseismal area were damaged, it is possible that considerable damage could have been caused by the Goodnow event, had it occurred beneath a large town.

Aftershock Distribution

Within 24 hours of the main shock a number of portable seismographs were established in the Goodnow epicentral area. A portable network

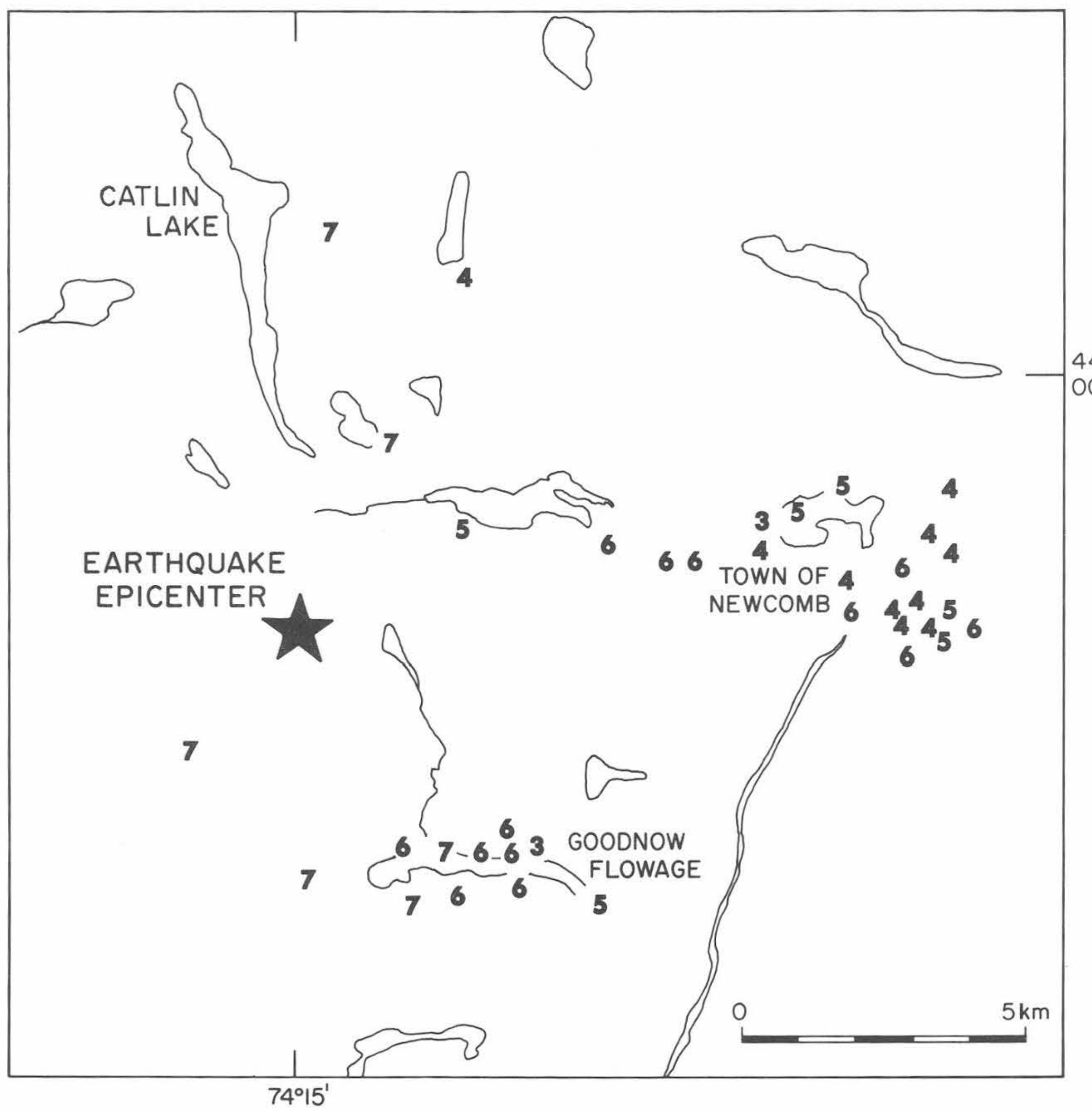

Figure 3. Epicentral intensity map of the Goodnow, Oct. 7, 1983 main shock. Numbers are Modified Mercalli intensities. Intensity VII reports are confined to a narrow zone elongated along the NNW trending Catlin Lake lineament. This orientation is parallel to the fault plane defined by the largest cluster of aftershocks and consistent with the steep, west-dipping nodal plane of the fault-plane solution for the main shock.

remained operative for 22 days after the main shock when three additional stations of the permanent network became operative (Figure 4). The data from this initial period yielded 93 accurate hypocenters (Figures 4-6) from which we can resolve 1) a planar clustering striking ≈ N 15° west and dipping 60° west, an orientation close to that of one of the planes in the fault plane solution of the main shock; 2) the geometry of this clustering, which appears to be annular when viewed normal to this plane with hypocenters concentrated at the border of a circle with a radius $r \approx 0.75$ km, in excellent agreement with constraints on the rupture of the main shock (Figure 7); 3) first motion data from the aftershocks, most of which are consistent with the fault plane solution of the main shock; 4) very few aftershocks outside of this annular cluster. The aftershock zone grows slightly in time, but only away from the rupture plane, not along this plane. About 10 days after the main shock activity seems to gradually migrate over four days about 1 km up and to the west, off the inferred rupture plane, possibly along a complimentary fault (Figure 8), which would dip shallowly eastward and nearly coincide with the downward extrapolation of the Blue Mtn. Lake fault active in the 1971-73 swarms (e.g., Yang and Aggarwal, 1981; Figure 9).

SEISMICITY AND GEOLOGY

Seismicity and Brittle Structures

The Adirondack massif is characterized by prominent sets of linear topographic features. The most prominent set strikes NNE and includes long linear valleys such as the Long Lake valley and the upper Hudson valley (Figure 4). Some of these appear to be fault controlled (Isachsen and McKendree, 1977). Another set of linears strikes WNW, this set includes the Raquette Lake lineament which has been tentatively associated with the seismogenic fault responsible for the 1975 sequence in that area (Yang and Aggarwal, 1981). Another linear in this set recognized by Isachsen and McKendree (1977) is the Catlin Lake lineament (Figure 4). This lineament is close to the surface extrapolation of the inferred rupture plane and may be controlled by the same steeply dipping fault. No prominent linear feature seems to be associated with the shallow-dipping Blue Mt. Lake fault (Yang and Aggarwal, 1981).

