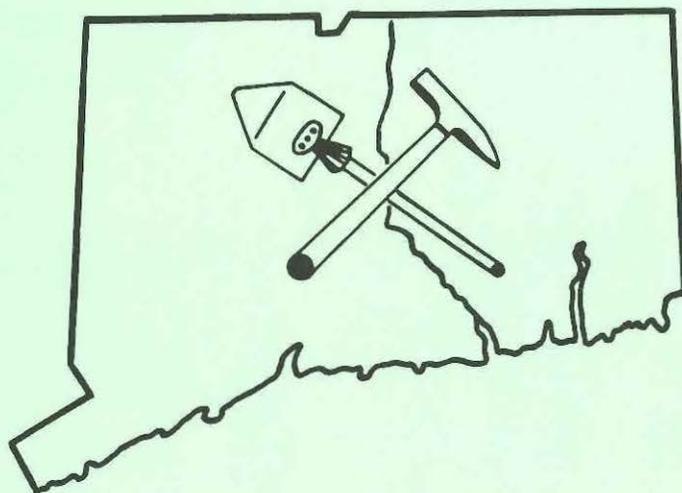


**Guidebook for Fieldtrips in Connecticut  
and Adjacent Areas of New York and Rhode Island  
New England Intercollegiate Geological Conference**

77TH ANNUAL MEETING

YALE UNIVERSITY  
NEW HAVEN, CONNECTICUT

OCTOBER 4, 5 AND 6, 1985



STATE GEOLOGICAL AND NATURAL HISTORY SURVEY  
OF CONNECTICUT

NATURAL RESOURCES CENTER

DEPARTMENT OF ENVIRONMENTAL PROTECTION

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OF NEW YORK AND RHODE ISLAND

Editor

Robert J. Tracy  
Department of Geology and Geophysics  
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New Haven, CT 06520

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Honorable William A. O'Neill, Governor of Connecticut  
Stanley J. Pac, Commissioner of the Department  
of Environmental Protection

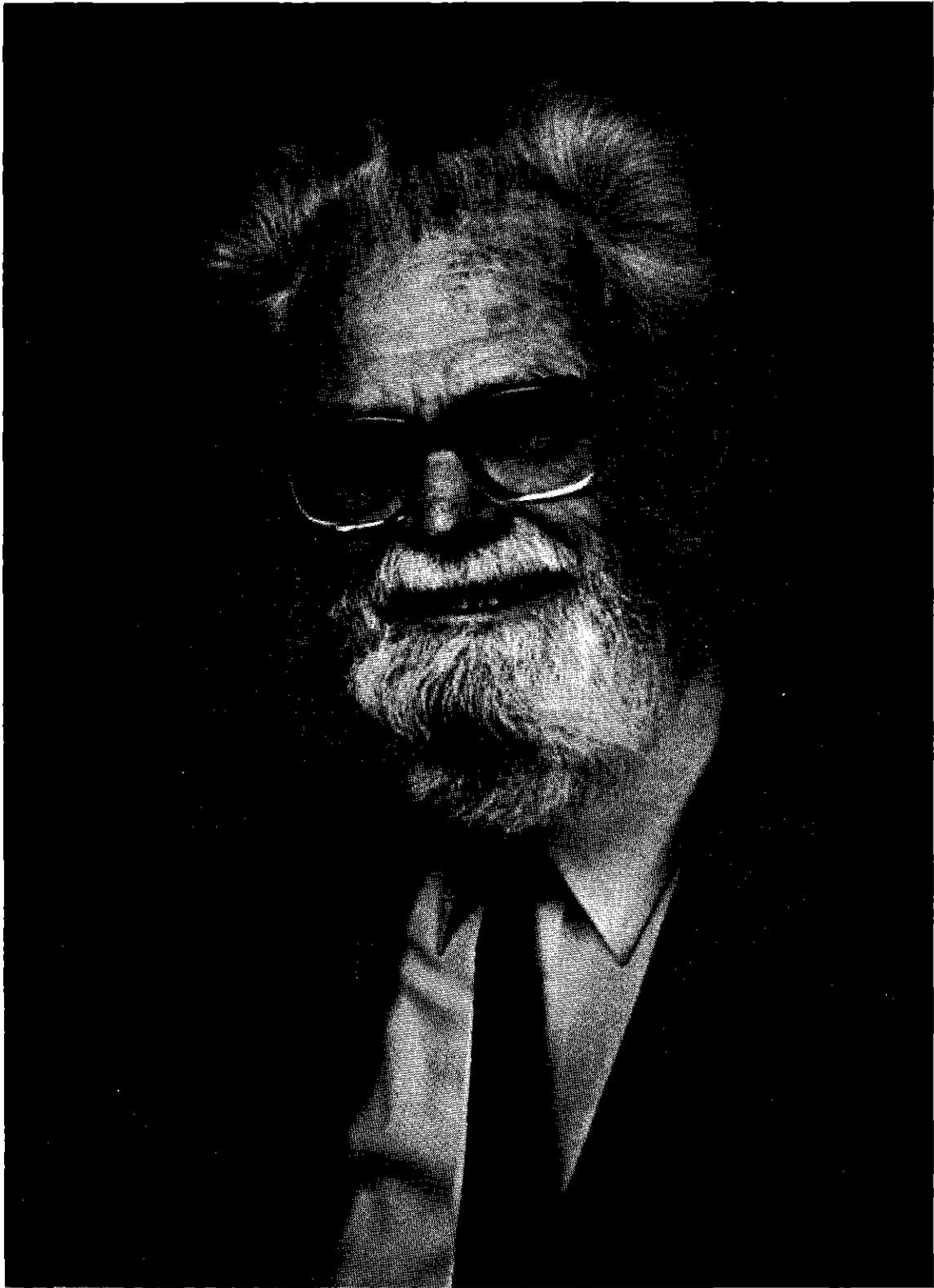
STATE GEOLOGIST  
DIRECTOR, NATURAL RESOURCES CENTER  
Hugo F. Thomas, Ph.D.

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Telephone: (203) 566-3540.

## EDITOR'S PREFACE

The arrival of the 1985 N.E.I.G.C. Meeting in New Haven, only three years after the 1982 meeting in Storrs, indicates the vitality and breadth of ongoing geologic studies in Connecticut. One man who is to a large extent responsible for the current health of geological studies in Connecticut is Professor John Rodgers of Yale. He has accomplished this through an enlightened combination of personal research and responsible criticism of others' research, coupled with his overseer's role as compiler of the very recently published Geologic Map of Connecticut. I suspect that there are few working in Connecticut geology who have not been influenced by John's interested observation (sometimes uncomfortably sharp!). It has been a great pleasure for me to assemble this guidebook as a permanent reminder of John Rodgers' commitment to field geology and his impact on geological studies in Connecticut.

I would like to thank several people for their assistance in the assembly and editing of this guidebook. Sidney Quarrier of the Connecticut State Geologic and Natural History Survey has provided great help (and moral support in tough times), especially in the printing of the book. Craig Dietsch and Kim Waldron of Yale have been very helpful in the production of the guidebook and the organization of the N.E.I.G.C. meeting in New Haven. Judy Couture, Linda Phillips and Beth Lofquist of the Yale Geology Department have cheerfully and generously helped me with many of the less glamorous but necessary details of meeting organization and guidebook preparation. And last, but not least, I would like to thank the authors of the trip guides for providing me with excellent contributions which should make this a most useful guidebook both now and in the future.



NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE  
77TH ANNUAL MEETING

HONORING

JOHN RODGERS

For nearly four decades John Rodgers has been an active and vital part of Connecticut geology. In his teaching at Yale University, in his involvement with the work of the Connecticut Geological and Natural History Survey and the U.S. Geological Survey, and in other associations too numerous to name, he has enthusiastically provided generations of students and colleagues with information and inspiration. Often acting as the focal point of intense discussion of ideas and interpretations, he has continuously encouraged the exploration of many facets of the State's geology. John's outlook has never been narrow, and his linguistic ability and wide travel have broadened not only his own perspective, but others as well. His skill at synthesizing diverse views is amply demonstrated in his compilation of the 1985 Bedrock Geological Map of Connecticut which incorporates the results of more than thirty years of detailed mapping by many different (and differing) geologists. John's part in geological activity in Connecticut is a phenomenon as distinct as the Moodus Noises; and when he favors us with his impromptu piano recitals, far more melodious. It is a great pleasure, therefore, to honor John Rodgers as friend and challenging colleague at this 77th meeting of N.E.I.G.C.

Joe Webb Peoples

NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE

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A special word of thanks is in order to all the secretaries and illustrators whose extra effort at producing crisp, camera-ready copy on a very tight schedule has made this guidebook the attractive piece of work that it is.

The fact that a locality is described in this guidebook does not imply that the public has access to the locality. Stopping on a limited access highway is forbidden by a regulation of the State Traffic Commission, which prohibits all vehicles from stopping or parking on any part of the highway. These regulations also prohibit pedestrians on any limited access highway. Field trip features on these highways can be viewed from other ground. In other instances, stops on private property require permission of the owner. Anyone planning to go on this field trip should check carefully the suggested stops, and do nothing to jeopardize their use by geologists in the years ahead.

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## GEOLOGY OF THE WATERBURY DOME

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## INTRODUCTION

The purpose of this field trip is to examine and discuss rocks within and flanking the Waterbury dome. Questions concerning the origin and evolution of these rocks have been longstanding and important geologic problems still remain. Current understanding of, and controversy surrounding, the stratigraphy, structure, and metamorphism of the rocks seen on this trip will be placed in a regional context. The information presented here represents part of a Yale University Ph.D. dissertation.

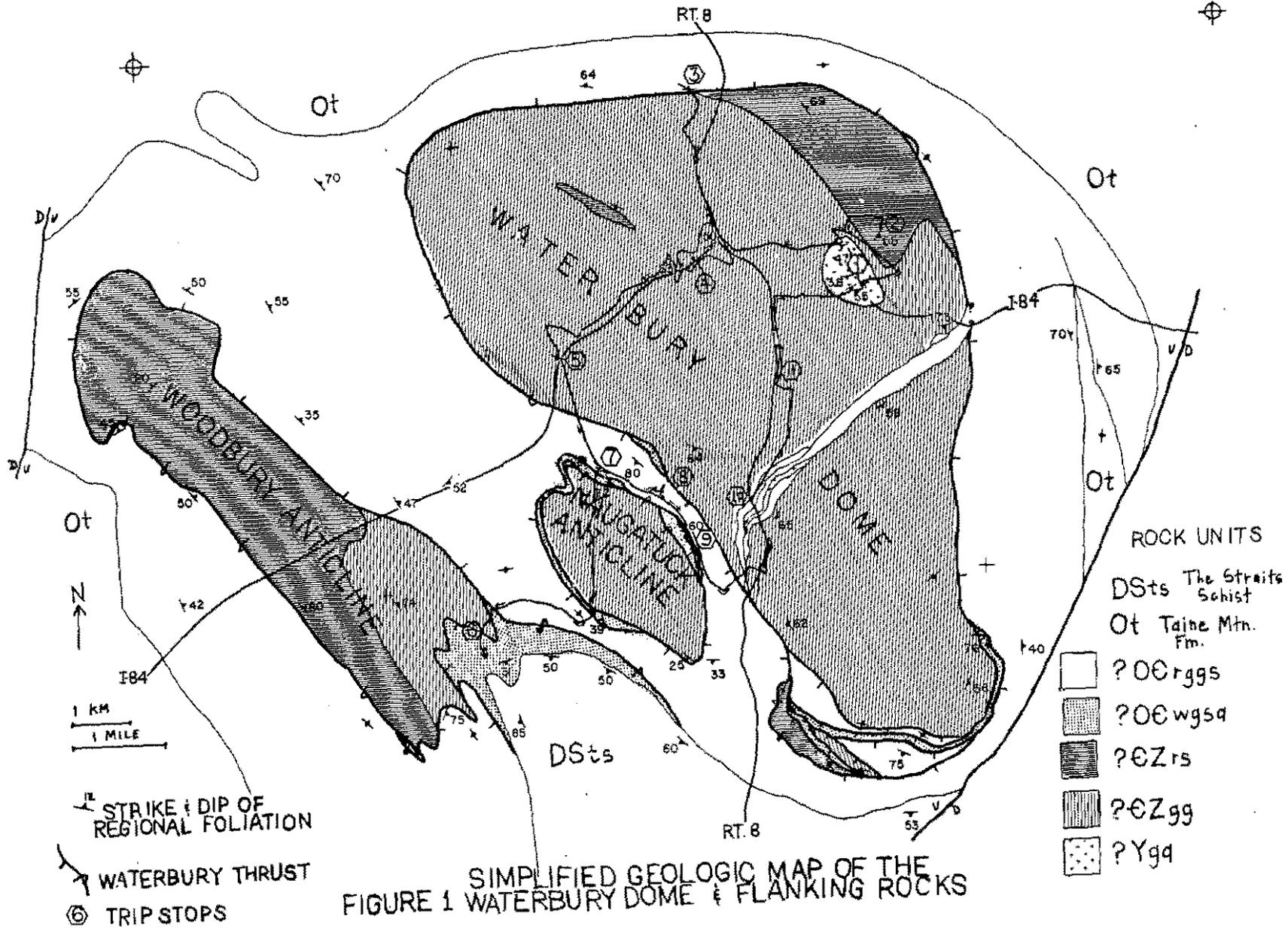
The area in question covers the southern two-thirds of the Waterbury quadrangle (Gates and Martin, 1967), and adjoining parts of the Naugatuck (Carr, 1960), Southington and Mt. Carmel (Fritts, 1963), Southbury (Scott, 1974), and Woodbury (Naruk, 1978) quadrangles. As compiler for the new Connecticut bedrock map, Rodgers (1985) modified and integrated these quadrangle maps but the map pattern shown on the new state map has been changed by my work. Unfortunately, my mapping was not completed in time to be incorporated on the new state map.

Outline of major structures and redefinition of the Waterbury dome Remapping has established major fold structures that dominate the map pattern, in particular, one large dome and two previously unrecognized doubly-plunging anticlines (Figure 1). These structures form a more-or-less parallel set trending roughly N40<sup>0</sup>W and are separated by similar trending synclines. Dominating the eastern half of the map area is a large, internally complex dome to which I prefer to restrict the name "Waterbury dome". It is precisely on the northern, eastern, and southern flanks of this structure that steeply inclined regional foliation wrap around a "central complex of gneisses" and traditionally define the Waterbury dome. Although its map pattern is that of a rather broad culmination, foliation within the Waterbury dome is characteristically steep. On the southwestern flank of the Waterbury dome and separated from it by a narrow syncline, is the Naugatuck anticline, overturned towards the northeast. West of the Naugatuck anticline is the larger Woodbury anticline, nearly isoclinal and overturned towards the southwest.

## REGIONAL GEOLOGICAL SETTING

The Waterbury dome is the southernmost of the series of domes in the Connecticut Valley Synclinorium in western New

A1-2



SIMPLIFIED GEOLOGIC MAP OF THE  
FIGURE 1 WATERBURY DOME & FLANKING ROCKS

England. In western Connecticut, the Synclinorium consists of amphibolite-grade Lower(?) to Middle Paleozoic feldspathic clastic and volcanoclastic rocks that have been intruded by small, mostly granitic to dioritic plutons with ages ranging from 440 to 383 m.y. (Mose and Nagel, 1982).

The western boundary of the Synclinorium in Connecticut is Cameron's Line, separating the Synclinorium from the massifs of North American basement of Grenville age and their autochthonous and allochthonous cover. Cameron's Line is generally regarded as a major tectonic boundary (Rodgers, 1985) and is equivalent to the Whitcomb Summit thrust of Zen, et. al. (1983). To the east, the Mesozoic Hartford Basin separates the Synclinorium in Connecticut from the Bronson Hill Anticlinorium.

The Waterbury dome is more-or-less equidistant -- 30-40 kilometers (20-25 miles) -- from North American basement and its cover to the west, dome gneisses of the Bronson Hill to the east, and Avalonian basement to the southeast, that crops out in the southernmost domes of the Bronson Hill Anticlinorium along the Connecticut coast, as far west as East Haven. Thus the age and correlation of units in the Waterbury dome have important regional tectono-stratigraphic implications.

#### STRATIGRAPHY

Five major lithologic units are now recognized within the map area:

(1) unit ga -- gneiss and amphibolite. Unit ga consists of masses 10-20 meters across of more-or-less texturally homogeneous granite, granodiorite, and quartz-diorite interlayered with migmatized gneiss of similar composition and subordinate hornblende gneiss, amphibolite, and biotite-garnet gneiss. In the field, granite gneiss of unit ga is a distinctive white, and less commonly pink rock with large feldspar megacrysts. Hornblende gneiss and amphibolite layers range from 0.5-2 meters thick.

(2) unit gg -- gneiss and granulite. Unit gg consists of a heterogeneous assemblage of light- to dark-gray layered, migmatitic, and massive plagioclase-quartz-biotite+/-alkali feldspar+/-garnet+/-muscovite+/-kyanite gneisses and granulites. Present throughout unit gg in subordinate amounts are biotite schist, biotite-muscovite schist, quartzite, amphibolite, and calc-silicate rock. No regular pattern of distribution of these rock types has yet been found on which to base an internal stratigraphy.

Underlying large tracts of unit gg are rusty-weathering layered rocks consisting of alternating biotite-kyanite schist and quartz-microcline-plagioclase gneiss; individual layers are 1 centimeter or less thick and the schist layers

form distinctive ridges on outcrop. This particular layered gneiss is texturally gradational into units (see below).

(3) unit rs -- rusty schist. Unit rs consists of fine-grained, commonly migmatitic biotite-quartz-plagioclase-muscovite-garnet-microcline schist and schistose gneiss. Scattered throughout unit rs in minor amounts are small boudins of calc-silicate rock, quartzite, and amphibolite.

(4) unit wgsa -- white granitic gneiss, schist, and amphibolite. Unit wgsa consists of irregularly interlayered white weathering, medium-grained quartz-microcline-plagioclase-biotite-muscovite+/-garnet gneiss, rusty-weathering fine- to medium-grained granular quartz-plagioclase-biotite-muscovite-garnet schist and schistose gneiss, amphibolite, and quartzite.