Plumb and others (1984) measured strain relaxation, rock anisotropy data, and conducted borehole fracturing experiments in this region to assess *in situ* stress. The various techniques gave internally consistent results. Bearings of maximum strain relaxation (ε_1) are generally aligned with topographic contours and often the mechanically stiff direction of borehole cores. Furthermore, ε_1 is aligned with the inferred ENE regional stress, local p-axes, Precambrian structures, and local joints. They hypothesized that this alignment of ε_1 with other structures is the result of a feedback between tectonic stress and the process of jointing during the development of local topography.

We are currently investigating brittle features in the Goodnow epicentral area (Figure 10). This work is at a very preliminary stage. We have found evidence of brittle faulting along the EW contact zone between the Quartzo-feldspatic basal gneiss in the Goodnow Mt. area and

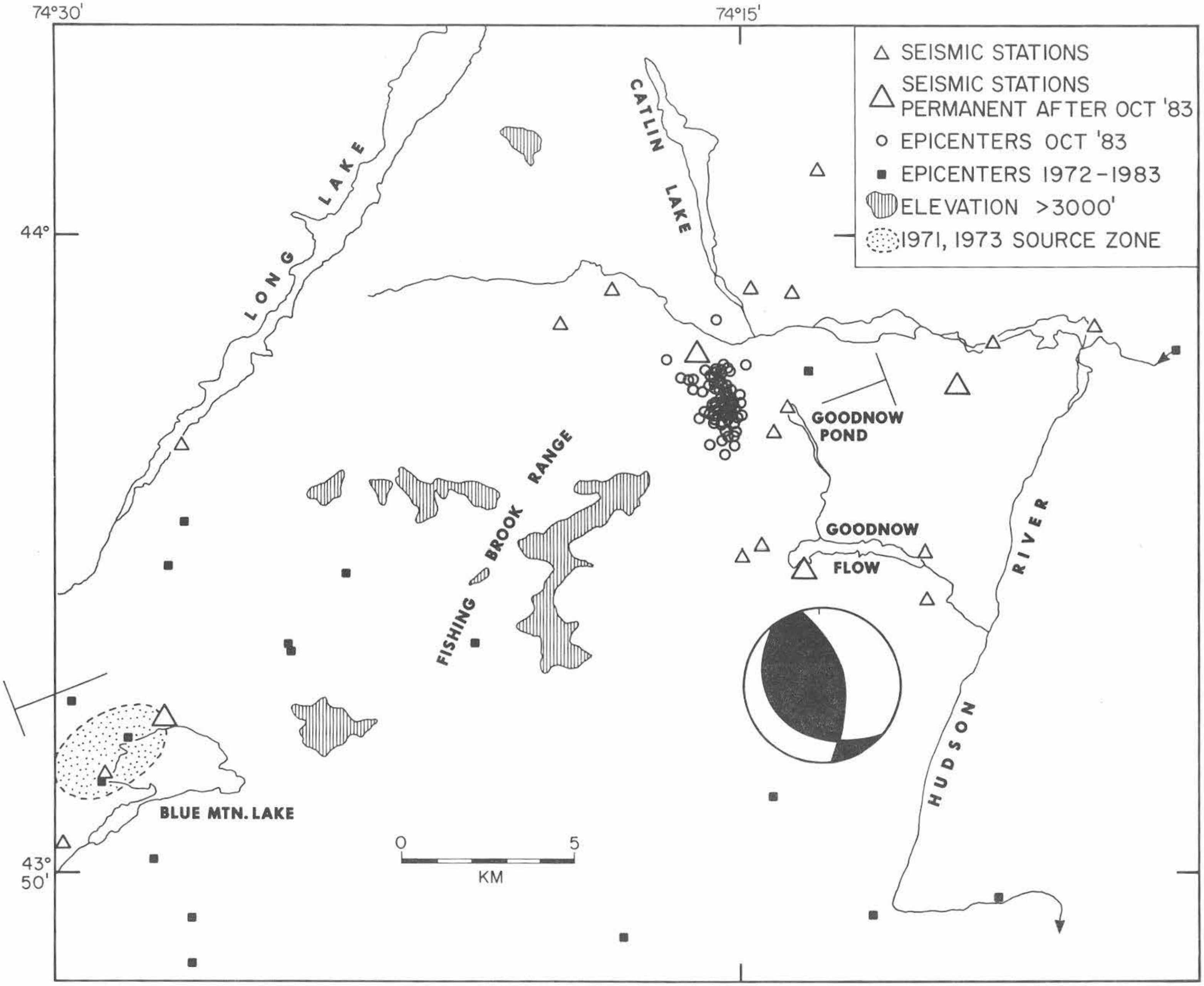

Figure 4. Blue Mt. Lake-Goodnow area of the central Adirondacks. Seismicity from 1972-83 (squares) is located by the regional seismic network and from Oct. 7-29, 1983 (circles) is located by the network of temporary stations (triangles; L-DGO and USGS). The area of the 1971 and 1973 Blue Mt. Lake swarms is also indicated (shaded). Large triangles are stations of the permanent network after Nov. 1, 1983. Catlin Lake and Long Lake are part of linear topographic features possibly associated with brittle faults (Isachsen and McKendree, 1977).

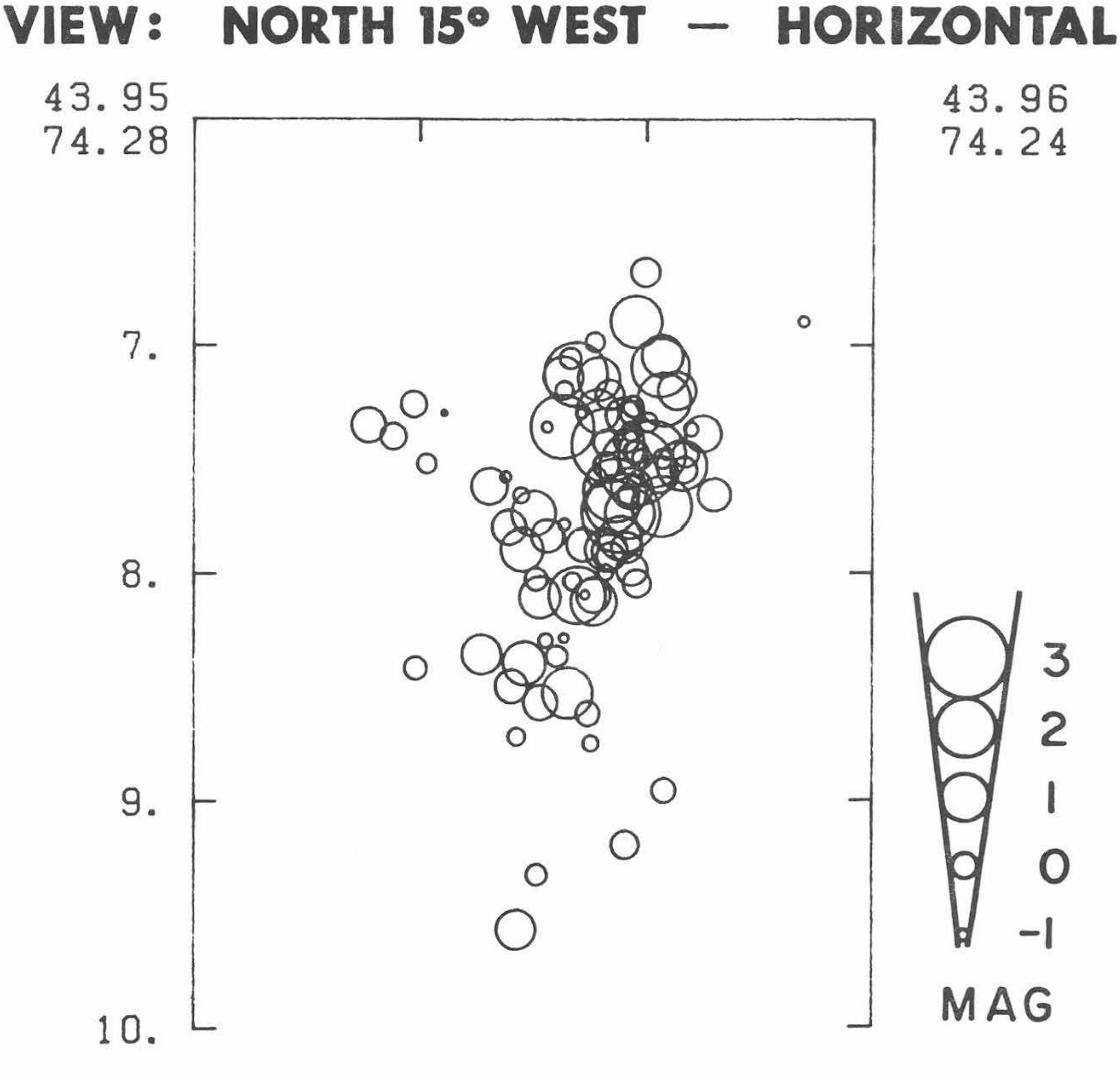

Figure 5a. North 15° west view of the Goodnow aftershocks. This section was chosen among many other sections with strikes differing from this by as litle as 5°, to yield the narrowest scatter in hypocenters about a plane, presumably the plane of the main rupture. Thus, the hypocenter data agree well with the first-motion data and indicate that the NNW-striking plane dipping steeply to the west in the fault-plane solution is the main rupture plane.

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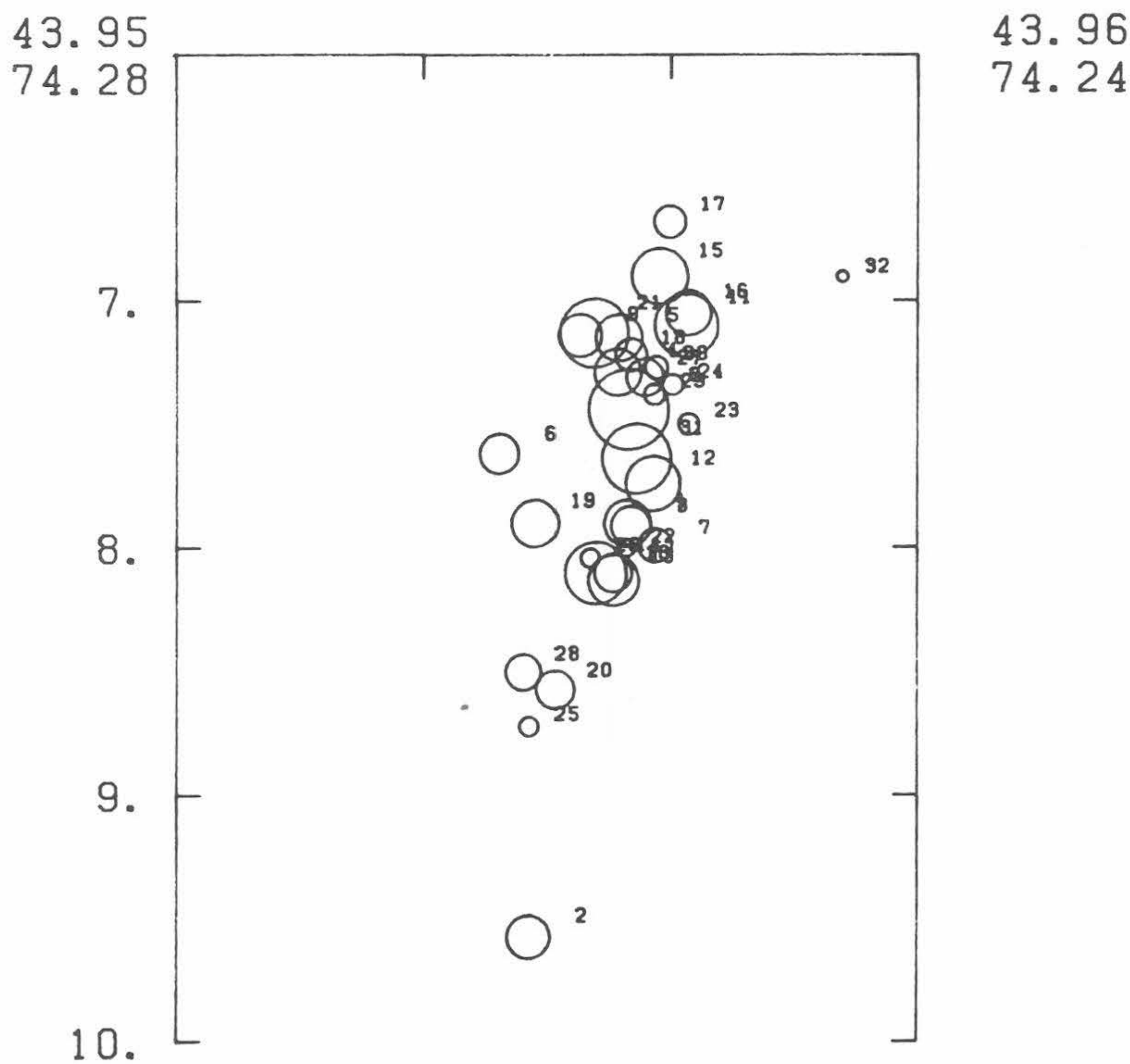

Figure 5b. First 32 aftershocks (recorded during first 5 days). Note how distribution is more planar than in **Figure 5a** and defines the rupture zone despite lower quality of locations than for later aftershocks. Thus scatter in Figure 5a is probably real.