(5) unit rggs -- rusty gneiss, granulite, and schist. Unit rggs is dominated by rusty-weathering quartz-plagioclase-alkali feldspar-biotite-muscovite+/-garnet+/-kyanite gneisses and granulites. Present throughout the unit are distinctive, although subordinate, graphitic, generally sulfidic, quartz-plagioclase-orthoclase-biotite-kyanite-muscovite-garnet schists, and amphibolites. Other minor rock types include quartzite, hornblende gneiss, granitic gneiss, mica schist (without graphite), and calc-silicate rock. As with unit gg, no regular pattern of distribution of individual rock types has yet been found on which to base an internal stratigraphy.

#### INTRUSIVE ROCKS

A group of intrusive trondhjemitic, tonalitic, and quartz-dioritic rocks, and several small bodies of granite and granodiorite are present throughout the map area and intrude all of the mapped units. These rocks occur as small sills and plugs commonly not exceeding several tens, and in some cases a few hundred, meters in their longest dimension. Granitic pegmatites are also scattered throughout the area.

Recognition of (meta-)intrusive rocks in the map area is not always simple because upper amphibolite grade metamorphism in many instances has blurred the distinction between intrusive igneous rocks and "ortho"-gneisses and granulites produced by in situ partial melting. This is particularly true for the trondhjemitic and tonalitic rocks that intrude unit gg because their mineralogy is similar to the paragneisses of this unit.

The age(s) of the intrusive rocks have not yet been determined.

#### STRUCTURE

Major regional structural features Structurally above the Waterbury dome and plunging off it to the north and south,

refolded nappes are outlined by the Silurian-Devonian The Straits Schist. Dieterich (1968) proposed the following sequence of folding for the evolution of these structures (see Trip B-1, Skinner and Rodgers, Figure 2):

- (1) regional scale, west-directed isoclinal folding,
- (2) regional scale, east-directed isoclinal "backfolding",
- (3) tight to open folding, trending N-S and NW-SE.

Mose and Nagel (1982) reported an Rb-Sr whole-rock isochron age of  $383 \pm 5$  m.y. for the undeformed, pegmatitic Nonewaug granite which cross-cuts deformed The Straits Schist.

Hall (1980) recognized that the isoclinal folds outlined by The Straits Schist do not involve the "core" rocks of the Waterbury dome as it was then mapped. Remapping has shown that the inferred decollement exists at the outer contact of unit rrgs; thus rocks exposed in the cores of the Waterbury dome (as here defined) and the Woodbury and Naugatuck anticlines, as well as unit rrgs which overlies them, were detached from isoclinal folding of Acadian age.

Sequence of deformation in the map area. Rocks in the map area record at least seven deformation events. The regional foliation ( $S_2$ ) in the five mapped units, including coarse migmatitic layering, is interpreted to have been produced by an early stage of thrusting ( $d_2$ ) that established the contact relations now observed. The age of thrusting has not been directly established. Relicts of older fabrics within the regional foliation in these rocks demonstrate that thrusting was superposed on older structures.

After thrusting, the rocks were deformed by four phases of folding; the earliest phase ( $F_3$ ) has only been recognized west of the Naugatuck anticline. The map pattern is dominated by folds trending NW ( $F_4$ ) that can be traced directly into The Straits Schist. The Naugatuck and Woodbury anticlines are  $F_4$  anticlines. These  $F_4$  anticlines, trending NW, are deformed by younger folds trending NE ( $F_5$ );  $F_5$  folds superposed a doubly-plunging geometry on the  $F_4$  anticlines. Likewise, the Waterbury dome formed by the interference of  $F_4$  and  $F_5$  folds. Domes produced by fold interference confirm Hall's (1980) view. Whether interference was accompanied by buoyant rise of rocks now in dome core is uncertain.

The youngest folds are broad open folds with horizontal to steeply plunging axes with variable trends. Brittle faults post-date all other structures and are present throughout the map area.

$d_2$  thrusting and the Waterbury thrust The contacts between map units that outline the Naugatuck and Woodbury anticlines and the Waterbury dome are interpreted to mark major  $d_2$  faults or fault zones, interpreted as thrusts. Evidence to support this interpretation includes (1) map scale as well as

local truncation, and out of sequence juxtaposition of units along these contacts; (2) gneisses exposed along and adjacent to these contacts have augen and broken and strung out ribbons of quartz-feldspar, and have boudins truncated by shear foliation; (3) tabular bodies of trondhjemite with mylonitic foliation are found along these contacts; and (4) very fine-grained, thin, anastomosing bands of kyanite that surround lens-shaped aggregates of recrystallized quartz and biotite have been found only in schists along and adjacent to these contacts; these bands can be interpreted as mylonite.

Where unit rggs surrounds the Waterbury dome and units gg and rs exposed in the cores of the Naugatuck and Woodbury anticlines, it forms an overlying tectonic cover. The contact that separates unit rggs from the rocks below it has been informally named the Waterbury thrust. A fault (thrust?) zone that cuts across the Waterbury dome (Stop 10) indicates that the dome itself consists of two large, thrust imbricated parts.

#### METAMORPHISM

The rocks of the map area form an upper-amphibolite facies metamorphic high within the regional amphibolite facies terrane of the western Connecticut highlands. Previous petrologic work has been limited, for the most part, to the identification and mapping of mineral assemblages. O'Connor (1973) mapped isograds in the area and discussed conditions of metamorphism.

Detailed petrologic and textural evidence supports a polyphase metamorphic evolution for the map area involving three "phases": (1) early kyanite-garnet grade metamorphism,  $M_1$ , (2) high P/high T metamorphism involving migmatization,  $M_2$ , and (3) "retrograde" metamorphism, but still in the amphibolite facies, involving recrystallization and hydration,  $M_3$ .

The P-T conditions of  $M_2$  and  $M_3$  metamorphism have been estimated by combining careful petrographic analysis of mineral assemblages and textures with quantitative thermo-barometric techniques. Details of the methods used, the problems associated with the calculations, and the mineral compositions used are given by Dietsch (1985).

Estimates of the conditions of  $M_2$  metamorphism are  $T = 700 - 765^\circ \text{C}$  and  $P = 8.5 - 9.0 \text{ kb}$ , and for  $M_3$  metamorphism,  $T = 550 - 590^\circ \text{C}$  and  $P = 4.8 - 7.5 \text{ kb}$ . Geochronologic results from the region indicate that the superposed, lower grade metamorphic effects are Acadian.

Isograds O'Connor (1973) mapped two isograds: (1) a staurolite-disappearance isograd marking the upper limit of the assemblage staurolite-muscovite-quartz, and (2) a kyanite K-feldspar isograd marking the first appearance of

kyanite and K-feldspar in mutual contact without intervening muscovite. Muscovite is present in all pelitic and semi-pelitic rocks above the kyanite K-feldspar isograd. All of the stops on this trip, except stops 3 and 6, lie well above the kyanite K-feldspar isograd. The occurrence of co-existing kyanite and K-feldspar in migmatitic leucosomes, combined with analysis of phase equilibria and thermobarometric calculations, indicate that the stable co-existence of kyanite and K-feldspar did not result from the subsolidus dehydration of muscovite. Petrologic relations are consistent with melting reactions involving quartz, plagioclase, and muscovite producing the co-existence of kyanite and K-feldspar. Thus the kyanite K-feldspar isograd delimits the critical assemblage kyanite + melt.

O'Connor (1973) and Naruk (1978) reported a few isolated occurrences of sillimanite and only along the northern margin of the Waterbury dome can a sillimanite isograd be drawn. Ubiquitous kyanite attests to the relatively high pressure during all phases of metamorphism.

On the eastern flank of the Waterbury dome a normal fault separates staurolite-bearing The Straits Schist from kyanite K-feldspar grade rocks. On the southern flank of the Waterbury dome and south of the Naugatuck and Woodbury anticlines, staurolite-bearing The Straits is interpreted to be in thrust contact with units wgsa and rggs. North of the Waterbury dome, the limit on the occurrence of staurolite coincides with the contact between the Taine Mountain Formation and unit rggs. This contact marks an Acadian decollement.

CORRELATIONS

A "three-tiered" stratigraphy is proposed for the five major units:

unit rggs -- Rusty-weathering mica-quartz-plagioclase gneiss of unit rggs along the western flank of the Woodbury anticline and gray-weathering quartz-plagioclase-biotite gneiss and granulite along the southern and eastern flanks of the Waterbury dome are lithically similar to Stanley's (1964) Scranton Mountain Member and Wildcat Member, respectively, of the Lower? Ordovician Taine Mountain Formation. Stanley (1963) defined the Taine Mountain Formation in the northern third of the Bristol dome in the Collinsville quadrangle. The carbonaceous, sulfidic, rusty-weathering schist of unit rggs is similar to the Middle Ordovician Hawley Formation of the so-called Hartland Belt of western Connecticut. Unit rggs is thus broadly correlated with these Lower? to Middle Ordovician rocks.

unit wgsa -- Unit wgsa is lithically similar to rocks that Hall (1976) mapped as various members of the Hartland Formation in the White Plains-Glenville area of southwesternmost

Connecticut and adjacent New York. Unlike Hall's (1976) Hartland Formation, no internal stratigraphy with unit wgsa has been recognized. Unit wgsa has only been mapped as large, fault-bounded "slivers" along the Waterbury thrust where the thrust outlines the Waterbury dome and the Naugatuck and Woodbury anticlines; thus, it is possible that rocks mapped as unit wgsa along the Waterbury thrust are meta-mylonite. Complicating matters, the white granitic gneiss included in unit wgsa may be an intrusive rock.

unit rs -- The gnarly, rusty-weathering biotite-garnet-kyanite schist and schistose gneiss of unit rs is in places lithically identical to the allochthonous schistose rocks in western Connecticut mapped as Cambrian to Proterozoic Z(?) Waramaug Formation, Manhattan Schist, and Canaan Mountain Schist. Correlation of unit rs with this group of units is perhaps the strongest of all of the proposed correlations.

unit gg -- The rusty-weathering, layered gneisses of unit gg that are texturally gradational into unit rs are lithically similar to rocks mapped by Martin (1970) as Waramaug Formation in the Torrington quadrangle. Where unit gg consists of interlayered gray to green plagioclase-quartz granulite, biotite-muscovite schist, quartzite, and quartz-rich gneiss (Stop 11), it is lithically similar to the clastic rocks that characterized the low Taconic sequence of Zen (1967) in southern Vermont and western Massachusetts and adjacent New York. Unit gg is thus broadly correlated with the Cambrian to Proterozoic Z(?) clastic sequence of the Taconic allochthons (including "Group 3 Taconicallochthons" of Stanley and Ratcliffe, 1983).

unit ga -- No good lithologic counterpart of these rocks is present in the region. Only unit ga contains large masses and sills of leucocratic granitic gneiss which lie among other, thoroughly migmatized, gneisses that may represent ultrameta-morphosed greywackes and/or dacitic volcanics. The presence of the granitic gneiss in unit ga leads me to propose that this unit represents Precambrian basement and I further suggest that unit ga correlates with basement of Grenville age exposed in the massifs of western New England and in the Chester, Athens, Rayponda Lake, and Sadawga Pond domes in southern Vermont. If the correlations of units rs and gg are reasonably correct, correlation of unit ga with Grenville basement would be entirely consistent with regional structural and stratigraphic relations.

These stratigraphic correlations are presented as a possible model. I share the view summarized by Stanley and Ratcliffe (1983) and shown by Zen, et. al. (1983) that the major map units of the Connecticut Valley Synclinorium are largely -- if not exclusively -- in fault contact. What do correlations mean in such a pseudo-stratigraphy?

If the correlations of units rggs, rs, and gg are correct, then the Waterbury thrust marks the boundary between rocks of the Hartland Belt (unit rggs) and rocks of the Taconic sequence (units rs and gg) and is equivalent to Cameron's Line in western Connecticut (Rodgers, 1985). If the Waterbury thrust is indeed a resurfacing of Cameron's Line then the high P/high T  $M_2$  metamorphism is considered to be equivalent to high P (and lower T) Ordovician metamorphism recognized and dated within the Synclinorium in north-central Vermont.

ROADLOG            WARNING!    Many of the stops on this trip are roadcuts along major highways. To reach some stops requires walking along, and crossing, ramps and roads. Be aware of the traffic and proceed with caution!

Mileage  
total interval  
0.0        0.0    BEGIN TRIP. Commuter parking lot, Hamilton Ave., Waterbury. (All of the stops are located within the Waterbury quadrangle except Stop 6).

STOP 1. (2 hours approximately) Tectonic window? and type area of unit ga in the core of the Waterbury dome. Turn left (north) from commuter lot and walk to Washington Ave. Turn left on Washington and walk up to traffic light. At light, cross over to roadcut.

The dominant rock types here are medium-grained granodiorite and tonalite that are "soaked" with leucocratic muscovite- and muscovite-biotite granitic gneiss. Seams of biotite-garnet gneiss may represent migmatitic melanosomes. Gray, fine-grained sill-like bodies of trondhjemite are also present and this particular rock type will be seen throughout the area. Field relations seen at Stops 4, 5, 8, and 10 demonstrate that intrusion of trondhjemite probably spanned  $d_2$  deformation.

At the east end of this cut walk up slope to huge roadcut on the east side of Pine Hill. A large body of leucocratic, migmatized granodioritic gneiss (average mode plg 38, qtz 33, ksp 18, bio 8, mus 3) is well-exposed a short distance along the cut. Observe that the large K-feldspars are porphyroblasts, cutting across the local foliation. In thin-section, the muscovite present in this granodioritic gneiss (and in the intrusive trondhjemites) is poikiloblastic; this and traces of chlorite attest to retrograde metamorphic effects. This leucocratic granodioritic gneiss is present only here at Pine Hill and in exposures in the immediate area, and it is recognized as an integral part of unit ga. Without detailed geochemical data it is not possible to determine its protolith or the protoliths of the other ultra-metamorphosed rocks of unit ga.

A large, E-W-trending recumbent fold and E- and NE-trending  $F_5$  "dome-stage" folds are present within this roadcut but are best seen from a vantage point away from the roadcut.

Return to Washington Ave. and walk under overpass to Hamilton Ave. Turn left and walk down to traffic light. Turn left at light to roadcut on I-84 entrance ramp.

Various mesoscopic structures are well-exposed here. Strongly sheared tonalite is cross-cut by granitic veins and dikes, and mylonitic biotite-rich melanosome(?) layers are conspicuous. Crenulations in these layers are axial planar to large NE-trending  $F_5$  folds. Note the thin phyllonite seam along the southeast-dipping limb of the large  $F_5$  fold, presumably developed at a relatively late, cool stage of "doming" and related to the arching of a stiff gneissosity.

At west end of this cut, climb southwestwards over the hill to roadcuts on westbound I-84. Ramp-like  $F_5$  folds can be seen across I-84 near the western end of the Pine Hill roadcut; asymmetric minor folds within the "steep zone" of this fold have a consistent SE-over-NW sense of movement.

Amphibolites and layers of hornblende gneiss within unit ga are well-exposed in the roadcut along westbound I-84. Note the folded internal foliation within the large boudin at the west end of the cut, truncated by the dominant foliation of the enclosing gneisses. This  $S_i$  may be a relict of  $d_1$  or older deformation.

Climb back over hill to entrance ramp and Hamilton Ave. and return to vehicles.

- |     |     |   |
|-----|-----|---|
| 0.0 | 0.0 | Exit from commuter lot: turn left on Hamilton Ave., get in right lane.  |
| 0.2 | 0.2 | At traffic light turn right on Silver St. (Rt. 69). Where road widens, get in left lane.  |
| 0.6 | 0.4 | At first light, turn left on East Main St.  |
| 1.3 | 0.7 | Traffic light. Make first right past light on Wall St. (drug store on corner).  |
| 1.5 | 0.2 | Bear right on Wall, continue straight at stop sign up steep hill.   |
| 1.9 | 0.4 | Just before crest of hill, turn right on Hobart St. Park at the end of Hobart on small dirt pull-out opposite last house on west (right). |

STOP 2. (30 minutes approximately) Type area of unit rs.  
Follow path at the end of Hobart to outcrops on the west side  
of hill.

The distinctive field appearance of this migmatitic schistose gneiss has enabled it to be separated from rocks previously mapped together as "Waterbury gneiss" and leads to the proposed correlation of unit rs with the Waramaug Formation-Hoosac Schist-Canaan Mountain Schist. The assemblage of unit rs here is biotite-kyanite-plagioclase-muscovite-garnet-microcline-pyrrhotite-pyrite. Small boudins of garnet amphibolite, calc-silicate rock, and rare quartzite are present and are characteristic of the unit.

Migmatitic leucosomes cross-cut the dominant  $S_2$  foliation and contain microxenoliths of schist, and are also isoclinally folded parallel to  $S_2$ .

Follow outcrops south where a large body of dirty-white weathering, medium-grained (anatectic?) granite cross-cuts unit rs. Minor open to chevron folds with NNE-trending axes deform the granite-unit rs contact. If this granite is indeed anatectic then its crystallization age would place limits on the timing of both deformation and metamorphism.

Return to vehicles.

Return to Wall St. Turn left and retrace Wall down the hill to East Main St. Prominent ridges seen to the west from the crest of Wall are held up by NW-striking, steeply dipping units gg and rs of the north-central portion of the Waterbury dome.

- |     |     |  |
|-----|-----|--|
| 2.6 | 0.7 | Turn right on East Main St.  |
| 2.7 | 0.1 | At first light, turn left on Hamilton Ave.   |
| 3.0 | 0.3 | At light, turn right to I-84 West (Danbury) then bear left onto I-84. KEEP RIGHT on I-84.                            |
| 4.2 | 1.2 | Take Exit 20, Rt. 8 North (Torrington). Stay left on Rt. 8.  |
| 5.2 | 1.0 | Take Exit 35, Rt. 73 (Oakville/Watertown). Move into right lane on Rt. 73 exit.                                      |
| 5.7 | 0.5 | At first light, turn right on East Aurora St.  |
| 6.0 | 0.3 | Turn left on Gear St.  |
| 6.2 | 0.2 | Turn left on Huntingdon Ave., then make immediate right on Brookside Dr. Park near crest of hill just past outcrops. |

STOP 3. (30 minutes approximately) Unit rggs adjacent to inferred contact with unit gg on northern flank of the Waterbury dome.