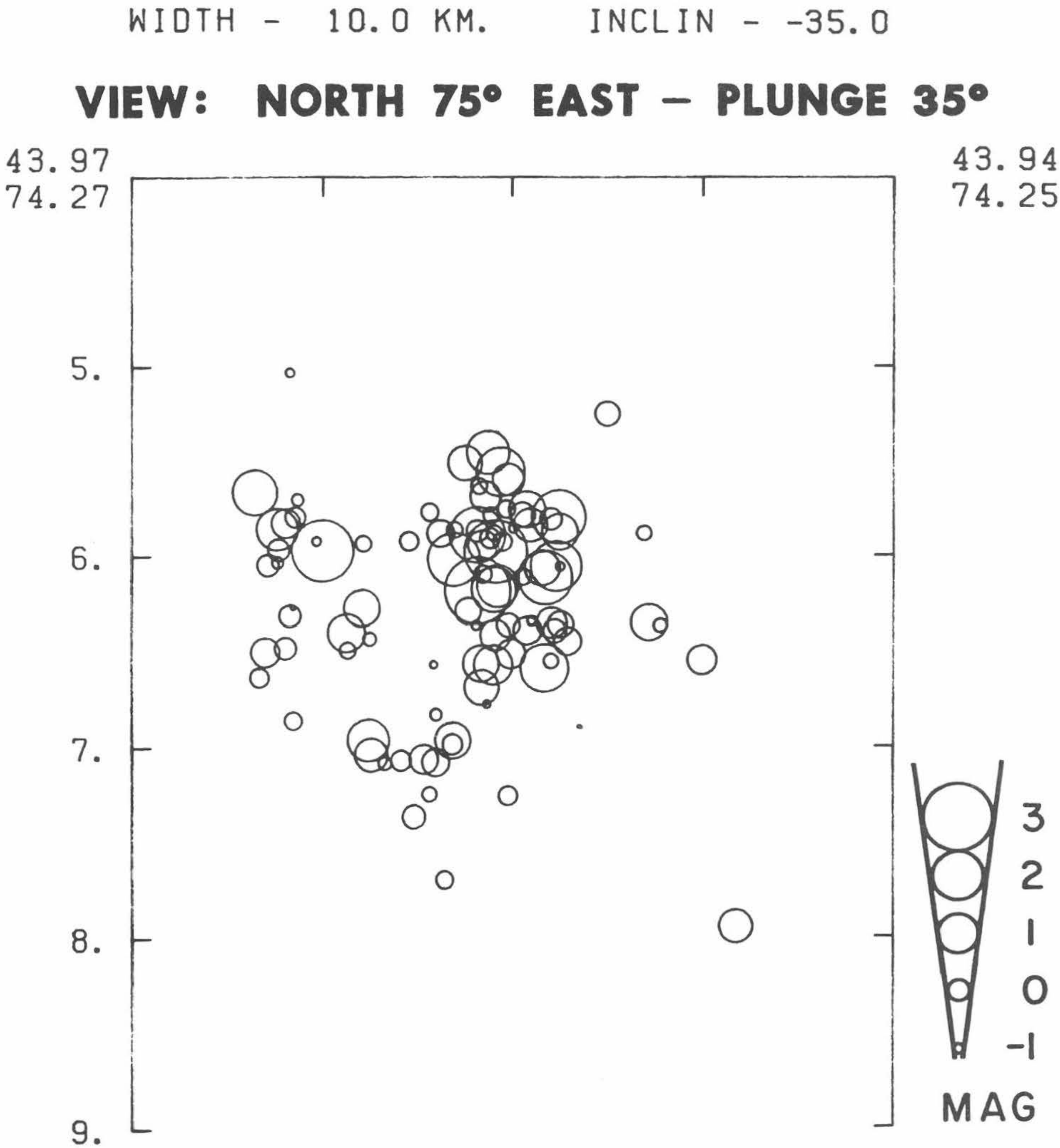

Figure 6. The Goodnow aftershock zone viewed perpendicular to the inferred plane of the main rupture (i.e., viewed north 75° east at a plunge of 35°; note that the plane of this section is not vertical and the numbers do not reflect true depth).

From the data in this figure we estimate the main rupture to be about 1.5 km in diameter and extend from about 7 to 8½ km depth.

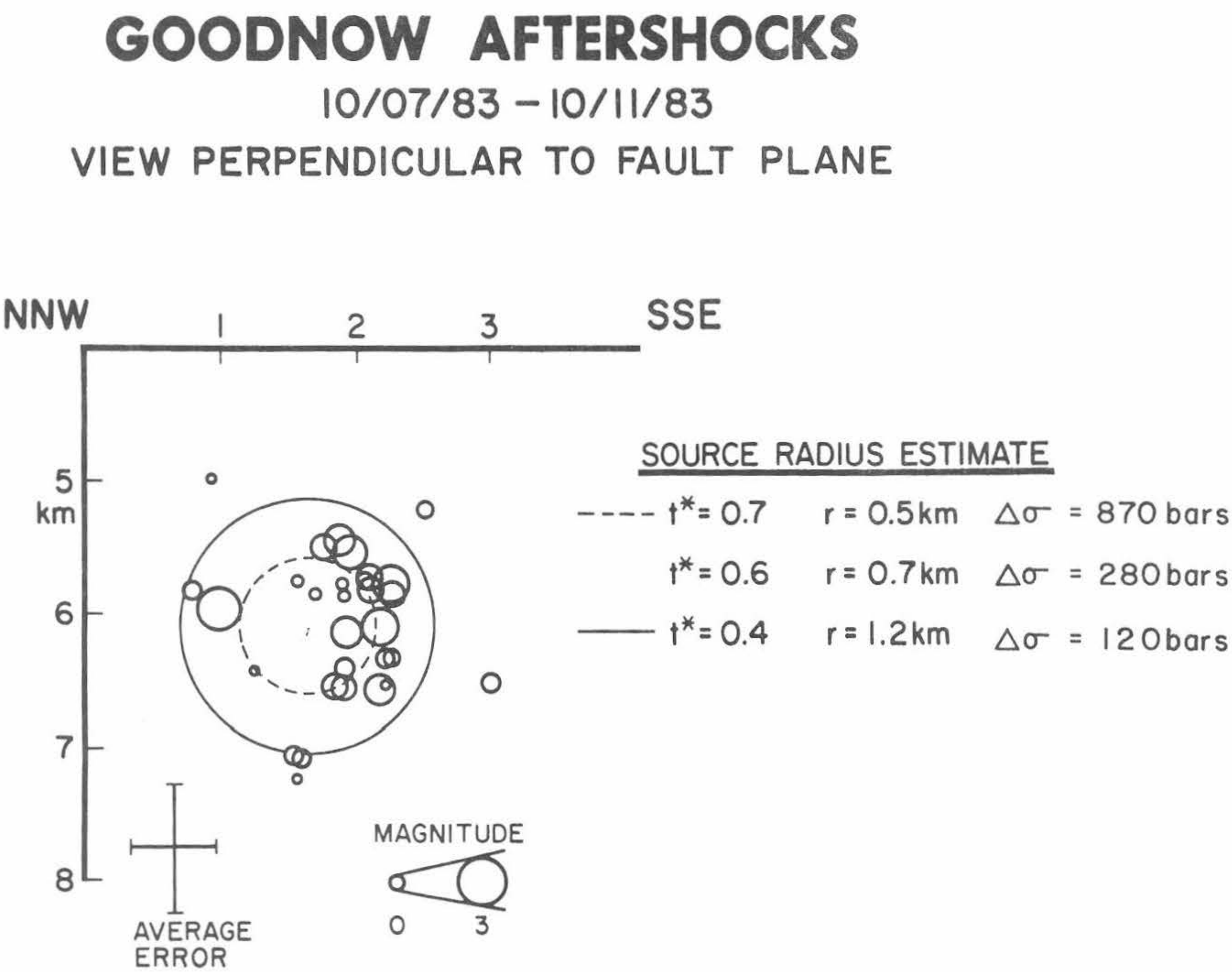

Figure 7. The first 29 well recorded aftershocks of the Goodnow earthquake viewed perpendicular to the inferred fault plane. The circles give the range in rupture dimension inferred from modeling the short period teleseismic P waves, assuming Mo = 2.5×10^{23} dyne-cm (obtained from long period Raleigh waves), a velocity of rupture = 0.9 Vs, and rupture nucleation at the center; t* is an attenuation parameter. The circles are centered at the hypocentral depth inferred from the moment tensor inversion. These data indicate that the aftershocks are confined for the most part to the rupture or near its outer edge where stress is expected to be concentrated (from Suarez, et al., in preparation).

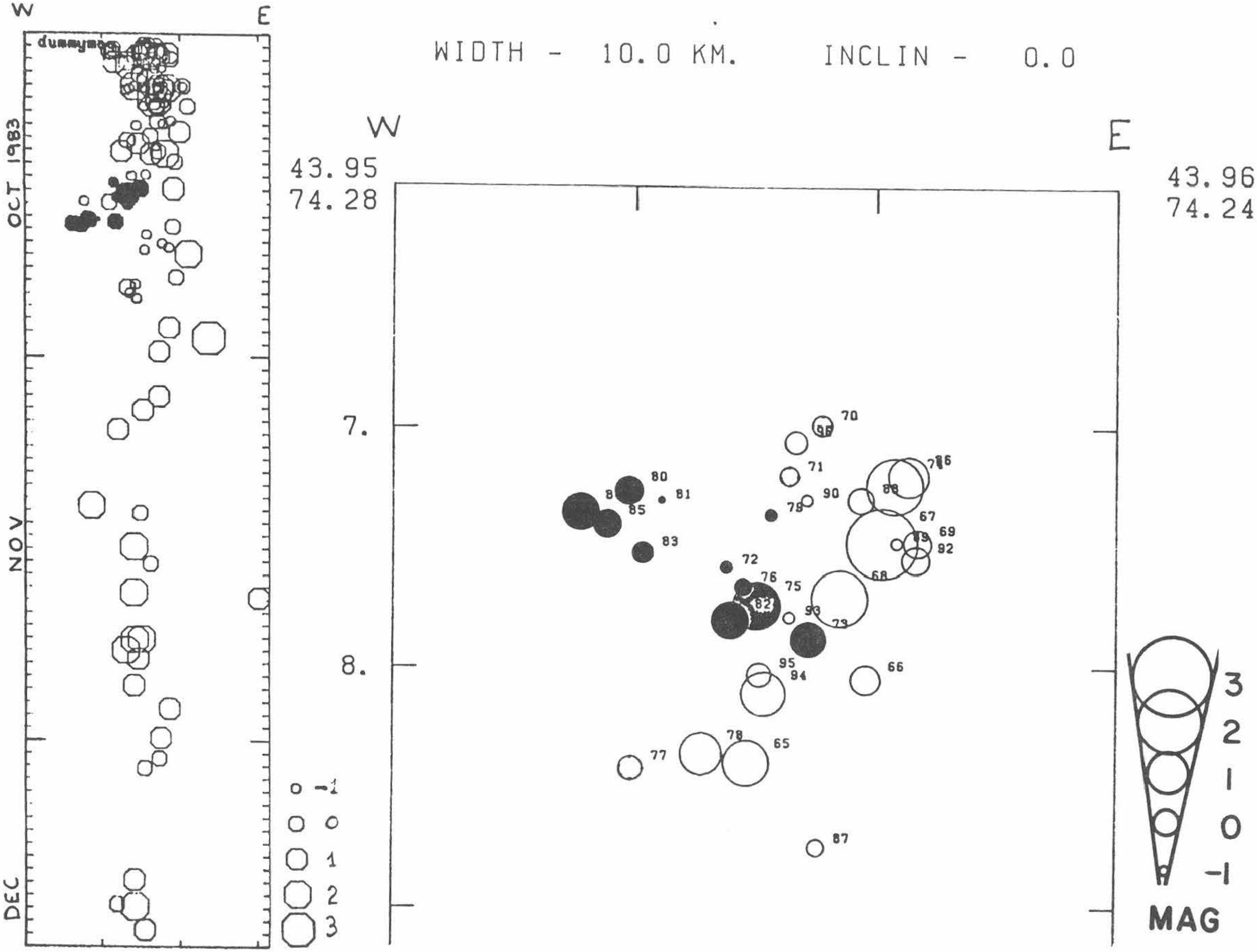

Figure 8. Time-space data suggesting the propagation of slip along a conjugate fault away from the main rupture. On the left is a time-space plot of the Goodnow epicenters projected on the line of the section in on the right (view along main upture as in Fig. 5). Most of the earthquakes in a tight westward migrating sequence occur on a plane dipping eastward and extending about 1 km from the main rupture (blackened symbols). This migration seems to take about 4 days.