These rocks lie within O'Connor's (1973) kyanite-garnet-biotite zone just below the kyanite-K-feldspar isograd. The dominant rock types here are light to dark gray, fine- to medium-grained quartz-plagioclase-biotite-muscovite+/-garnet gneiss and granulite, and brown weathering biotite schist. Interlayered garnet-bearing granitic gneiss and cross-cutting, medium-grained garnet-bearing pegmatite are also present. At the south end of the roadcut (on the east side) is a small knob of green weathering tremolite-quartz-phlogopite rock; more extensive exposures of this and similar rock types can be seen at Stop 4.

The rocks are striking about E-W and are vertical or dip steeply north off the dome. Note the broken and strung-out layers of quartz and feldspar and the porphyroclasts of feldspar. A strong lineation plunges steeply to the NNW and is well developed in the pegmatites near the north end of the cut, defined by feldspar porphyroblasts.

Detailed petrologic study, including microprobe analysis, has been done on schists from this roadcut with the assemblage quartz-plagioclase-muscovite-kyanite-biotite-garnet-pyrrhotite. In garnets with quartz inclusion-filled cores and clear rims, that have overgrown the  $S_2$  schistosity, a compositional discontinuity coincides with the boundary between inclusion-filled cores and clear rims; this indicates two episodes of garnet growth. Quantitative temperature estimates here were obtained here using Thompson's (1976) empirically derived garnet-biotite Fe-Mg exchange thermometer (Thompson's equation is also used for the other temperature estimates cited below). Using rim compositions of these same garnets and rims of adjacent biotite, temperature estimates range from 570-605° C. A minimum pressure of 6.1 kb was obtained using the P-T contours of ternary Fe-Mg-Mn garnet composition of Tracy, et. al. (1976) for the assemblage garnet + cordierite + kyanite + quartz. This is a minimum pressure because no cordierite is present. These are estimates of  $M_3$  metamorphism.

Return to vehicles.

Use driveways at crest of hill just past outcrops to turn around. Retrace Gear St. and E. Aurora St. to Rt. 73 (Watertown Ave.).

6.9      0.7      At light, turn left on Rt. 73 to Rt. 8 South/I-84. Follow Rt. 8 south. KEEP RIGHT

7.9      1.0      Take Exit 33, I-84 West (Danbury). KEEP RIGHT

- 8.6      0.7      Take first exit, Exit 18, West Main St./ Highland Ave. Roadcuts on right are Stop 4. Keep/bear left on exit ramp to West Main St.
- 8.9      0.3      At light at end of ramp, turn right and make immediate right into A&P parking lot. Park.

STOP 4. (1 1/4 hours approximately) Antiformal "bow-tie" of unit rggs in the core of the Waterbury dome. Retrace exit ramp to roadcuts along I-84 that begin just east of the exit. WARNING! Proceed with caution to the roadcuts! Where the exit ramps splits is a very dangerous area! Examine the rocks while slowly making your way ENE down the hill, stopping just past the Highland Ave. overpass.

Interpretation of the map pattern and the location of the "bow-tie" right along the major NW-trending antiformal axis of the dome, combined with quantitative P-T results from rocks collected from this roadcut that give the highest calculated pressure estimates from the dome, lead to the conclusion that this area of unit rggs occupies a small culmination; i.e. unit rggs in the "bow-tie" is structurally beneath surrounding unit gg. Unit rggs here is interpreted as part of an imbricate thrust sliver along a  $d_2$  thrust contact at the base of unit gg.

The dominant rock types here are rusty-weathering, sulfidic quartz-plagioclase-mica-garnet-kyanite gneiss and schistose gneiss, and blue-gray, medium-grained tonalitic and trondhjemitic gneiss and granulite. Subordinate rusty-weathering, fine- to medium-grained, friable sulfidic schist with and without calc-silicate pods is also present.

Schistose gneisses from this roadcut have the assemblage quartz-plagioclase-microcline-muscovite-kyanite-biotite-garnet-pyrrhotite-ilmenite-rutile. Matrix rutile and rutile in contact with garnet are rimmed by ilmenite and these textures are suitable for quantitative barometric calculations. Bohlen, et. al. (1983) calibrated the reaction  $3 \text{ ilmenite} + \text{kyanite} + 2 \text{ quartz} = \text{almandine} + 3 \text{ rutile}$ ; this reaction has a positive  $dP/dT$  slope with almandine and rutile on the high P/low T side. Model pressures calculated from this reaction in combination with garnet-biotite thermometry yield P-T conditions of  $730^\circ \text{C}$  and 8.8 kb. These results are consistent with kyanite K-feldspar grade metamorphism and are interpreted to represent conditions obtained after  $d_2$  thrusting.

Good igneous textures can be seen in gray tonalite at the first outcrop adjacent to the highway. Tonalite exposed on the eastern side of the small "quarry" about 10 meters farther along the roadcut. can be seen cross-cutting a (pre- $S_2$ ?) gneissosity.

Scattered along this roadcut at irregular intervals are small isolated blocks of green rock. The largest block of green rock is located just to the west of the Highland Ave. overpass. Observed assemblages of these rocks are diopside-calcite-plagioclase, cross-cut by hornblende veinlets and chromite vienlets; tremolite-Cr-bearing phlogopite-chromite; hornblende-epidote-diopside-apatite; and tremolite-phlogopite-quartz-sericite-calcite. Mineralogical zonation is developed in green rocks in contact with gray tonalite and suggests that intrusion of tonalite produced Ca-metasomatism of ultramafic rocks.

Beneath the overpass is a "mini marble cake" migmatite. Note the cross-cutting 2-mica melts which represent the last stage of melting. The crystallization age of these melts could place an upper limit on the age of migmatization.

Return to vehicles. Exercise caution when crossing exit ramp to Highland Ave. on your way back to West Main St. Watch traffic and proceed quickly across this dangerous intersection!

- |      |     |   |
|------|-----|---|
| 9.0  | 0.1 | Exit from A&P lot at traffic light. Turn left on West Main St. Keep right around right hand curve where road widens and at light, mileage (9.3/0.3), continue straight on Chase Pkwy. |
| 10.4 | 1.4 | At the end of Chase Pkwy., turn right on Rt. 64 West.   |
| 10.7 | 0.3 | At first light, turn left on Rt. 63 South.  |
| 11.1 | 0.4 | I-84/Rt. 63 intersection. Park on shoulder just before I-84 overpass next to roadcut.   |

STOP 5. (20 minutes approximately) Unit gg in the core of the Waterbury dome.

Exposed in this roadcut is one variety of the migmatitic plagioclase-quartz-biotite gneisses of unit gg, here with pink feldspar-bearing leucosomes. Note the zonation of the SW-dipping, garnet-bearing dike and the extremely elongated calc-silicate boudin within the gneiss.

At the south end of this roadcut, a massive body of blue-gray trondhjemite has generally concordant contacts. A small xenolith of foliated plagioclase-quartz-biotite-muscovite-garnet gneiss can be seen near the northern contact of the trondhjemite, a few meters above ground level, showing the intrusive origin of the trondhjemite.

Walk around the north end of this cut to the smaller roadcuts of migmatized gneiss along the east side of the entrance ramp onto I-84. Rocks collected here by Clark and Kulp (1968)

yielded a best-fit Rb-Sr whole-rock "errorchron" age of 455+/-50 m.y. Biotite from this roadcut analyzed by Clark and Kulp gave a conventional K-Ar age of 345+/-13 m.y. (both results are recalculated numbers using the decay constants of Steiger and Jager, 1977).

Return to vehicles.

- 12.4      1.3      Turn right on Allerton Farms Rd.
- 13.8      1.4      Prominent ridge seen from crest of Allerton Farms Rd. is held up by staurolite-bearing The Straits Schist, dipping to the south and southwest. This ridge of The Straits cores the structurally lowest regional recumbent fold of Acadian age that overlies the Waterbury dome.
- 14.0      0.2      At stop sign, turn left on Jones Rd.
- 14.1      0.1      At stop sign, turn right on Neuman St.
- 14.3      0.2      At stop sign, turn right on Rubber Ave.
- 14.5      0.2      Carbonate block exposed on right hand side of Rubber Ave. Blocks of carbonate, calc-silicate rock, quartzite, and amphibolite have been mapped at numerous localities at the base of The Straits Schist and are considered to be a basal member of The Straits in Connecticut. Hatch and Stanley (1973) correlated these rocks at the base of The Straits with the Middle Silurian Russell Mountain Formation of western Massachusetts.
- Rocks of unit rrgs are exposed along the entrance to the Southwood Apts. just past the block of carbonate.
- 15.4      0.9      At small grass traffic triangle, bear left down hill on Guntown Rd.
- 15.9      0.5      Cemetery on left; at mileage 16.0/0.1 past cemetery, turn around at paved drive on left hand side. Park next to cemetery at mileage 16.1/0.1.

STOP 6. (45 minutes approximately) (Naugatuck quadrangle) Traverse across unit wgsa-unit gg contact along the southern margin of the Woodbury anticline. Walk about 50 meters up Guntown Rd. and then cut left up hill through the woods to outcrops along the small ridge. Examine the rocks while slowly following outcrops northeastwards. Continue until ridge turns north into wide gully.

Most of these outcrops consist of granoblastic, leucocratic granitic gneiss, mapped as unit wgsa. Present at irregular intervals are layers of amphibolite, quartzite, lustrous tan-weathering feldspathic biotite-muscovite schist, and gray weathering, medium-grained granulite. These subordinate rock types become more common as the contact with unit gg is approached. The contact is drawn where dark, rusty-weathering migmatitic schistose rocks begin. Unit gg here contains co-existing kyanite and K-feldspar but rocks mapped as unit wgsa show no evidence of this upper amphibolite grade metamorphism.

L. Hall (pers. comm., 1983, along these same outcrops) likened this sequence of rock types to the contact he mapped between Harland Formation (unit wgsa) and Manhattan Schist (unit gg) in southwesternmost Connecticut.

Four fold sets can be seen in unit wgsa along this ridge. These folds have been correlated with the same sequence of structures mapped by Dieterich (1968) to the south and south-east in rocks involved in the sequence of Acadian folding. Structures correlated with the second phase of Acadian isoclinal folding ( $F_3$  in the sequence outlined above) are only present west of the Naugatuck anticline. An isoclinal  $F_3$  fold with an axial planar foliation is exposed about 15 meters SW of the contact in the upper portion of the outcrop. Mica crenulations in its hinge are remnants of an older foliation ( $S_2?$ ). This fold is folded by NW-trending  $F_4$  chevron folds with axes plunging steeply to the SE. Two sets of younger open folds produce undulations on the outcrops;  $F_5$  folds have axes plunging gently to the SW and  $F_6$  folds have subhorizontal axes trending NNW.

From exposures of unit gg, descend down hill to road and return to vehicles.

Retrace Guntown Rd. and Rubber Ave. back to Neuman St.

- |      |     |  |
|------|-----|--|
| 17.5 | 1.6 | At stop sign on Rubber Ave., turn left on Neuman St.                                   |
| 17.7 | 0.2 | Turn left on Jones Rd. Watch this tricky intersection, yield to traffic on your right. |
| 17.8 | 0.1 | Turn right on Allerton Farms Rd.   |
| 19.4 | 1.6 | At stop sign, turn right on Rt. 63. Pull off immediately on shoulder and park.         |

STOP 7, optional. (15 minutes approximately) Unit rggs in  $F_4$  syncline between the Waterbury dome and the Naugatuck anticline.

The roadcuts on both sides of Rt. 63 expose large bodies of amphibolite within unit rggs. A variety of migmatitic, sulfidic gneisses and schists, and sulfidic granulites characteristic of unit rggs are exposed. In the amphibolites, ilmenite inclusions in sphene are ubiquitous but other high pressure petrologic indicators are absent. Amphibole compositions are Ca-rich and compositional zoning or overgrowth features indicative of high pressure are absent.

Near the south end of the roadcut on the east side of Rt. 63 a brittle fault is marked by a well-exposed carbonate breccia, a feature common to faults of Mesozoic age in the region.

Return to vehicles.

Continue south on Rt. 63.

20.1      0.7      Turn left into Hop Brook Dam, Army Corps of Engineers parking lot. If gates at the end of the lot are open, follow road across dam to small circle. Park.

STOP 8. (1 hour approximately) Units rggs and gg across their contact on west-central margin of the Waterbury dome.

Examine unit rggs in the small quarry at the east end of the Hop Brook dam. Unit rggs consists here of rusty-weathering, sulfidic schist with interlayered quartzite and small calc-silicate pods. Minor F<sub>4</sub> folds plunging gently to moderately to the SE are well developed here in the sulfidic schist. On top of the quarry, sulfidic schist is in contact with a gray weathering, kyanite schistose gneiss that is similar to rocks seen elsewhere along and near the unit gg-unit rggs contact that outlines the Waterbury dome.

Leave the quarry and follow the wide path to the SE down the east side of the dam; continue along the path, bearing left (NE) at the bottom of the hill, until you emerge from the woods along the side of the dry overflow reservoir. Unit rggs in contact with a sill of gray trondhjemite is exposed in the southernmost outcrop along the west side of the dry reservoir.

Where the cuts resume northwards from this outcrop of unit rggs, unit gg is exposed, consisting of a heterogeneous sequence of strongly deformed, green- and blue-gray, medium-grained migmatitic gneisses and granulites. Compositions range from quartz-diorite and tonalite (generally the darker layers) to granodiorite (generally the lighter layers) and garnet and kyanite are common minor phases. These rocks are similar to unit ga seen at Stop 1 but the large bodies and sills of leucocratic granodiorite and pink feldspar-bearing granities of unit ga are absent here. Thin, green-gray

schist layers which are characteristic of unit gg are common in the gneisses here and help to distinguish the two units. Dark green garnet amphibolites and calc-silicate rocks with the assemblage hornblende-plagioclase-clinzoisite-diopside-quartz-garnet-sphene-rutile are common in the southern 50 meters of the large continuous cut.

A variety of minor structures are present, including  $F_4$  folds that plunge to the SE like those seen in unit rggs in the quarry, but of particular interest are small shear zones and mesoscopic folds with mylonitic axial planar fabrics that deform migmatitic leucosomes and leucocratic 2-mica granitic veins. These structures have a consistent SE-over-NW sense of asymmetry similar to asymmetric structures associated with  $F_5$  folding of unit ga. What is the relation, if any, between these two sets of structures?

Retrograde metamorphism ( $M_3$ ) of the gneisses exposed in the dry reservoir can be seen in thin-section. Remnants of garnet are embayed by plagioclase, muscovite, and biotite, and remnants of kyanite in schist layers are completely enclosed in large mats of muscovite. P-T conditions of retrograde metamorphism affecting these rocks have been estimated using the equations of Ghent and Stout (1981) for the reactions pyrope + grossular + muscovite = 3 anorthite and almandine + grossular + muscovite = 3 anorthite. Average rim compositions of garnet, plagioclase, biotite, and muscovite that form the retrograde textures yield  $T = 575^\circ \text{C}$  and  $P = 6.4 \text{ kb}$ .

After looking at rocks on the north side of the concrete barrier, return to vehicles.

20.4      0.3      Leave parking lot, turn left on Rt. 63.  
NOTE: total mileage assumes parking across the dam for Stop 8. Subtract 0.2 from total mileage if you did not cross the dam to park for Stop 8.

20.9      0.5      Pull off Rt. 63 over curb and park.

STOP 9, optional. (15 minutes approximately) Intrusive plug of trondhjemite cross-cutting unit rggs

The intrusive contact of a medium-grained plug of trondhjemite is well exposed up along the central portion of the outcrop. Xenoliths of calc-silicate gneiss and sulfidic schist of unit rggs are present in the central and northern portions of the outcrop.

21.7      0.8      At first light, turn left on Rt. 68 East.

22.1      0.4      Turn right, follow Rt. 68 over Rt. 8.

22.3      0.2      At light, turn left on Golden Ct.

- 22.4 0.1 At stop sign, turn right on Golden Ct. (old Rt. 8), follow entrance ramp onto (new) Rt. 8 North. KEEP RIGHT
- 23.0 0.6 Take Exit 29, South Main St., Waterbury.
- 23.2 0.2 At light at end of ramp, continue across South Main St. into parking lot. Park.

STOP 10. (1 1/4 hours approximately) Fault -- thrust? -- zone marking the imbrication of the Waterbury dome. From the parking lot, cross South Main St. at traffic light. Proceed quickly up the road leading to the Naugatuck Industrial Park to where the roadcut ends. Examine the rocks while slowly returning to South Main.

In this cut, characteristic rock types of units ga, gg, and rggs, striking NE and dipping SE, as well as small plugs of unfoliated, fine-grained trondhjemite, are imbricated in an irregular sequence. The different rock types are commonly separated by isolated layers and lenses of green rock similar to those seen at Stop 4. Many of these green rocks have one the high variance assemblages tremolite-chromite, tremolite-quartz-chromite, or diopside-hornblende-chromite. Chromite veins are present in some of these 2- and 3-phase rocks. Some of the green rocks have calc-silicate assemblages; observed assemblages are diopside-hornblende-Cr-bearing epidote-chromite and diopside-Cr-bearing epidote-garnet.

Also present in this cut are blocks of bona fide ultramafic rock -- metamorphosed harzburgite -- with the assemblage olivine (fo<sub>91-97</sub>)-enstatite (en<sub>91-96</sub>)-tremolite (mg/mg+fe<sub>95-99</sub>)-Fe-Cr spinel. The presence here of harzburgite lends support to the view that the family of chromite-bearing rocks began life as ultramafics.