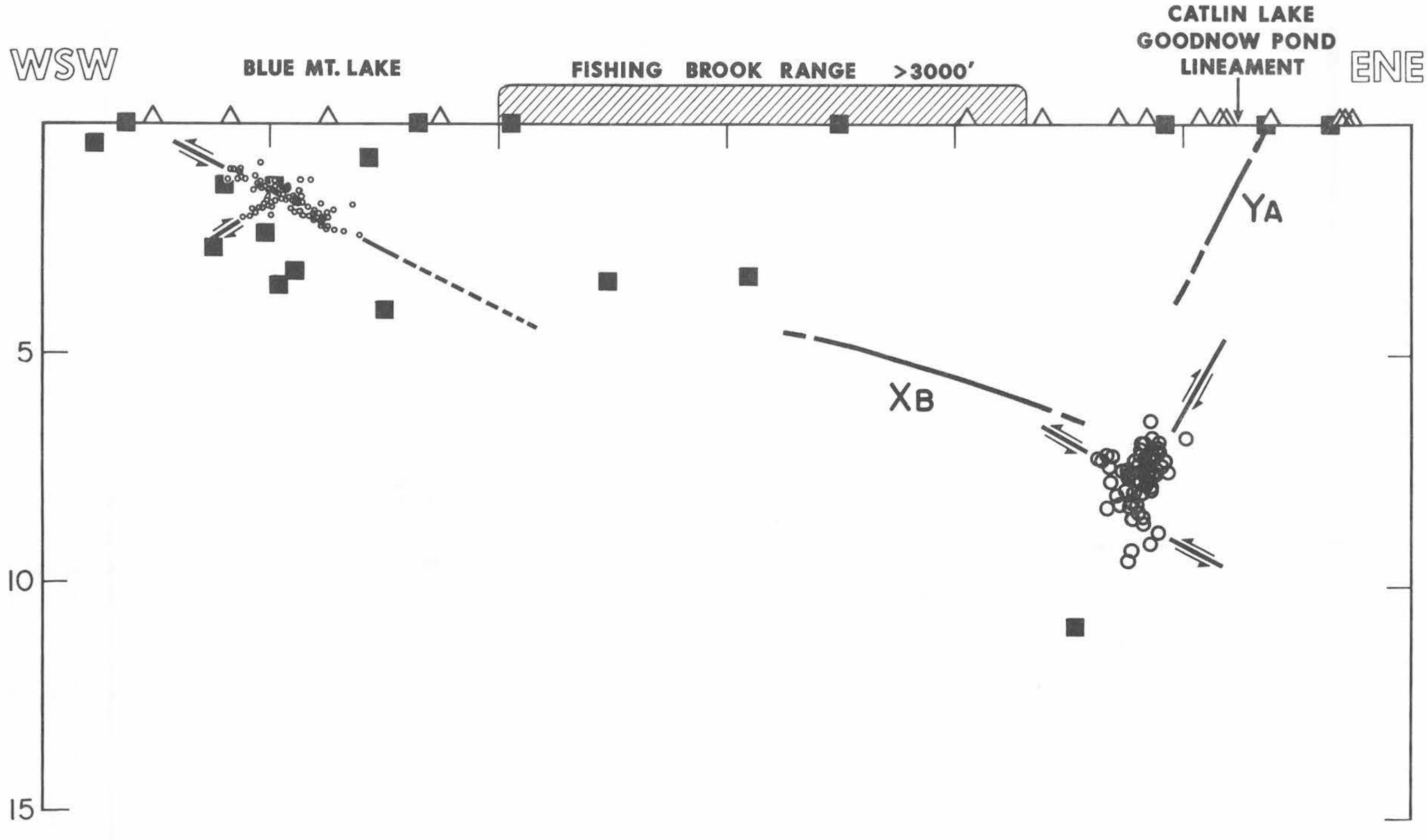

Figure 9. Section through the Goodnow and Blue Mt. Lake area (located in Fig. 4; no vertical exaggeration). Seismicity in Fig. 2 ± 20 km from plane of section is included (same symbols). Hypocenters for the 1971-73 Blue Mt. Lake swarms are from Yang and Aggarwal, 1981. Active faults delineated by spatial and temporal distribution of hypocenters and by first-motion data are indicated. Depth control for events located only by the regional network is generally poor. It is possible that seismicity after the Blue Mt. Lake swarms and before the Goodnow event (squares) was on the same system of faults active during the well-located sequences in 1971, 1983, and 1983. X_b and Y_a are possible reflectors identified on COCORP reflection data (Klemperer et al., 1984).