Detailed petrologic/microprobe study of gneisses of unit rggs from this fault zone produced estimates of both peak M<sub>2</sub> and retrograde M<sub>3</sub> metamorphic P-T conditions. Conditions of M<sub>2</sub> metamorphism were estimated using garnet-rutile-ilmenite-kyanite-quartz reaction textures (see discussion Stop 4): T = 740° C and maximum P = 10.9 kb. Conditions of M<sub>3</sub> metamorphism were estimated using reaction textures involving the breakdown of garnet, aligned parallel to migmatitic layering, to plagioclase. Calculations using Newton and Haselton's (1981) equations to describe the reaction grossular + 2 kyanite + quartz = 3 anorthite, combined with temperatures calculated from adjacent garnet rim-biotite rim compositions, yield conditions of T = 590° C and P = 6.3 kb. These results are in close agreement with those from Stop 8. Olivine-spinel Fe-Mg exchange thermometry from samples of metamorphosed harzburgite yield temperatures in the range 565-580° C.

At the bottom of the hill, follow South Main St. under the Rt. 8 overpass and walk up to the outcrops of the west side of Rt. 8. These outcrops expose mylonitic migmatitic gneisses and gray tonalite cut by mylonitic pink feldspar-bearing pegmatites. From atop these outcrops, look across Rt. 8 to the roadcut that is a lower extension of the rocks just traversed. The large-scale imbrication of rock units is interpreted to mark an important fault zone but the timing of faulting is not completely clear. Fabrics indicative of shearing are only locally present. The metamorphosed ultramafic rocks do not contain shear fabrics; foliation within the fault zone wraps around them. This suggests that the present configuration of the ultramafics, as well as the other rocks, could have been produced by faulting at a late stage in the structural development of the dome.

Return to vehicles. Be aware of traffic when crossing South Main St.!

Leave parking lot. Turn left on South Main St.

24.4 1.2 Turn left on Leonard St.

24.6 0.2 Turn right on South Leonard St. (to Rt. 8).

24.8 0.2 Park on right hand side of street or pull into open area on left.

STOP 11, optional. (20 minutes approximately) Migmatitic unit gg in the core of the Waterbury dome. Leave road and walk northwestwards towards Rt. 8 and then north to the small roadcuts on the east side of the highway.

Migmatitic gneisses of unit gg in this roadcut are inter-layered with various gray and green plagioclase-quartz granulites and gneisses, and biotite-rich schists. The northern portion of the cut typifies the complex relationships between deformation and migmatization in high grade rocks. Relict structures that pre-date the main migmatization can be found in blocks within the migmatitic gneissosity that have a truncated internal gneissosity. Complex cross-cutting relationships can be seen between two generations of leucosomes and the local shear foliation present in the layers of schist, and at least two sets of folds.

Looking westwards across Rt. 8, a large body of trondhjemite can be seen in the northern portion of the roadcut. The extent of brittle faulting present within the dome is typified by the brittle faults seen directly across the highway.

Return to vehicles.

Continue north on South Leonard St.

- 25.6 0.8 At light, turn right on Washington Ave.
- 25.9 0.3 At light at South Main St., turn right then immediate left across South Main. Follow Washington Ave. up and over hill.
- 26.6 0.7 At light, turn right on Hamilton Ave.
- 26.8 0.2 Commuter lot on right. END TRIP.

To New Haven, turn right from commuter lot on Hamilton Ave./Rt. 69. Follow Rt. 69 south to its end at Whalley Ave. in the Amityville section of New Haven, about 17 miles. At traffic light, turn left on Whalley. Follow Whalley, through the Westville section of New Haven, until it merges with Broadway, about 3 miles. Continue straight, Broadway turns into Elm St. (one way).

To Kline Geology Lab, follow Elm past New Haven green. At (east) end of green, turn left on Church St. At fifth light, turn into KGL parking lot.

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ORDOVICIAN DUCTILE DEFORMATION ZONES IN THE HUDSON  
HIGHLANDS AND THEIR RELATIONSHIP TO METAMORPHIC  
ZONATION IN COVER ROCKS OF DUTCHESS COUNTY, NEW YORK

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Introduction

The results reported here stem from a program of detailed mapping of fault zones within the Ramapo seismic zone in New York, New Jersey and Pennsylvania funded by the USGS Reactor Hazards, and Earthquake Hazards Programs and by the Nuclear Regulatory Commission. The objective of these studies is to establish the relationship between geologic structure and low level seismicity that has been well located in this area. The rocks traversed on this trip are dominated by Paleozoic age semiductile shear zones in Proterozoic basement gneiss that forms the seismogenic source rock in the Ramapo seismic zone. Knowledge of the structure and mineralogy of these shear zones should provide important clues to the understanding of deeper parts (depths greater than 10 km) of present day seismogenic structures. On a regional scale these structures localized brittle failure in Early Mesozoic time to create the Triassic-Jurassic Newark basin (Ratcliffe, 1980).

Recent 1:24,000 mapping of the Middle Proterozoic basement rocks and the Cambrian through Ordovician cover sequence of the northern Hudson Highlands by Ratcliffe and Burton has defined a complex system of semiductile deformation zones (Fig. 1) that are largely responsible for deformation of the basement rocks in the Taconic orogeny. We have traced these narrow 0.5 to 100 meter thick zones of intense deformation and determined that the mineral assemblages contained in them record a prograde sequence from chlorite to sillimanite plus muscovite grade that is colinear with the Barrow zonation in Dutchess County described by Balk (1936) and Barth (1936) and most recently studied by Vidale (1974, a, b), Bence (Bence and McLelland, 1976) and McLelland and Fisher (1976) (Fig. 2).

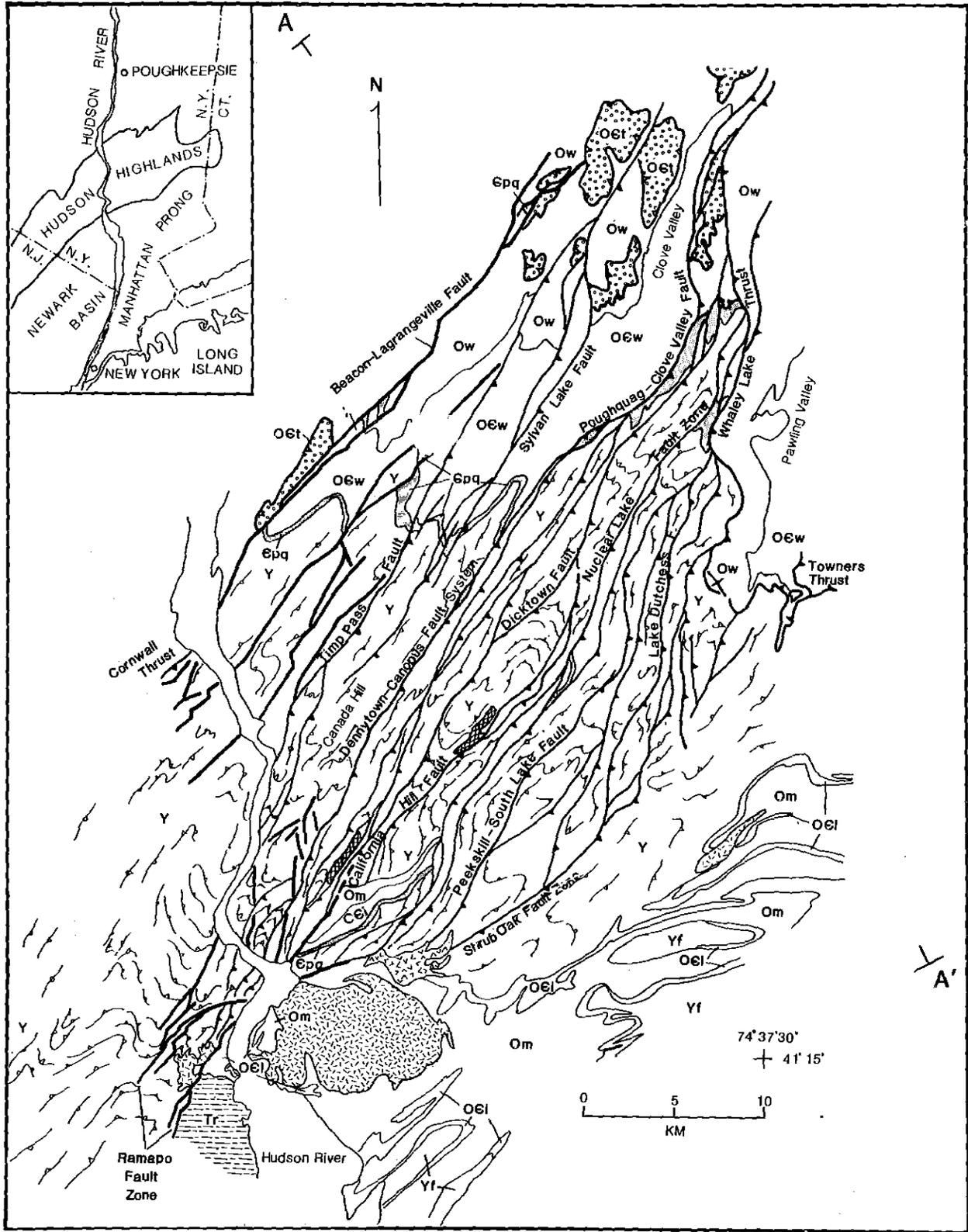
<sup>40</sup>Ar/<sup>39</sup>Ar mineral dates from muscovite, biotite and hornblende from basement rocks and Paleozoic cover rocks in this zonation suggest that the metamorphic zonation is Taconian, perhaps as old as 465 Ma., and not Acadian. (Bence and Rajamani, 1972, Dallmeyer and Sutter, 1976, Sutter and others, 1985).

The shear zones and Middle Proterozoic basement gneiss near shear zones contain new metamorphic assemblages that are consistent with Taconian remetamorphism. Biotite from the California Hill shear zone at staurolite grade gives an <sup>40</sup>Ar/<sup>39</sup>Ar plateau age of 436 ± 3 Ma. (John Sutter, personal communication, 1982).

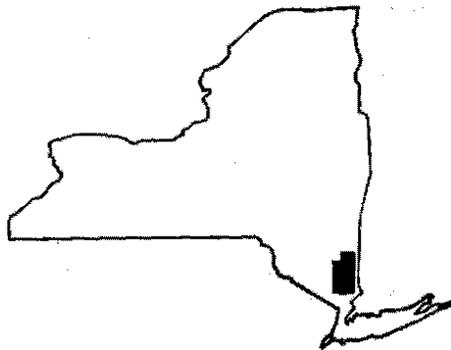
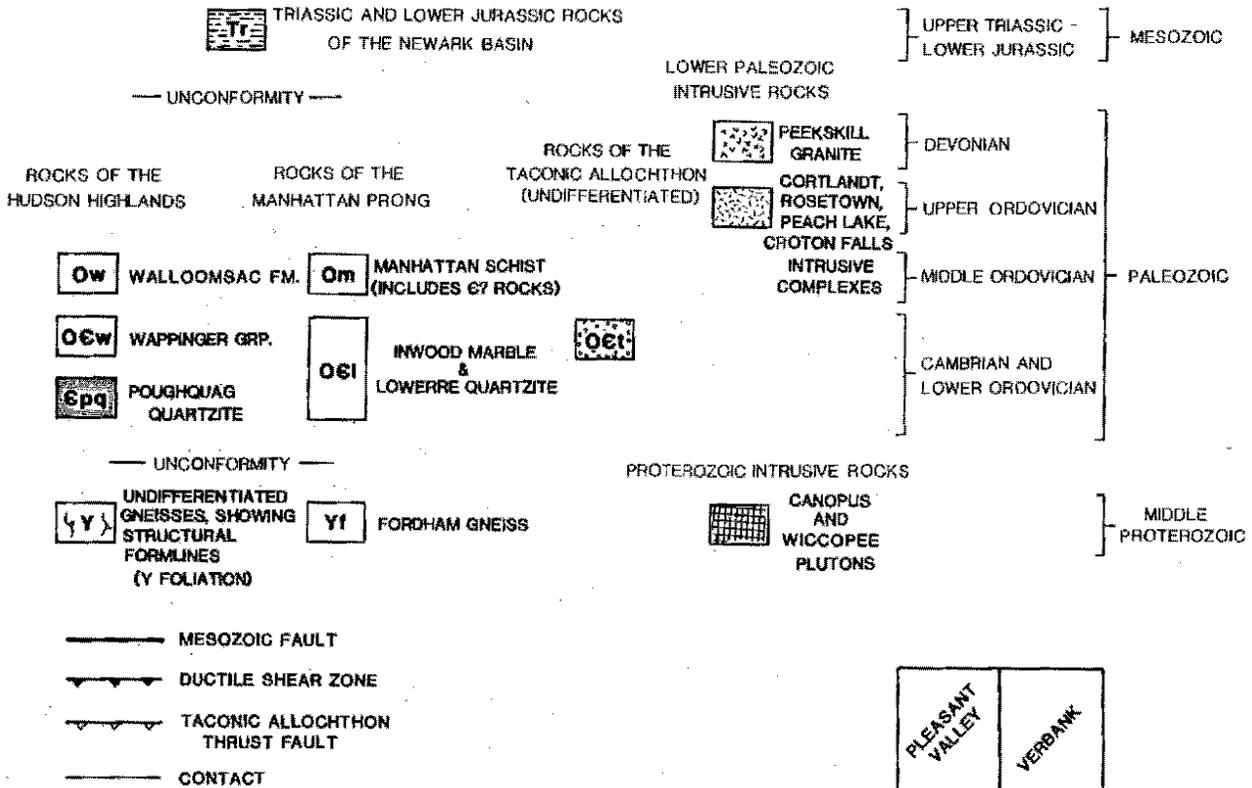
When traced northward these ductile deformation zones pass upwards, as strong S<sub>2</sub> deformation zones, into the Paleozoic cover rocks that contain the classical garnet through sillimanite grade assemblages of Dutchess County. Locally these zones of S<sub>2</sub> deformation are retrogressive and tectonic fore-shortening of the isograds in the cover rocks is indicated from structural studies.

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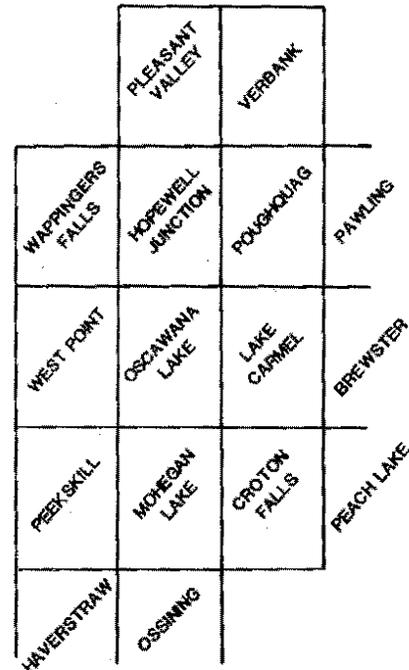
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EXPLANATION FOR FIGURE 1



LOCATION OF FIGS. 1,2,3



QUADRANGLE INDEX MAP FOR FIGS. 1,2,3

Figure 1 Simplified regional geologic map of the northern end of the Hudson Highlands showing semiductile deformation zones. Formlines show prominent regional F<sub>2</sub> foliation of Proterozoic age in basement rocks, barbs show dip.

At the time Vidale published her paper on vein assemblages (Vidale, 1974, a, b), she recognized that the occurrence of index minerals and critical assemblages was quite irregular and probably affected by structural and or retrogressive complications. She therefore did not publish her mineral assemblage data.

We have reexamined all of Rosemary Vidale's unpublished mineral assemblage data and her extensive thin section collection from this area as well as our own oriented sections from shear zones in the cover rocks in an attempt to relate the distribution of isograds to structural evolution of the area. We believe the data indicate that the pattern of metamorphic isograds in Dutchess County was strongly affected by faulting and deformation resulting from accumulation of strain in the quartzofeldspathic basement rocks arising from failure along semiductile shear zones. We relate strain softening of the basement rocks, oversteepening of the isograds and final tectonic emplacement of the isograd-containing rocks to a single but dynamically changing dynamothermal regime of Taconian age (Ratcliffe, 1983).

We have mapped basement and cover rock in detail in order to delimit zones of faulting; however, the lithic successions and details of the mineralogy of protoliths and their relationships will not be stressed on this trip. Those interested in the regional distribution of shear zones of this system exposed to the south in New Jersey may find a discussion and maps in Ratcliffe (1980). Comparable features from the Berkshire massif in Massachusetts have been described by Ratcliffe and Harwood (1975).

The purpose of this trip is to examine the structural details of the semiductile shear zones, associated folding, and mineralogy of the shear zones at various grades from chlorite through sillimanite grade and to relate these features to structures found in the cover rocks in Dutchess County. The route and stops are identified in Figure 3.

Finally we would like to thank John Rodgers who first introduced us in the 1960's to the "chicken and egg argument" regarding the role of metamorphism in controlling of styles of deformation in basement rocks. We are not certain which one is the egg in this case, but it seems clear that the form and style of the structures and the metamorphic patterns are interrelated in a very complex fashion.