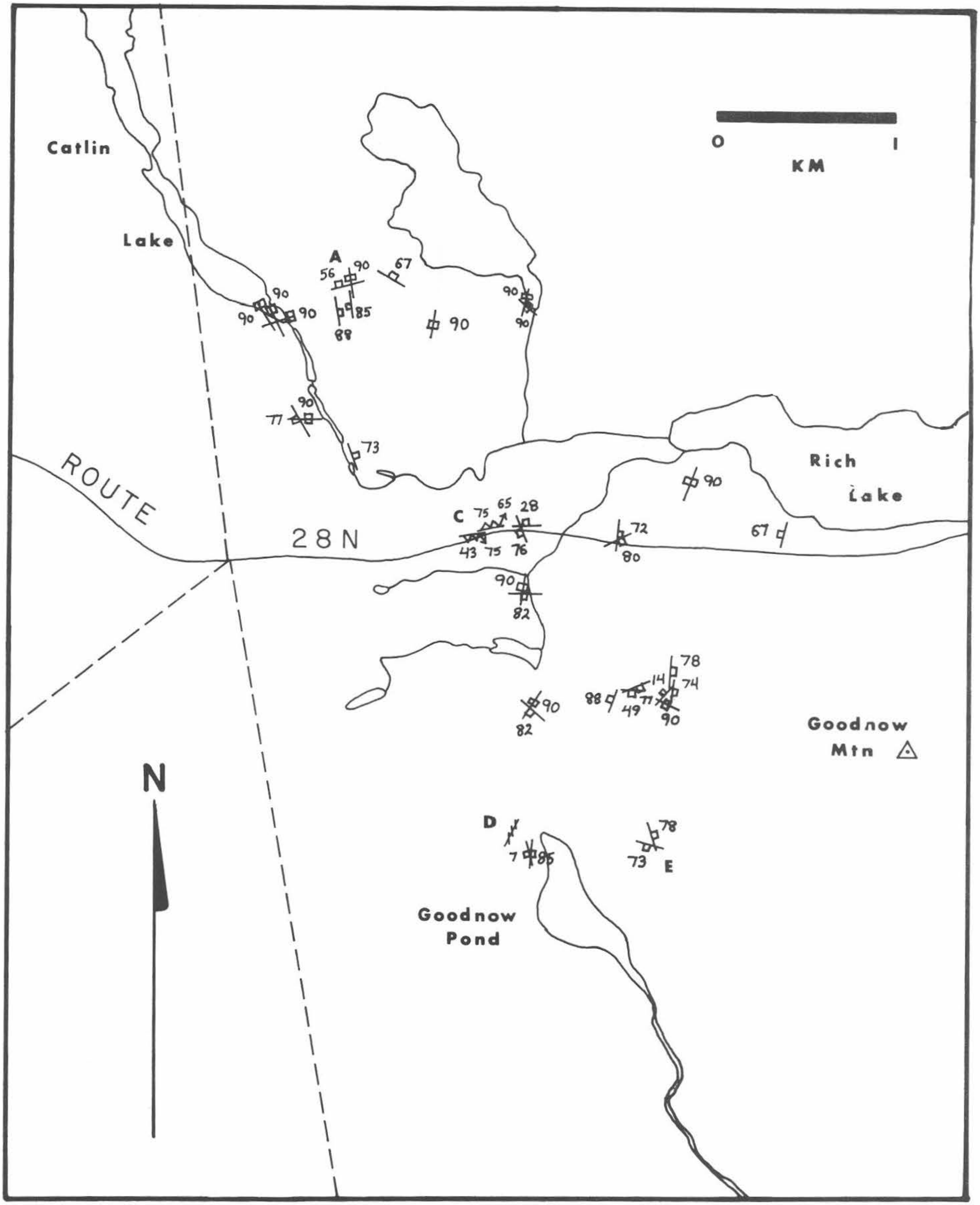

Figure 10. Joint and slickenside measurements made during a reconnaissance of the Goodnow epicentral area. Strike symbols with boxes denote joint planes; strike symbols with double barbs denote slickenside planes with rake of slickensides in plane given by arrow. Note that NNW trending joints give way to NNE trending joints to the east of the Catlin Lake-Goodnow Pond lineament. Data was mapped onto 1:25,000 scale air photos. **A.** Outcrop with N trending joint having aligned quartz crystal growth; **C.** Slickensided outcrop along Route 28N; **D.** Set of en echelon cracks; individual cracks trend 011°, crack train trends 027° (right-stepping offset); **E.** Joint oriented 163,78E contains possible gouge. Dashed lines are town boundaries.

the metasediments of the Grenville (series) to the north (McLelland and Isachsen, 1980). No evidence of faulting has yet been found along the Catlin Lake lineament, but this lineament appears to be joint controlled. The dominant joint set tends to strike NNE, parallel to the Long Lake lineament set, except in the vicinity of Catlin Lake where it strikes NNW.

Seismicity and Grenville Ductile Structure

A map of seismicity in the Adirondacks detected by the New York State seismic network during the last 10 years shows clustering of epicenters in well defined zones separated by aseismic zones (Figure 1). Two prominent zones strike approximately ENE and form broad arcs concave to the south in the western half of the Adirondack massif. When superimposed on structural (Figure 11) or lithologic (Figure 12) maps, these seismic zones appear to follow structural trends of Grenvillian age. This correlation between structure and seismicity cannot be interpreted as simple reactivation since the seismogenic faults appear to strike consistently NNW (Figure 2), at a large angle to the structural trends and seismic zones. The Goodnow earthquake is at the intersection between the ENE striking Central Adirondack seismic zone and another seismic zone striking NNW, parallel to the inferred plane of faulting, that can be traced southward to near the southern edge of the Adirondack massif.

McLelland and Isachsen (1980) have proposed a south verging Wakely Mt. nappe cored with the basal quartzo-feldspatic gneiss and covering most of the southern Adirondacks (Figure 12). The root zone of the Wakely nappe would closely follow the central Adirondack seismic zone. It is likely that a major structural boundary is associated with the central Adirondack seismic zone because 1) foliation data forms a band of subparallel and gently curving trends along this zone (Figure 11), in contrast to the adjacent aseismic regions where the foliation trends are relatively convoluted; and 2) the western half of this seismic zone corresponds to the boundary between areas of foliation that dip consistently southward to the north and northward to the south (Figure 11).

In summary, available earthquake and geologic data in the Adirondacks suggest that Grenville age structure is controlling some aspects of the seismicity. This result is in agreement with Plumb et al. (1984). This control cannot, however, be simple reactivation of Grenville structures, since seismogenic faults are at large angles to these structures. We are considering the possibility that seismicity is lithologically controlled. In this case, the relation between structure and seismicity would be a consequence of the control that structure has on lithology. The Adirondacks provide one of the best opportunities to carry out a detailed comparison between geology and seismicity because the seismogenic part of the crust is exposed and can be studied, and seismicity is relatively high and well monitored. Our current field investigation in the Goodnow - Blue Mt. Lake area of the central Adirondacks is directed at improving constraints on Grenville structures so that a reliable structural model can be developed and compared with the seismicity.

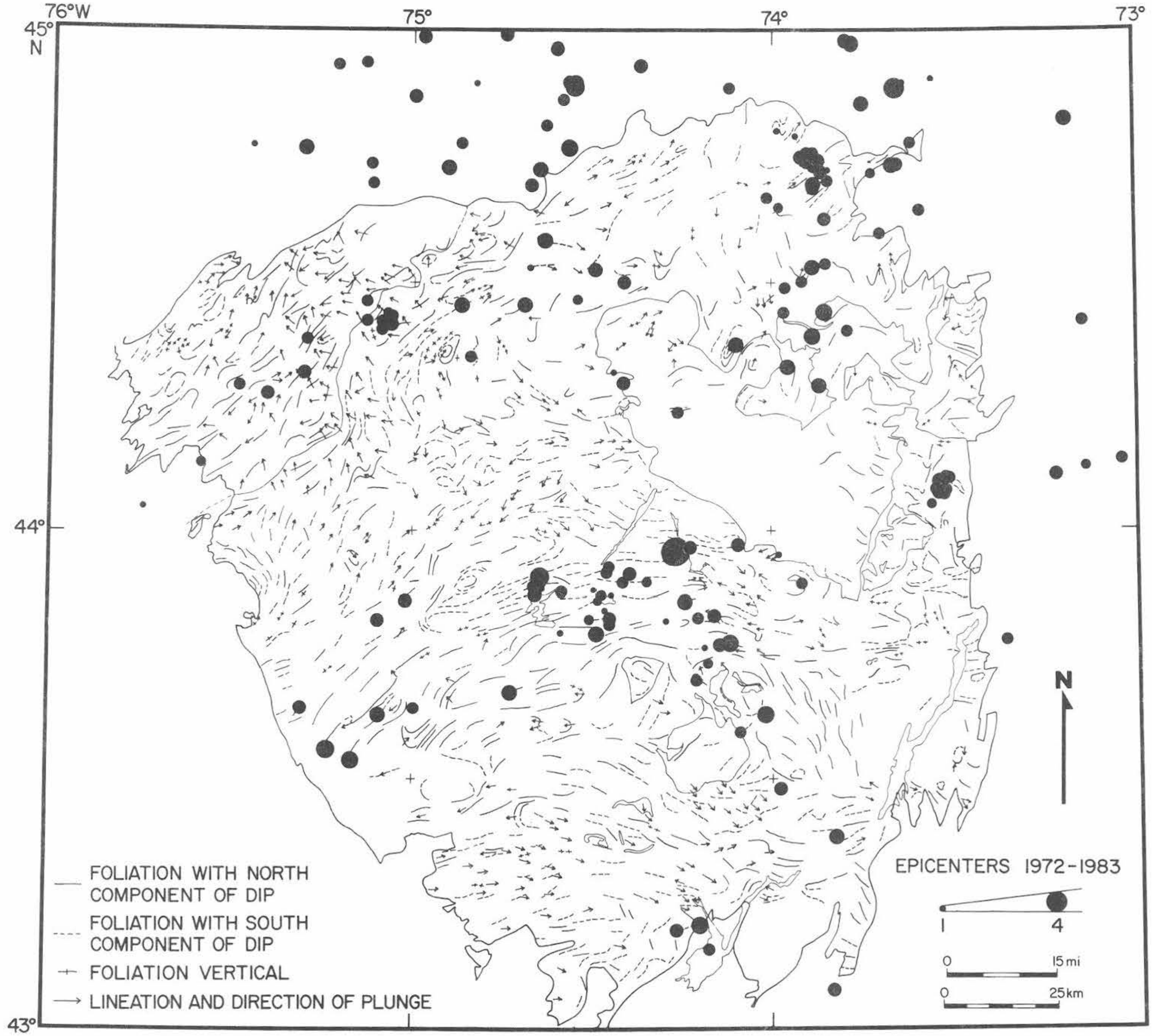

Figure 11. Epicenters from the N.Y. State Seismic Network, 1972-1983 (black dots) superimposed on foliation and lineation data extracted from the 1:250,000 N.Y. State geologic map (Fisher et al., 1970). Two arcuate belts of seismicity in the central and north-western Adirondacks are clearly related to Precambrian (Grenville) structural trends. Large domains where dips of foliation have either a north or south component can be recognized. The portion of the east-west seismic belt in the central Adirondacks which contains the Raquette Lake, Blue Mt. Lake, and Goodnow earthquakes is closely associated with the boundary between two zones with southerly and northerly dip of foliation, respectively.

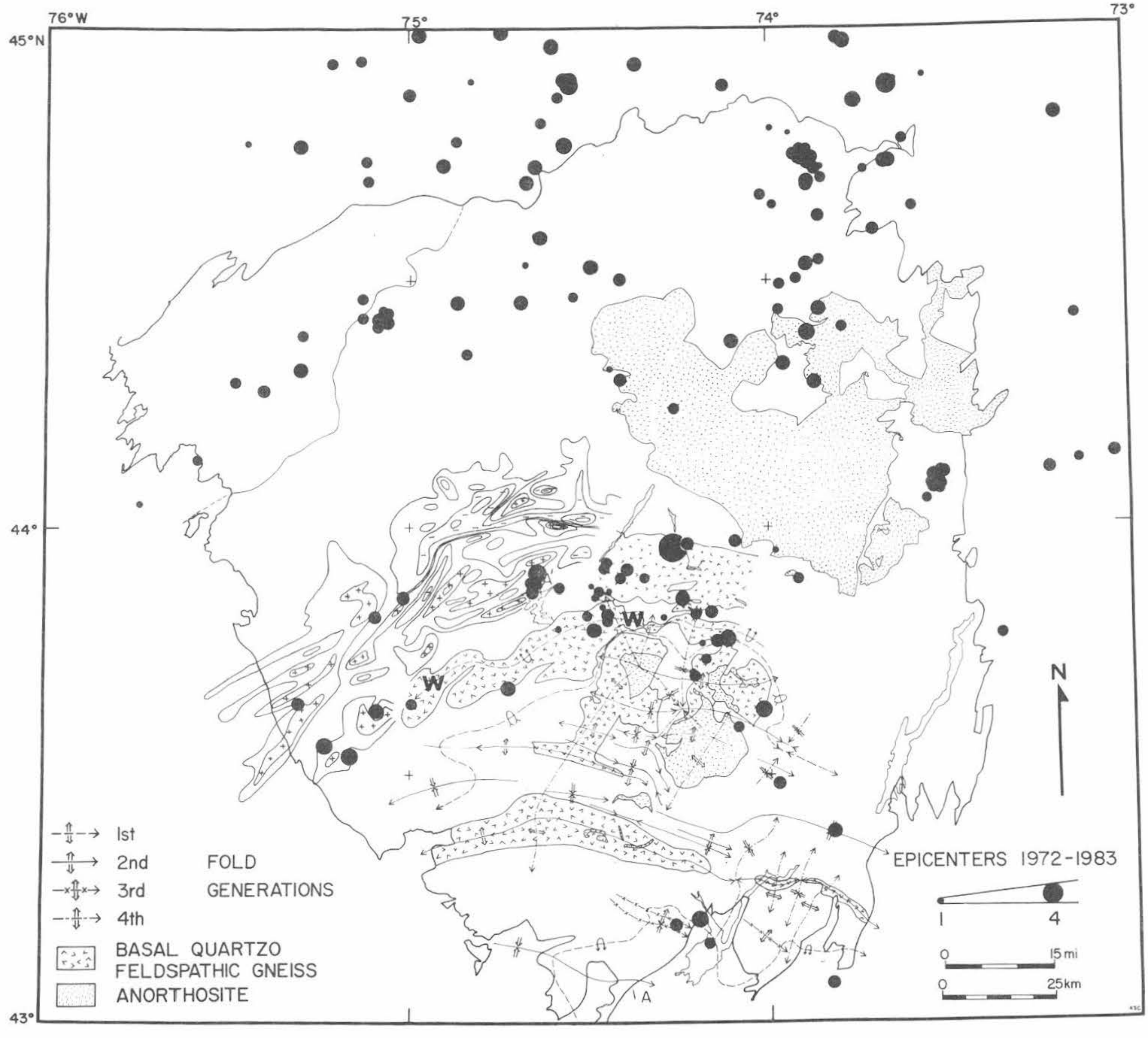

Figure 12. Epicenters from the N.Y. State Seismic Network, 1973-1983 (black dots) superimposed on a structural/lithologic map of the southern and central Adirondacks modified from McLelland and Isachsen (1980) and of magnetic data (Zietz and Gilbert, 1981) beyond the limit of McLelland and Isachsen's map (areas with '+' have magnetic intensity – 1200 gammas; areas with '–' magnetic intensity – 600 gammas). The quartzofeldspatic gneiss differentiated in this map is thought to be the basal stratigraphic unit and is tentatively interpreted as a pre-Grenville basement. Four phases of Grenville deformation have been identified from these data. W indicates trace of Wakely Mtn. nappe root zone. Note remarkable correlation between epicentral zones and arcuate Grenvillian structural trends in the central and northeastern Adirondacks.

FIELD TRIP STOPS

While field trip stops are not yet final at time of writing, likely stops include outcrops A, C, D, and E in Figure 10 and some of the Intensity VII effects (Figure 3).

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