### Regional geology

Middle Proterozoic gneiss that was strongly deformed and metamorphosed to hornblende or pyroxene granulite grade during the 1,000 Ma. "Grenville event" constitute the core of the Hudson Highlands and the basement (Fordham Gneiss) of the Manhattan Prong. Late Proterozoic, largely undeformed metadiabase dikes locally provide good strain markers within the basement. Lower Cambrian through Middle Ordovician cover sequence rocks locally are preserved as infolds within the Hudson Highlands, providing additional information for estimation of Paleozoic folding and metamorphic grade of remetamorphism of the basement rocks. The cover sequence in ascending order consists of: Lower Cambrian Poughquag Quartzite, Lower Cambrian through Lower Ordovician Wappinger Group (dolostone and limestone), Middle Ordovician Walloomsac Formation, (carbonaceous schist, quartzite and aluminous schist) and rocks of the Taconic allochthons (aluminous schist and phyllite) ranging from Late

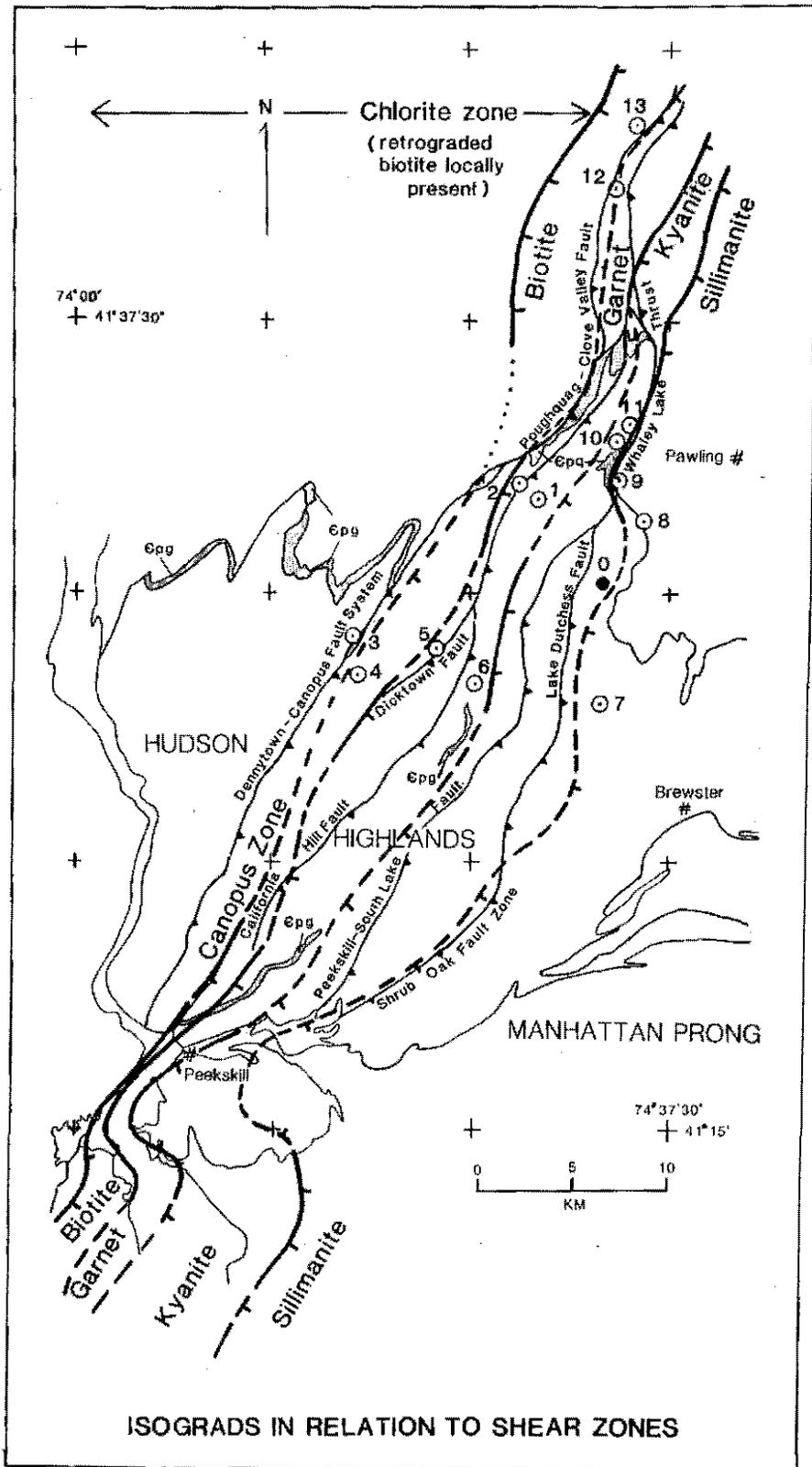


Figure 2 Simplified regional map showing isograds of Taconic age in relationship to shear zones in basement and cover rocks.

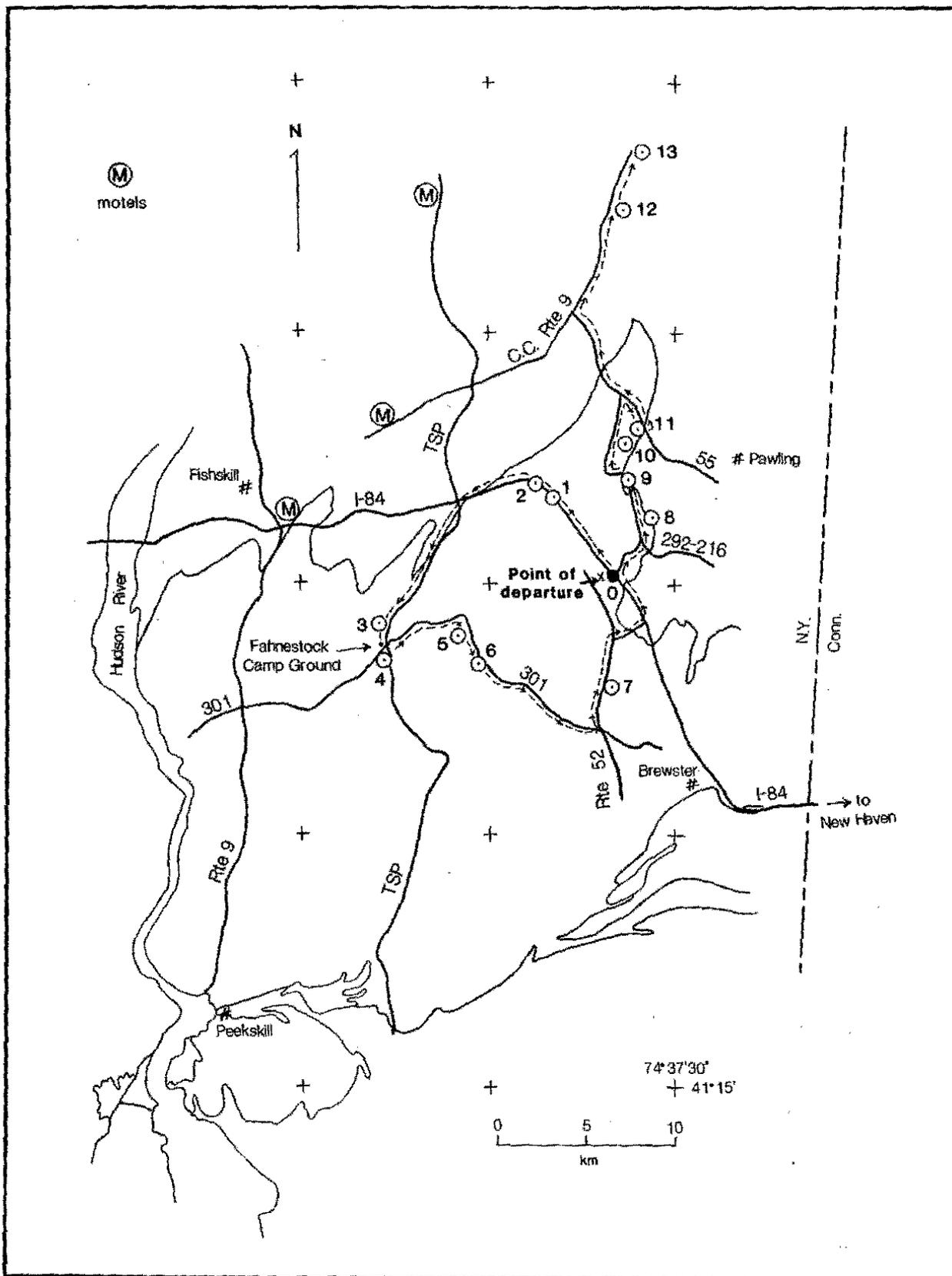


Figure 3 Route of field trip and stops.

Proterozoic through Middle Ordovician. Correlative rocks of the Manhattan Prong consist of: Lowerre Quartzite, Inwood Marble and Manhattan Schist. The Manhattan contains units correlative with the Walloomsac Formation as well as several possibly allochthonous units similar to, but not identical with rocks of the Taconic allochthon (Ratcliffe, 1968, Hall, 1968). Plutonic igneous rocks and dikes of the Cortlandt, Rosetown, Croton Falls, and Peach Lake intrusives of probable Late Ordovician age (Ratcliffe and others, 1983) cross-cut schistosity and Paleozoic shear zones in metamorphic cover rocks and basement in both the Manhattan Prong and Hudson Highlands. These rocks were intruded into the already highly deformed and tectonically thickened basement rocks near the end of the Taconic orogeny (Ratcliffe and others, 1983).

Formlines within Middle Proterozoic gneiss (Fig. 1) outline the attitude of the prominent foliation and gneissic layering ( $F_2$  foliation) of the Proterozoic deformation. Discontinuities marked by mylonitic rocks are identified as solid heavy lines with teeth that show dip direction of the shear zones. Section A-A' (Fig. 4) shows our interpretation of the relationship of shear zones to isograds.

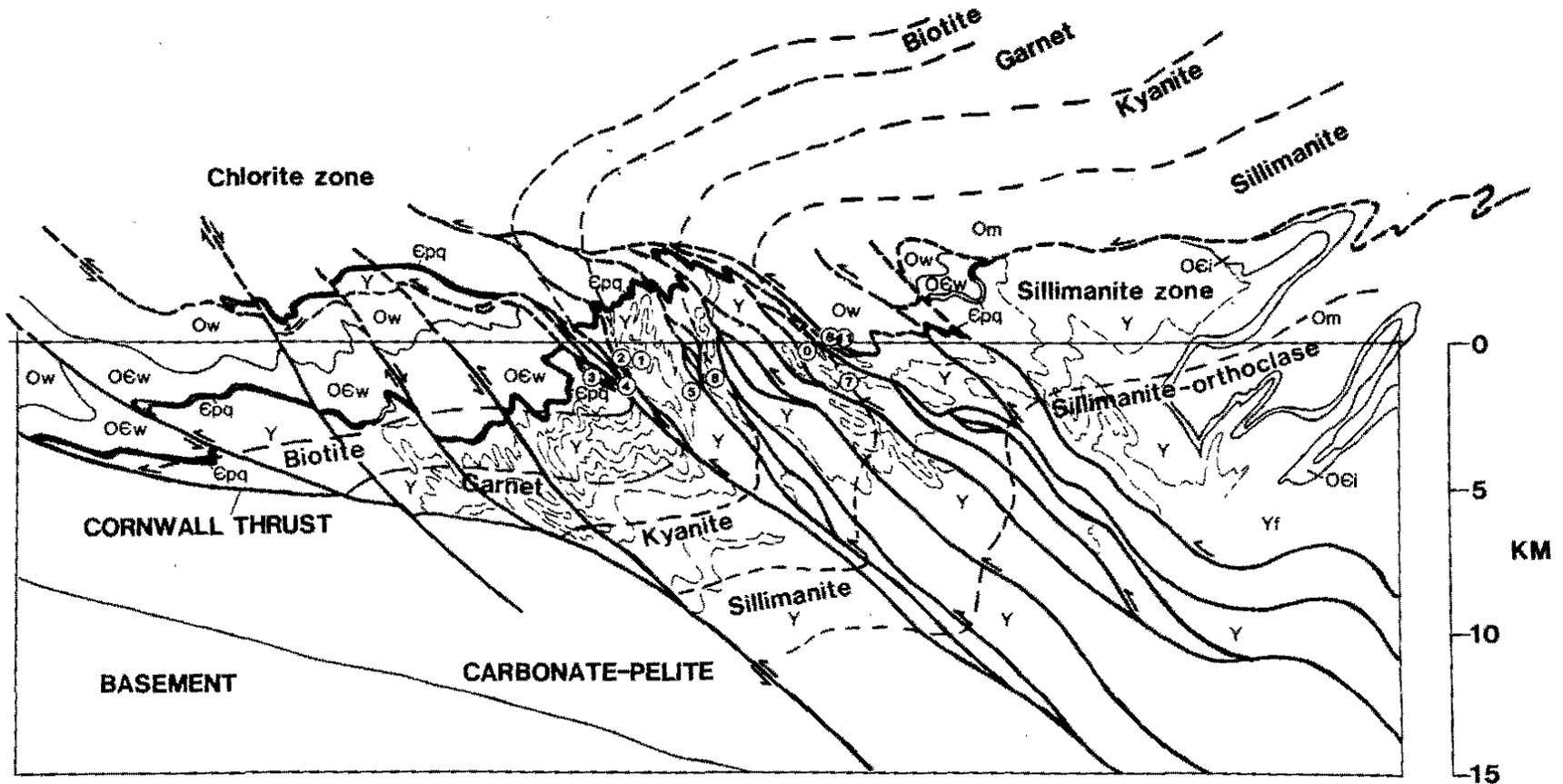
### Structural Terminology

Fault and fault zone rocks described in this guidebook are classified using the terminology of Sibson (1977). Semiductile faults, as opposed to brittle and ductile faults, are faults in which discrete dislocation is discernible and both ductile and brittle deformation characteristics are present as expressed by minerals with different ductilities. The protoliths for all of the fault-zone rocks in basement are moderately coarse-grained gneisses, containing hornblende-granulite-facies assemblages, but the bulk of the deformation in the shear zones was accomplished under metamorphic conditions much less intense than that of the formation of the protoliths. At low grades (chlorite and biotite zone) retrogression is pronounced and the final products are fine-grained diaphthoritic mylonites, that is, phyllonites. At higher grades, (garnet to kyanite grade) protomylonite, mylonite, mylonite gneiss and ultramylonite are common. At sillimanite grade coarser-grained fault rocks marked by coarse biotite, complete recrystallization and annealing of quartz ribbons, and triple-junction grain contacts are common although the unmistakable fault structures are present. We prefer the term blastomylonite for these rocks in order to distinguish these rocks from mylonitic rocks that preserve subgrain shapes formed by dynamic recrystallization. The sequence of fault-zone rocks described here thus range from phyllonite, through mylonitic gneiss, augen gneiss, protomylonite, and mylonite, to blastomylonite. With increasing metamorphic grade the ductile response of the basement rocks increases and fault-zone rocks change from having relatively brittle deformation characteristics to totally ductile fabrics. This change in structural style becomes pronounced in the transition zone between staurolite-kyanite zone and sillimanite zone and corresponds with the onset of dynamic recrystallization of plagioclase and of microcline in quartzofeldspathic rocks.

Folds associated with shear zones commonly have curved hingelines, and are purse-or sheath-shaped. They commonly pitch down the dip of either the schistosity "s" or "c" surfaces. Such folds are termed reclined folds (Rickard, 1971). The reclined nature of these folds is important because the hingelines in many exposures parallel the strong lineation or mullion struc-

# ISOGRADS IN RELATION TO SHEAR ZONES

## DUTCHESS CO. STEEP GRADIENT



### CROSS SECTION ACROSS HUDSON HIGHLANDS, N.Y.

Figure 4 Generalized cross section along approximate line of section A-A' in Figure 1, showing regional shear zones in relationship to isograds. Cross section was prepared from an earlier version of the map (Fig. 1), and contacts do not all agree precisely. Approximate locations of field trip stops are shown.

ture in the mylonite zones interpreted as the elongation lineation.

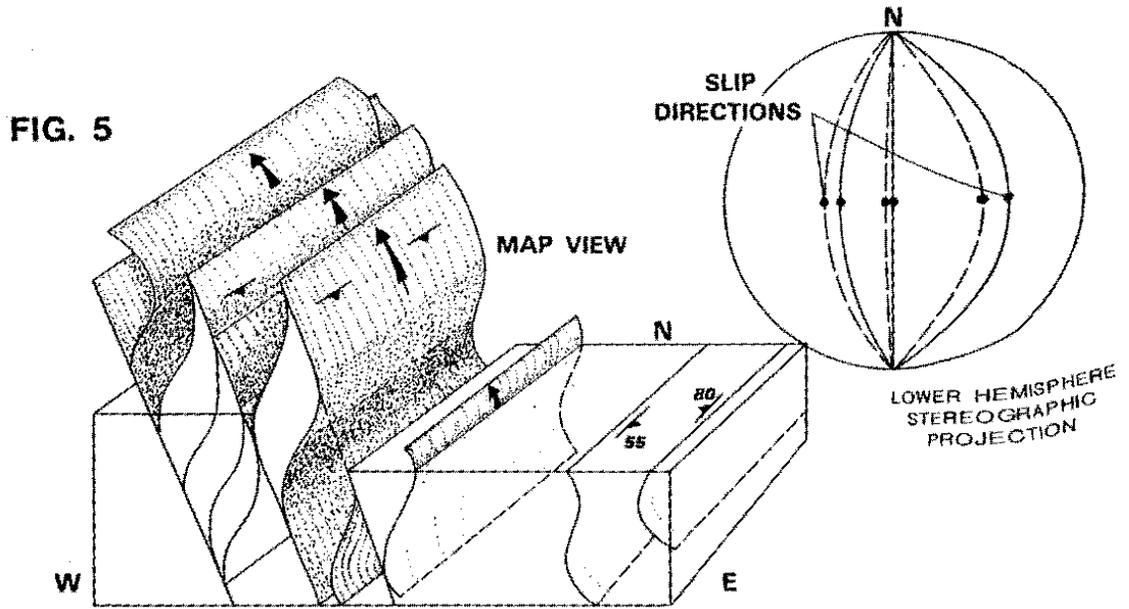
Textures seen in outcrop and in thin section are: "s", a schistosity axial planar to folds in the protolith near shear zones; "c", shear or mylonitic foliation parallel to the mylonite zones; and "sb", shear bands expressed as spaced dislocations in the mylonitic "c" fabric. Larger augen or porphyroclasts, principally of K feldspar, commonly show excellent retort structure in which tails of recrystallized materials or metamorphic products of the augen define the sense of displacement. Plagioclase crystals in some exposures exhibit brittle fracture, and "card deck" displacements, in which microdisplacements have a sense reversed to that of the bounding shear zone. These textures and structures and their kinematic significance are neatly described by Simpson and Schmid (1983). Photomicrographs are given in Plates 1-5.

Following the techniques outlined by Hansen (1971) the rotation sense and plunge of minor folds with amplitudes of 1 to 2 cm and wavelengths less than a centimeter within the mylonite fabric "c" are useful in determining tectonic transport. These minor folds commonly have curved hinge lines and are minor sheath folds as originally described by Hansen. Measurements from narrow zones are consistent and repeatable resulting in unusually narrow separation angles. Treatment of larger amplitude folds across an outcrop is less useful. The strong lineation on the mylonite surfaces, i.e. within the "c" surfaces, are rods or mullion-like structures related to the intersection of "c" and "s" or "sb" and are substantially parallel to the slipline as determined from the separation-angle technique.

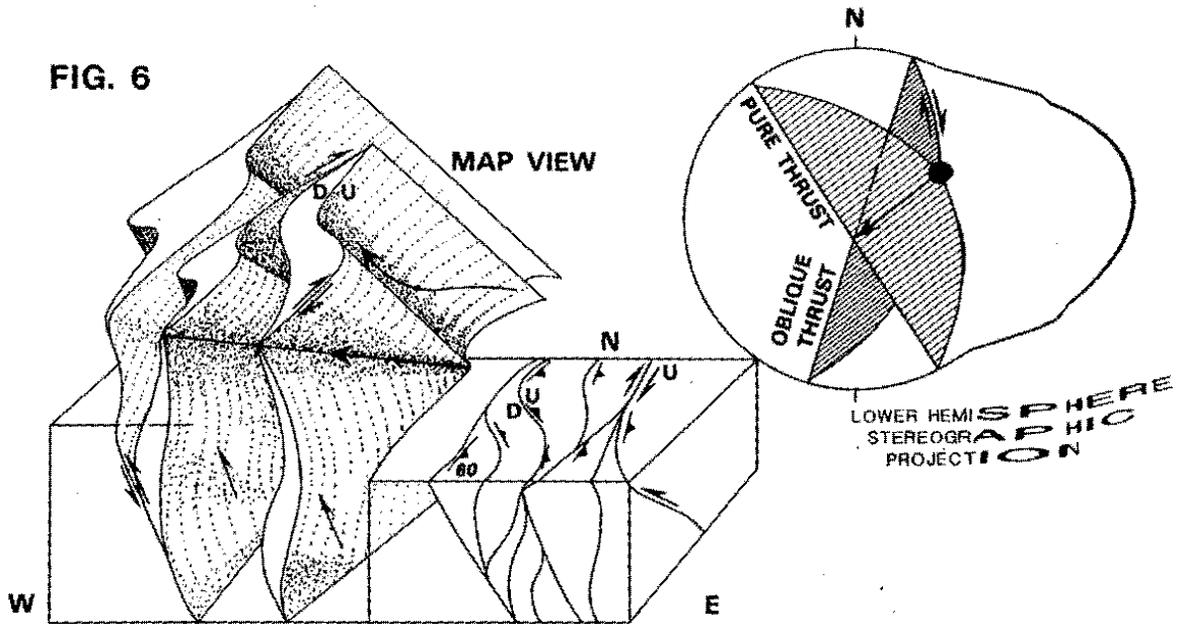
Throughout the area samples of phyllonite and mylonite from shear zones were collected and sectioned in a direction normal to "c" and parallel to the strong elongation lineation, and in a section normal to the lineation. The growth fabrics of fault minerals and protolithic minerals were noted. The results indicate that  $2M_1$  muscovite, chlorite and epidote grew in low grade rocks, followed by biotite, garnet, hornblende, staurolite-kyanite and finally by sillimanite, with microcline, and muscovite. The minerals are progressively stable within the shear zones (in rocks of suitable composition) in a zone approximately colinear with the mineral zonation in Paleozoic cover rocks of Dutchess County (Vidale, 1974a, b) Bence (in Bence and McLelland, 1975), and in the Manhattan Prong (Ratcliffe and others, 1983).

#### Description of the shear zones

A ten-kilometer-wide zone of closely spaced anastomosing shear zones transects the basement rocks east of the Dennytown-Canopus fault system (Fig. 1) based on our detailed 1:24,000 mapping. Shear zones 0.5 meter to 100 meters thick trend N-S to N.  $40^\circ$  E forming right-lateral sigmoidal patterns. Dips range from near vertical to  $45^\circ$  S.E. In cross section the shear zones also have sigmoidal forms with southeast-side-up sense of displacement. The entire package of shear zones is limited at its upper surface by a major thrust, the Whaley Lake thrust, that transports inverted Paleozoic cover-sequence rocks over gneiss. To the southeast this thrust is traceable to the area of the Towners thrust (Fig. 1). The floor of the package is the Canopus-Dennytown fault system that can be traced northward into the Poughquag-Clove Valley fault that places cover sequence rocks at garnet and higher grade over carbonate rocks of the Wappinger Group. The assemblage of



**SIGMOIDAL THRUST FAULT CONFIGURATION  
OF DUCTILE SHEARS  
(SLIP DIRECTION UP-DIP)**



**OBLIQUE THRUST FAULT CONFIGURATION  
OF DUCTILE SHEARS  
(SLIP DIRECTION, OBLIQUE RIGHT-SLIP)**



shear zones can be viewed as floor and roof of a complex duplex system with intense internal deformation. The Poughquag Quartzite at the base of the cover sequence is preserved in a series of fault-bounded synforms that record the passive style of the thrust-related folding.

The block diagrams in Figures 5, 6, and 7 portray conceptually the relationship of the semiductile shear zones to the folds in the cover sequence. In Figure 5 the map and cross section of normal thrust ramps are portrayed and the elongation lineations (mullions and sliplines) expected illustrated. In Figure 6 shear zones having sigmoidal plan views as well as cross-section form are illustrated. Slip directions for such faults are illustrated. This pattern is thought to represent the Hudson Highlands shear zones. Folding of the cover sequence into synforms formed over basement rocks deforming by such a sigmoidal oblique-thrust pattern is illustrated in Figure 7. Of particular note is the trend of folds in a more northwesterly direction than the trace of the master shears. The complex series of mylonite-related foliations and folds might normally be ascribed to multiple deformational episodes  $F_1$ ,  $F_2$  etc. but we believe that they are more logically explained by a single fold-thrust model described by Figures 6 and 7.

#### Sliplines and movement sense of shear zones

Slipline determinations after the technique of Hansen (1971) show a slip direction of approximately  $N. 77^\circ E.$  regardless of the strike of the shear zone, showing that the curvilinear traces of the shear zones are primary rather than folded. Additional data from west and south of the area shown in Figure 8 show sliplines ranging from due E. to  $N. 70^\circ E.$  The lineations mapped by Balk (1936) are approximately parallel to this lineation and date from the fold-thrust event.

#### Relationship of fold-thrust structures to folds and foliations in the cover rocks

The fold and thrust fabrics, and lineations related to the shear zones, represent at least the second dynamothermal event seen in the Paleozoic rocks (Fig. 9). Cover-sequence rocks attached to the basement gneiss contain a strong foliation or schistosity substantially parallel to bedding and contain beds highly folded about recumbent to inclined axial surfaces. These first generation folds are deformed in the shear zones into tight second generation  $S_2$  folds. Similarly, cover rocks above the Whaley Lake thrust contain folds and schistosity older than the thrust fabric. Near thrust faults the structural overprinting by the  $S_2$  (thrust-related) fabric is strong. This post-schistosity fabric continues northward into the cover-sequence rocks east and west of Clove Valley, where it is expressed locally by zones of phyllonite.

#### Metamorphic assemblages in shear zones and regional isograds

Previous workers do not all agree on the location of metamorphic isograds. Vidale (1974a, b) moved Barth's (1936) isograds further west than previously shown but Bence's isograds (in Bence and McLelland, 1976) differ markedly from Vidale's results. Additional data and reexamination of all of Vidale's thin sections has resulted in a minor modification (Fig. 9). Biotite first occurs west of Clove Mt. in the Pleasant Valley quadrangle as sporadically developed small crystals growing across a foliation but almost always

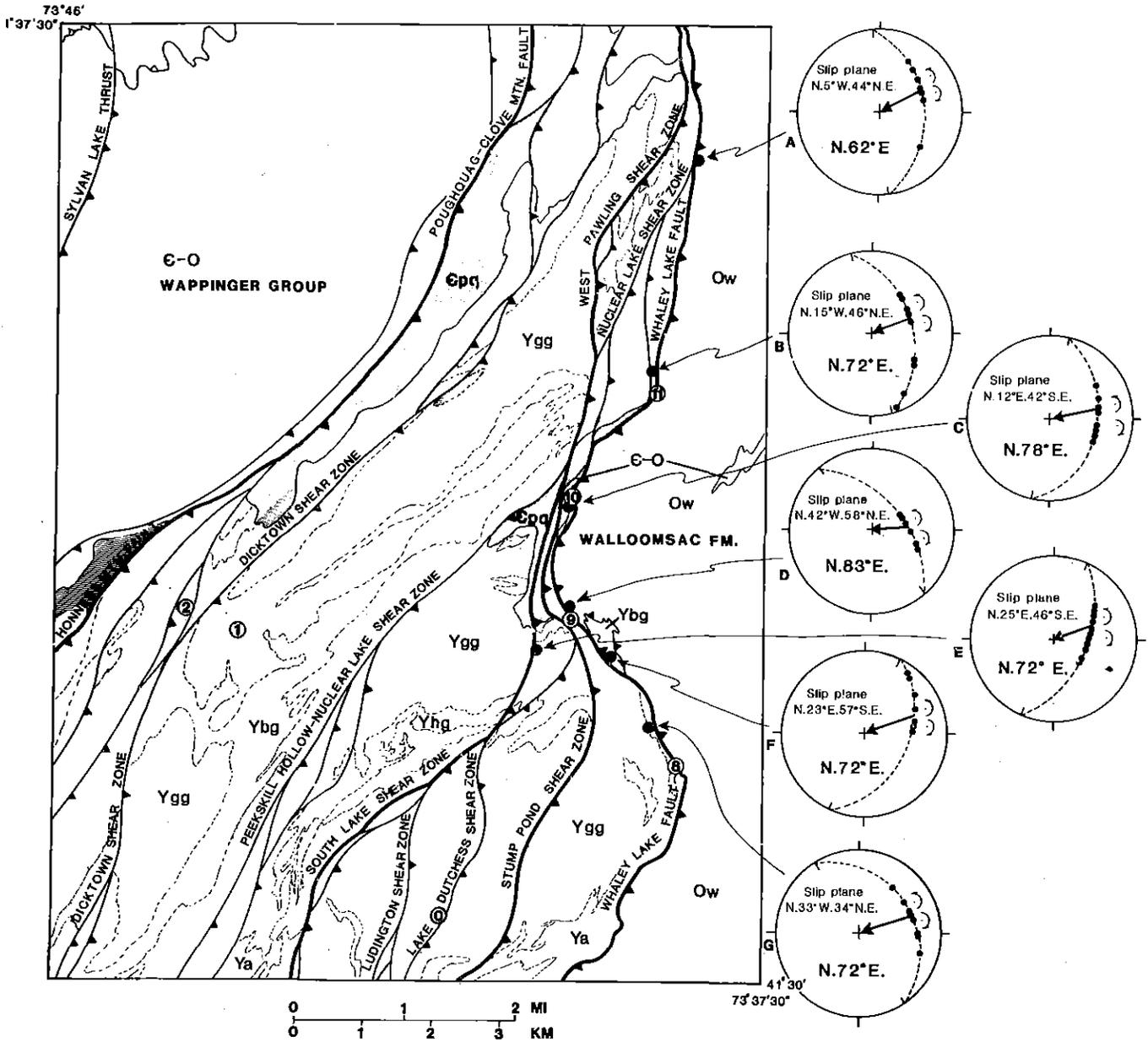


Figure 8 Generalized geologic map of the Poughquag Quadrangle showing slipline determinations from shear zones, outlines of marker units in Middle Proterozoic basement rocks, and field trip stops. Key to symbols: Ya, amphibolite, Ygg, granitic biotite gneiss, Ybg, biotite feldspar gneiss and paragneiss, Epq, Poughquag Quartzite, e-o, Wappinger Group, Ow Walloomsac Formation.

partially retrograded by a later foliation (Vidale, 1974a). East of Clove Valley two generations of biotite are widely developed: (1) as large post-schistosity ( $S_1$ ) non-oriented porphyroblasts that include schistosity and (2) as small, fine-grained crystals intergrown with muscovite and/or chlorite or as small isolated flakes in a second-generation foliation associated with  $S_2$  shear zones.

Chloritoid is strongly dependent upon the highly aluminous composition present only in the Taconic allochthonous rocks. It also occurs sporadically as post- $S_1$  crystals that are retrograded in  $S_2$  in rocks at the northern end of Clove Valley and along the foot of East Mountain. Above garnet grade chloritoid occurs as robust post- $S_1$  and as post- $S_2$  crystals.

Garnet first occurs at the north end of Clove Valley in isolated outcrops and again along the west foot of East Mountain east of a prominent  $S_2$  shear zone. Garnets are compositionally zoned (Bence and McLelland, 1976) and euhedral in rocks west of the Pleasant Ridge shear zone, but east of this, in staurolite-zone and higher grade rocks, have deeply corroded cloudy cores and clear but irregular overgrowths. Garnet is post- $S_1$  in most exposures but is syn- $S_2$  in rocks east of the Pleasant Ridge shear zone (Fig. 9). Although the staurolite isograd appears to be offset by the Pleasant Ridge shear zone, tiny euhedral staurolite, garnet and biotite appear as syntectonic minerals in  $S_2$  fabric in mylonite of this shear zone. This suggests that synmetamorphic recrystallization accompanied rapid uplift along the shear zone.

Sillimanite first occurs as fibrolite tails on kyanite kinked in  $S_2$  shear zones, or on biotite in the same fashion. The isograd for sillimanite, parallel to the Whaley Lake thrust, is controlled by these reactions in  $S_2$  just above the fault. Further east, away from the fault, staurolite and kyanite without fibrolite are widely present. A small area of fibrolite-bearing rocks is also present in strongly sheared kyanite-staurolite and kyanite schist along the Pleasant Ridge shear zone. On Figure 9, the deflection in the sillimanite isograd is produced by the preferential growth of fibrolite in Walloomsac schists having a strong  $S_2$  fabric near the Whaley Lake thrust. Here the sillimanite isograd may have been controlled kinetically by conditions in the dynamically recrystallizing rock.

From west to east in the basement rocks muscovite, epidote, chlorite, biotite, garnet, tremolite-talc, hornblende, staurolite, kyanite, sillimanite and microcline have grown in shear zones. Clear growth in the elongation direction indicate syntectonic crystallization in many specimens. Based on these observations re-metamorphism of the basement rocks and shearing were contemporaneous. In the cover rocks the observations differ.  $S_2$  shear zones with the same prominent N.  $70^\circ$  to N.  $80^\circ$  E. lineation are retrogressive and destroy previously formed biotite in areas west of Clove Valley and along the base of East Mountain (Stops 12, 13, Fig. 9). Rocks in the highlands on East Mountain and cover rocks south along the Whaley thrust contain clear indications of syntectonic thrusting and crystallization of sillimanite, kyanite, staurolite and biotite and garnet.

Mineral assemblages within the cover rocks affected by the shear zones indicate that the isograds for staurolite, kyanite and sillimanite trend more northeasterly than the shear zones (Fig. 9). However, mineral assemblages and growth fabrics indicate peak temperature assemblages are shared between foot-

DISTRIBUTION OF METAMORPHIC MINERALS  
AND GEOLOGY

DISTRIBUTION OF S<sub>1</sub> (F<sub>1</sub>) STRUCTURES  
AND S<sub>2</sub> SHEAR ZONES

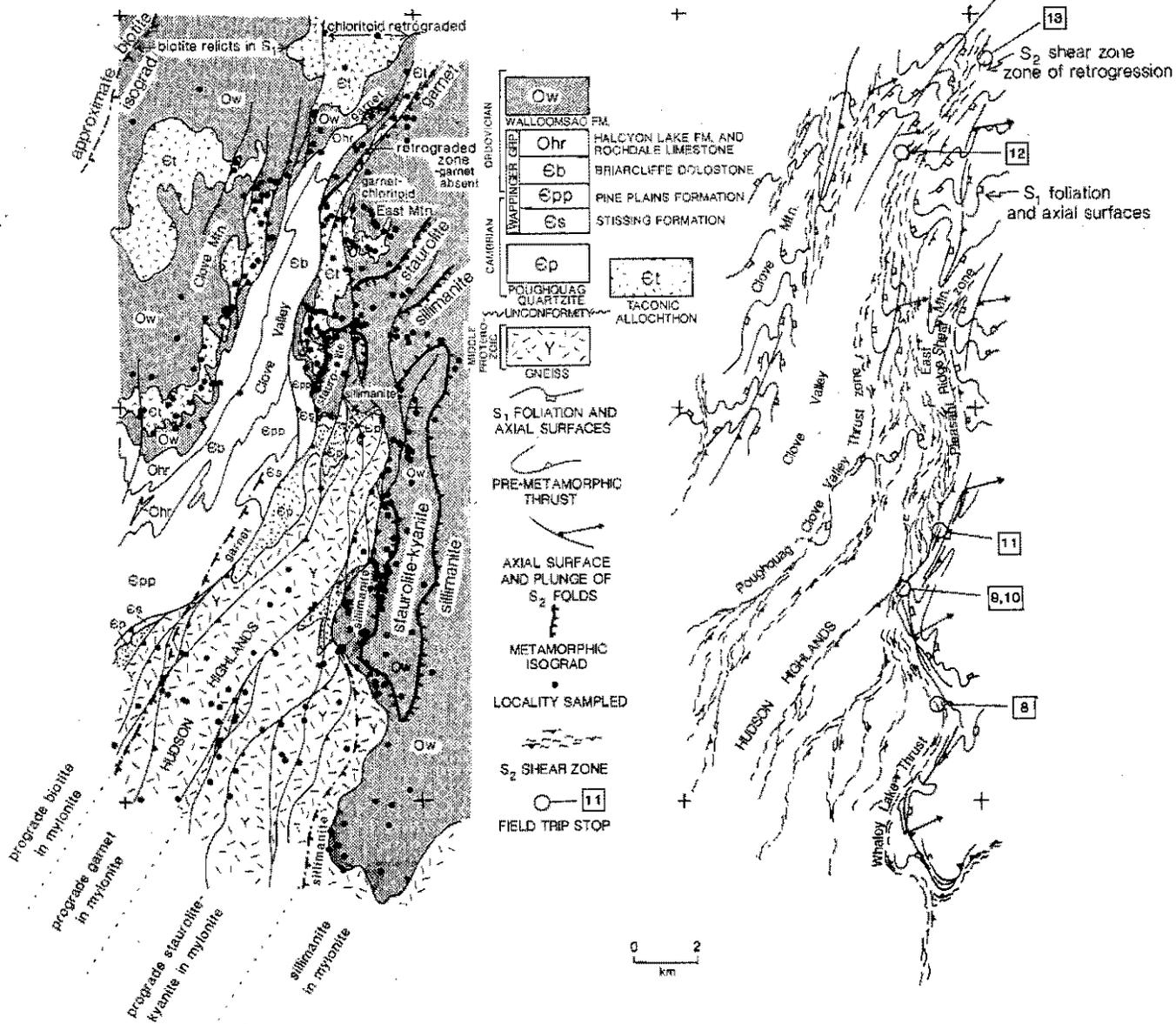


Figure 9 Generalized geologic map of the Verbank and Poughquag Quadrangles showing S<sub>2</sub> shear zones in cover rocks, metamorphic isograds, and pockets of relict (pre-S<sub>2</sub>) minerals. Isograds have tick marks on higher temperature side.

wall and hanging wall blocks of the shear zones at high grades. The mineral data therefore suggest that the peak metamorphic conditions in the high grade rocks formed synchronous with the semiductile faulting but that similar-style  $S_2$  faulting postdated peak conditions in the lower-temperature cover rocks, along the Poughquag-Clove Valley fault (Fig. 9) and to the west. Examination of the cross section along line A-A' (Fig. 4) shows that the metamorphic zonation from biotite to sillimanite is too narrow, less than 5 km wide (Fig. 9), to be accounted for by shallow-dipping regional isothermal surfaces, and steeply inclined isograds are required. The close correspondence between index minerals in the cover rocks and the shear zones in the basement rocks suggests that the isogradic surfaces were deformed by the same episode of ductile deformation that produced the shear zones in the basement.

The model we propose, therefore, involves thermal weakening of the quartzofeldspathic basement rock during the  $M_1$  event, uplift by formation of shear zones, strain softening, acceleration of this uplift, and final upward and westward movement of the core rocks and Barrovian assemblages contained in the cover rocks.

### Geochronology

$^{40}\text{Ar}/^{39}\text{Ar}$  data from biotite, muscovite, and hornblende from the basement and cover rocks affected by the Barrovian gradient described here yield ages ranging from 465 Ma to 370 Ma. (Bence and Rajamani, 1972; Dallmeyer and Sutter, 1976).

Basement rocks across the Hudson Highlands have been unevenly retrograded in the Taconic dynamothermal events to form new mineral assemblages. Petrographic data indicate that this retrogression is most pronounced near shear zones. Enclaves of non- or only partially-retrograded rocks exist between the shear zones. K-Ar and  $^{40}\text{Ar}/^{39}\text{Ar}$  geochronologic data from biotite, hornblende  $^{40}\text{Ar}/^{39}\text{Ar}$  ages from the northern Reading Prong and areas within the over-printed areas to the north are shown on Figure 10 (reproduced from Sutter and others, 1985). Included are data from Dallmeyer and Sutter (1976), Bence and McLelland (1976) and a new biotite plateau age of  $436 \pm 3$  Ma from the California Hill shear zone near Stop 6. Sample 11HP was taken near stop 2, 512 BP at stop 11; and 179 BP from cover rocks on East Mountain near stop 12. The biotite sample in the California Hill shear zone was collected from an area where protoliths adjacent to the shear zone yield hornblende and biotite plateau ages of 913 and 710 Ma, and conventional K-Ar ages on biotite are 800 Ma (Dallmeyer and Sutter, 1976 Figure 1).

$^{40}\text{Ar}/^{39}\text{Ar}$  plateau ages of hornblende and biotite between 470 and 377 shown in Figure 10 have been interpreted as cooling ages from a 460-470 Ma, Taconic metamorphic plutonic event (Dallmeyer and Sutter, 1976). The biotite plateau ages from cover rocks at  $444 \pm 3$  Ma (staurolite-kyanite grade), of  $418 \pm 5$  Ma at sillimanite grade, and a  $436 \pm 3$  Ma at staurolite-kyanite grade (California Hill shear zone) support the idea that the remetamorphism and mylonitization are of Taconic age. The regular increase in grade eastward found within the shear zones reported here reinforces the conclusion that the shear zones and deformation related to them are Taconic rather than Acadian or younger.



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Plates - Photomicrographs of shear zone rocks

The following plates contain photomicrographs of thin sections of shear-zone rocks arranged in order of increasing metamorphic grade. Each is approximately 1 cm across, taken in plane polarized light except where crossed nicols noted. References to Plates indicated in margin of road log as: Pl. 1A to read see Plate 1A.

Plate 1. Chlorite and biotite zone phyllonites from Dennytown-Canopus fault system, Stops 3 and 4.

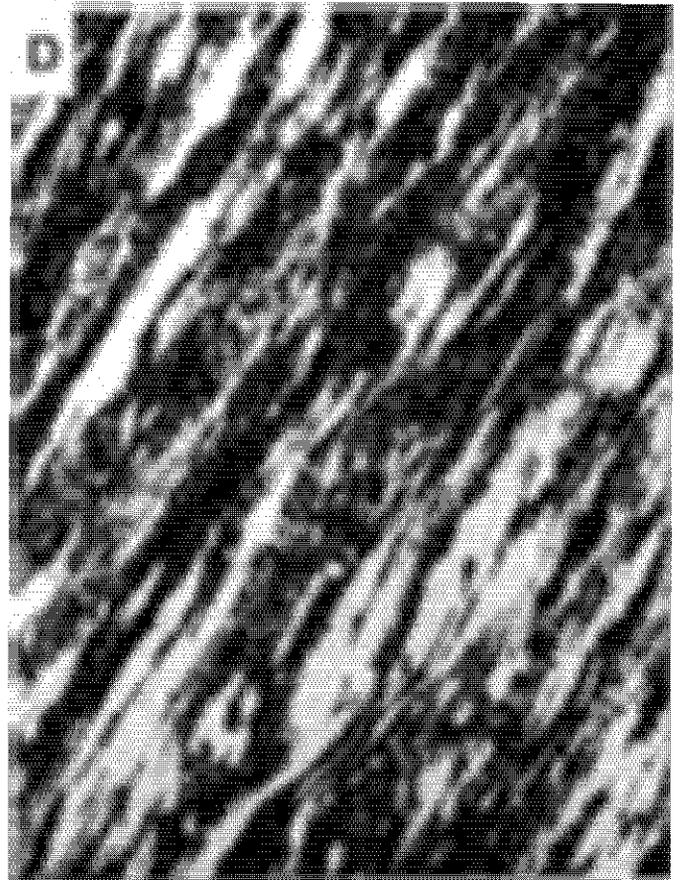
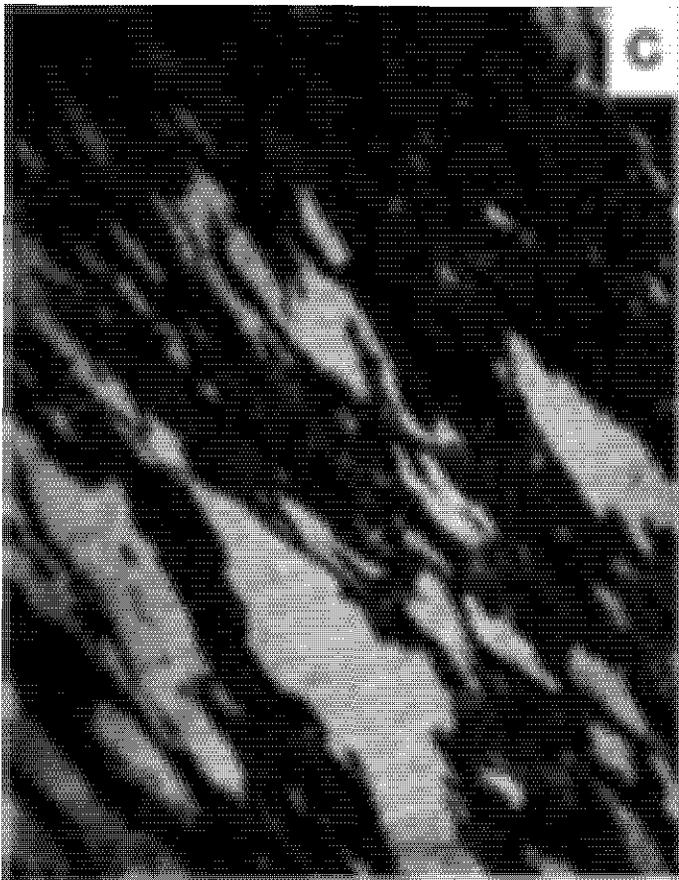
- A. Phyllonite from Proterozoic quartz plagioclase gneiss, Stop 3, section parallel to strike, N.E. at top, S.E. at right, matrix fine grained muscovite, epidote, chlorite, quartz and albite, biotite is shredded and replaced by muscovite, chlorite and opaques, clear sigmoidal flamboids of strained quartz, crossed nicols.
- B. Mylonitic Poughquag Quartzite, Stop 3, section parallel to strike, N.E. at top, S.E. to right, strained quartz pebbles show right sigmoidal shapes, matrix fine-grained muscovite, chlorite and quartz, crossed nicols.
- C. Phyllonite, Stop 4, Canopus shear zone near T.S.P., down dip section, top up, S.E. top right, clear areas folded quartz ribbons consisting of dynamically recrystallized subgrains in a matrix of muscovite, epidote, quartz and new biotite, relict Proterozoic biotite (dark spots) is shredded.
- D. Biotite-rich mylonite in Dicktown fault near Stop 5, horizontal section N. at top, E. at right, mylonitic foliation "c" consists of fine biotite, epidote, albite and muscovite, largely from breakdown of original plagioclase in gneiss, garnet in quartz ribbons are fractured, pulled apart and retrograded to chlorite. N. 50° E. shear bands show right-lateral component.

Plate 2. Biotite and garnet grade mylonite of the Dicktown and California Hill fault zones.

- A. Horizontal section of biotite-oligoclase-rich mylonite of California Hill fault showing right-lateral quartz, oligoclase tails on a large oligoclase porphyroblast, sigmoidal mylonitic "c". N.E. at top, S.E. at right. Biotite from this rock yielded  $436 \pm 3 \text{ } ^{39}\text{Ar}/^{40}\text{Ar}$  plateau age (Sutter, personal communication, 1982).
- B. Vertical section parallel to rods in muscovite mylonite, Stop 2, S.E. to right, showing brittle fracture of plagioclase and microcline in a matrix of ductile quartz, muscovite, biotite, epidote; ribbon of quartz are polygonized. Euhedral garnet present in some samples from this outcrop, garnet isograd.
- C. Vertical section parallel to rods of vertically dipping mylonite in South Lake shear zone near Stop 0, vertical mylonitic foliation top left to lower right corner. Sigmoidal mylonitic foliation shows up from the west sense of motion, matrix, muscovite, biotite and dynamically recrystallized quartz, plagioclase, microcline, and tiny euhedral garnet. (Upper garnet to staurolite-kyanite zone.)
- D. Horizontal section of augen gneiss in Nuclear Lake shear zone near Stop 11, showing right-sigmoidal "c" foliation, crosscutting shear bands "sb", N.E. at top, S.E. at right, matrix muscovite, biotite, and dynamically recrystallized plagioclase, microcline and quartz staurolite-kyanite zone.



Plate 1



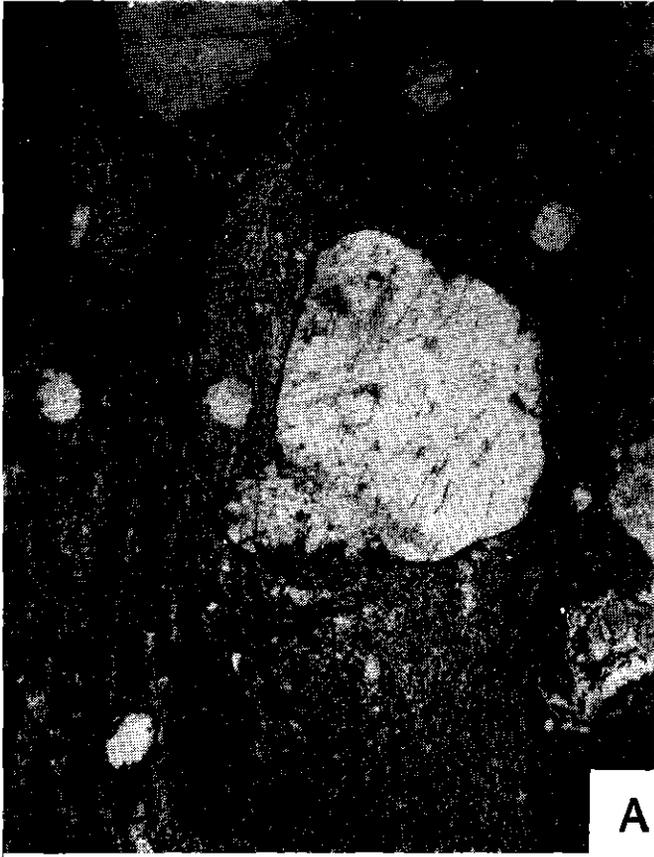
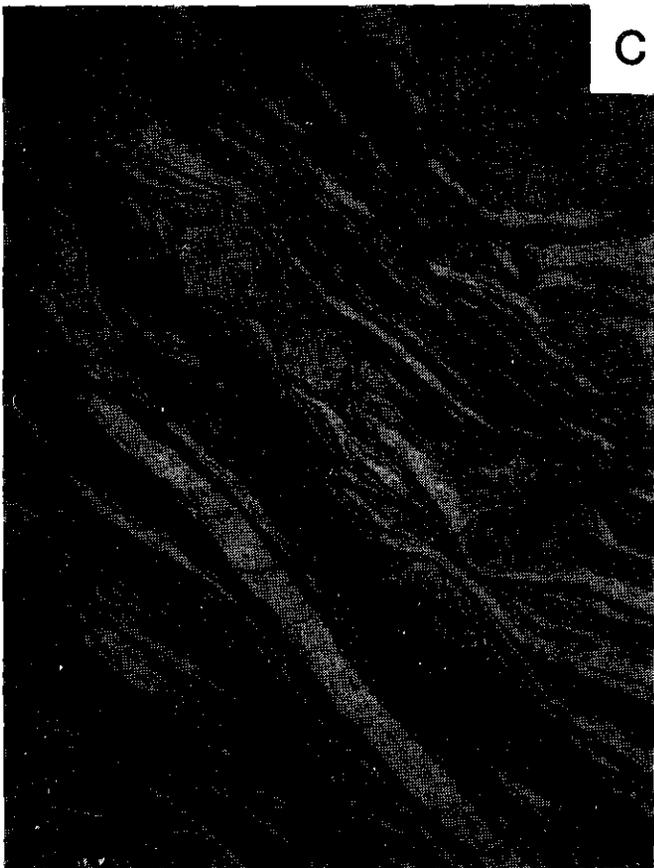


Plate 2



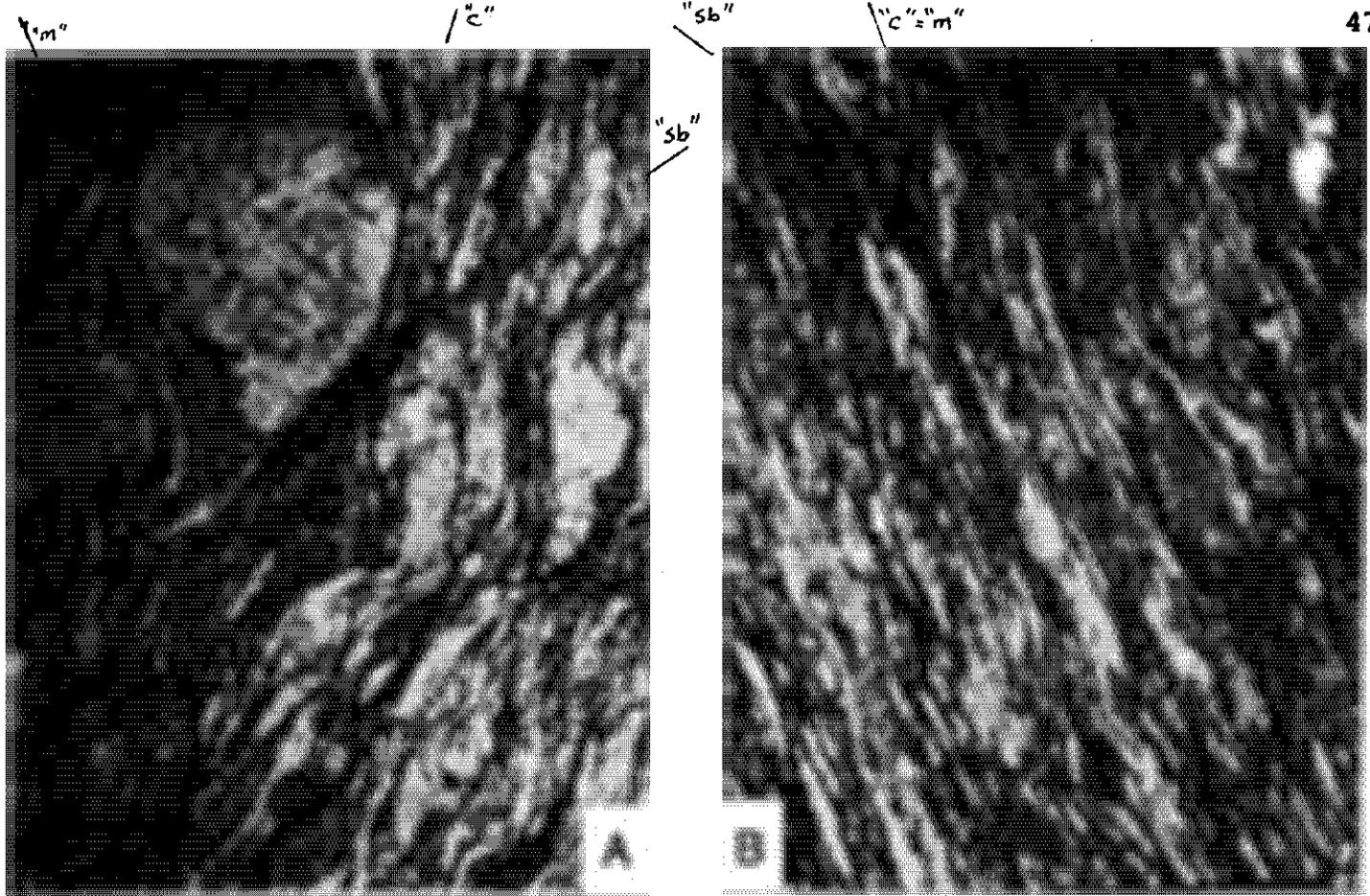


Plate 3



## Plate 3. Mylonites (staurolite-kyanite and sillimanite grade).

- A. Augen gneiss, Stop 10, cut parallel to strike, N.W. at top, N.E. to right. Mylonite matrix muscovite, biotite, plagioclase, microcline surrounding augen of microcline-perthite, ductile flamboids of quartz have annealed texture. Small plications suggest shortening perpendicular to "s", right-lateral sigmoidal "c", and shear bands "sb".
- B. Vertical section parallel to rods, in same rock as A, clear area polygonal quartz in flamboids, microcline light grey, plagioclase clear, biotite and muscovite define "c" and "s", N.E. to right.
- C. Vertical section parallel to rods, blastomylonite at Stop 7, E. to right, ramp structures defined by dynamically recrystallized quartz microcline (grey), and plagioclase with annealed polygonal quartz, muscovite and biotite define schistosity. The loss of all brittle deformation characteristics marks the change to completely ductile, penetrative deformation of the basement gneiss at the kyanite- sillimanite transition.

## Plate 4. Structures in cover rocks. Whaley Lake thrust.

- A. Biotite, garnet, plagioclase, fibrolite, quartz Walloomsac schist with strong  $S_2$  folds and fibrolite growing in  $S_2$  subparallel to thrust. E. to right. Stop 11.
- B. Mylonite in Walloomsac, 6 cm above contact with mylonite gneiss Stop 11, vertical section parallel to rods, and  $S_2$  hinges, polygonal quartz in ribbons, ramp faults, and "augen" of coarse  $M_1$  biotite with syntectonic tails of fine biotite in  $S_2$ .
- C. Mylonitic Walloomsac 1 cm above contact with Pine Plains Formation, Poughquag-Clove Valley fault, showing shredded  $M_1$  biotite and muscovite-rich mylonite, thrust direction from right (N.E.) to left (S.W.).
- D. Mylonite in Walloomsac at staurolite grade along Pleasant Ridge fault, section perpendicular to hingelines of  $S_2$  folds, matrix contains tiny euhedral staurolite, garnet, biotite and muscovite; synmetamorphic thrust fabric.

Plate 5.  $S_2$  fabrics in cover rocks, Stops 12-13.

- A. Mylonitic Walloomsac at north end of Clove Valley near Stop 12, showing shredded  $M_1$  biotite in retrogressive matrix of muscovite, chlorite and calcite, horizontal section.
- B. Phyllonitic Taconic rocks at Stop 13, horizontal section showing low angle intersecting shears. Biotite and muscovite in  $S_2$ , parent rock may have had small garnets--like C below.
- C. Euhedral small garnet, ilmenite, biotite, plagioclase, muscovite, chlorite quartz schist, Stop 12, showing ilmenite oriented in  $S_2$ , and locally bent where passing from garnet into  $S_2$  matrix, where  $S_2$  shearing is strong, near top, matrix is chlorite-rich and resembles D below.
- D. Mylonitic muscovite, chlorite quartz  $S_2$  phyllonite from same outcrop as C; matrix is rich in chlorite and iron-stained chlorite or stilpnomelane, no garnet or coarse grained muscovite present in C remains. Interpreted as a rock retrograded from C along  $S_2$  fabric.

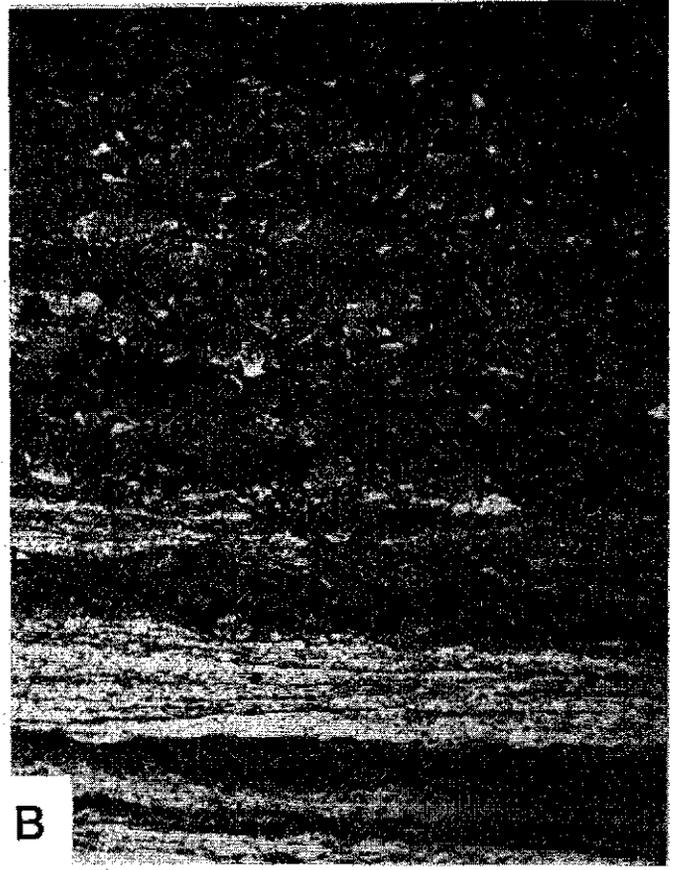
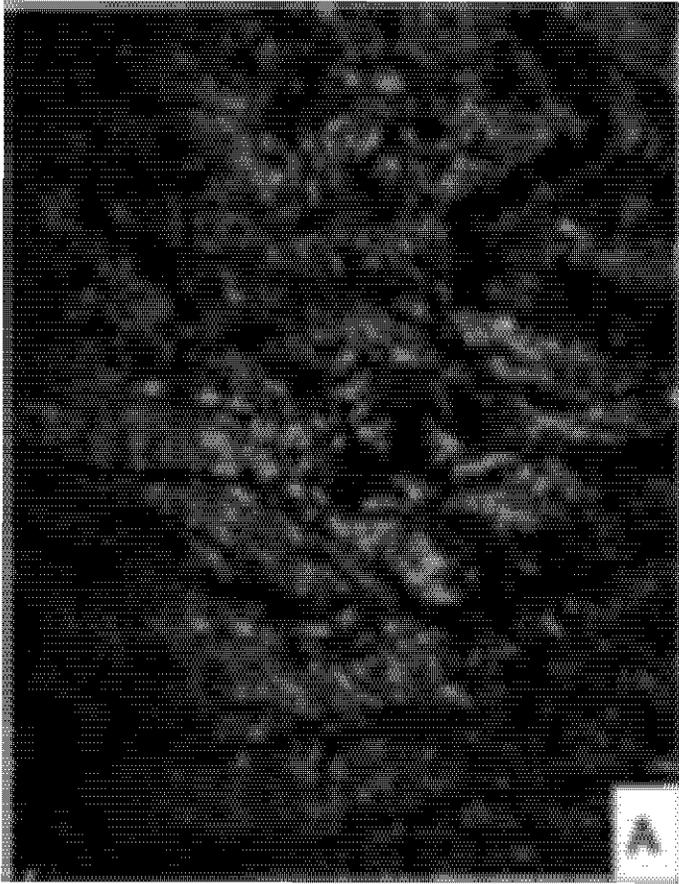
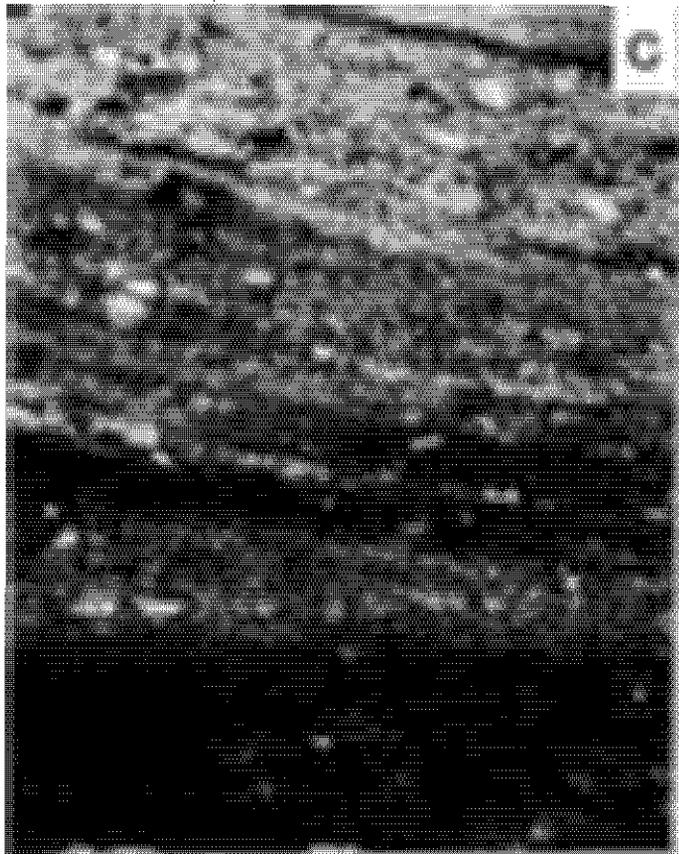


Plate 4



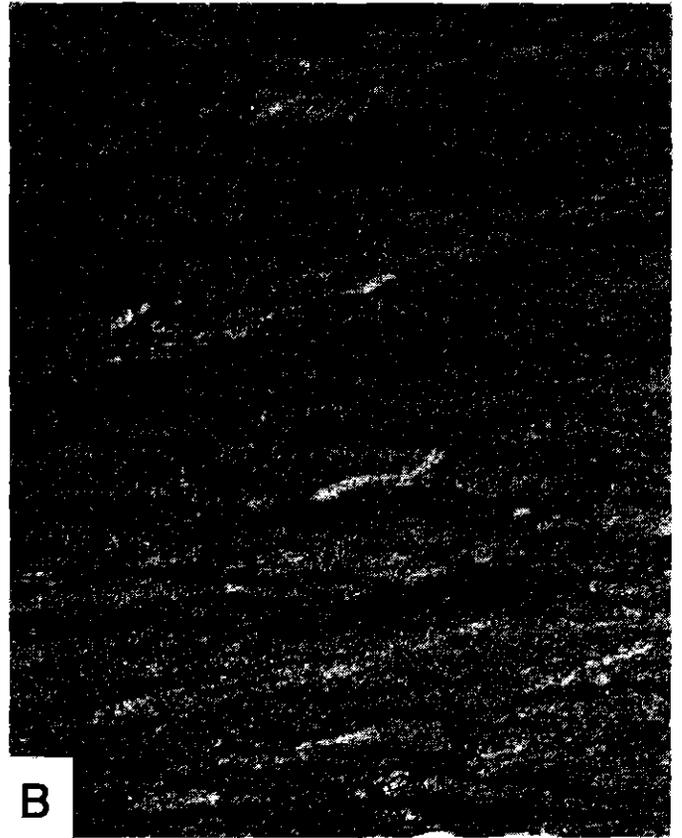
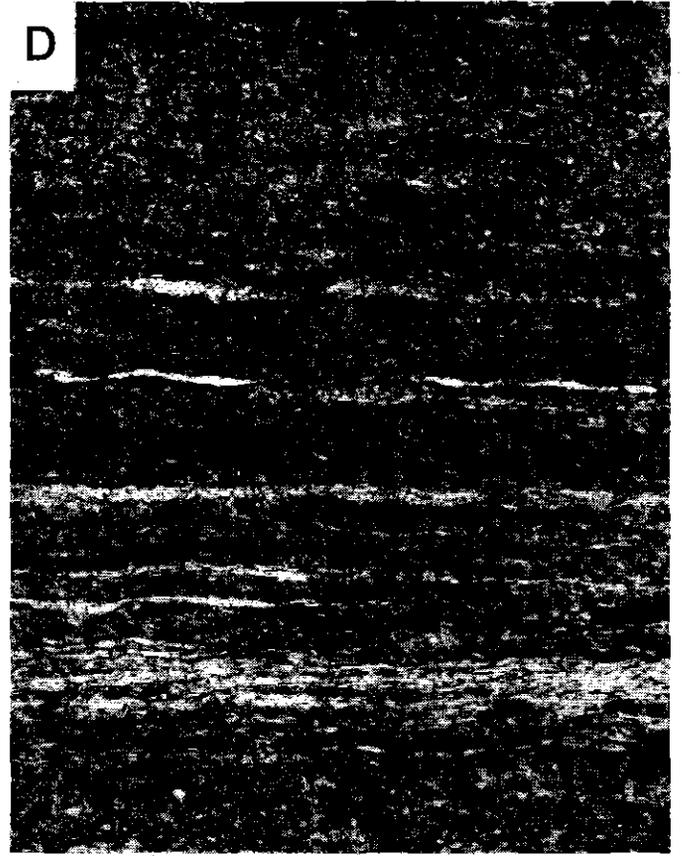


Plate 5



Road log (see Plates for photomicrographs)

(For route and stop locations refer to Figure 3.)

Starting point: entrance ramp to westbound lane of I-84 (Exit 17) at Route 52, Ludingtonville, N.Y. in the Poughquag quadrangle. Introduction 8:00, leave 8:30. References to plates pertinent to Stops given in margin.

Stop 0: Introduction at first outcrops in westbound lane to I-84 at Ludingtonville, staurolite-kyanite zone, Poughquag quadrangle (Figs. 2,8) (30 min).

The trip begins within the interior of the basement massif in rocks strongly affected by Paleozoic remetamorphism. A small but well-developed shear zone in granitic gneiss and amphibolite can be seen in the north side of the entrance ramp. Strongly developed folds in gneiss and amphibolite have axial surfaces expressed by a moderately strong schistosity "s" striking N-S to N.15 W. Biotite lineations plunge southeast parallel to the hinge lines of these folds. In the vicinity of the shear zone the schistosity swings into parallelism with the N. 35° E. trend of the strong "c" or shear foliation, in a right-lateral configuration. Locally shear bands trend N. 60° E. and produce microoffsets in a right-lateral sense. In vertical section, ramps indicate motion up from the southeast or east. A strong lineation plunges N. 75-80° E. in the plane of the schistosity "s" as an intersection lineation, or down the "c" surfaces, producing a mullion structure common to all mylonites in the Hudson Highlands. The structural elements seen in this small exposure are common throughout the shear zones and basement rocks in the northern end of the Hudson Highlands. At garnet grade or lower (Stops 1 and 2) the zones of strain are spaced and penetrative internal deformation by "s" is rare except within shear zones. Outcrops at this stop exhibit the type of structures illustrated by Figure 6.

Cumulative mileage

- 00 - head west on I-84 (Poughquag quadrangle)
- 1.5 - outcrops of amphibolite, rusty calc-silicate gneiss with near-vertical foliation
- 2.6 - well foliated grey biotite-feldspar gneiss (Ybg), minor amphibolite (Ya) intruded by biotite granite gneiss (Ygg)
- 3.5 - large outcrop of Ygg and Ybg -- park on entrance ramp near west end of roadcut
- 3.8 Stop 1:

(20 min) Middle Proterozoic granitic gneiss, biotitic feldspar

gneiss intruded by Middle Proterozoic granite, Poughquag quadrangle. This stop and the previous two large outcrops illustrate well the nature of the basement rock that is largely unaffected by Paleozoic faulting or mylonitization. This material was the protolith for almost all of the shear zones we will visit. Note the coarse grain size, the coarse gneissosity and style of the Proterozoic folds.

The roadcut exposes a suite of gneisses which are a product of partial assimilation of mafic country rock by intrusive granitic material. The darker-colored gneiss represents the relict parent material and was originally hornblende or hornblende-biotite gneiss. It contains blue-green hornblende altered to biotite, primary biotite, plagioclase and quartz. Small felsic "sweatouts" in the mafic gneiss contain equal proportions of microcline, plagioclase, and quartz.

The lighter-colored gneisses are more clearly intrusive in character and contain abundant microcline, plagioclase, and quartz. Country-rock contamination is evidenced by scattered grains of hornblende altered to biotite and badly corroded garnets. Primary biotite defines the Proterozoic gneissic foliation. A coarse-grained pegmatite appears to cross-cut the other granitic gneisses and may represent a later intrusive event.

With the possible exception of the pegmatite, strongly aligned biotite flakes in the gneisses indicate that intrusion and crystallization took place during or before formation of the dominant mid-Proterozoic gneissic fabric ( $F_2$ ). This fabric is the structure outlined in Figure 1 within Proterozoic rocks. Strained quartz is common in all the rocks, and partial but widespread alteration of plagioclase and breakdown of biotite to epidote and muscovite are indications of Paleozoic retrogression. (Reboard cars and drive to extreme west end of parking lot.)

An hornblende  $^{40}\text{Ar}/^{39}\text{Ar}$  incremental release age of  $928 \pm 20$  Ma (Fig. 10) has been determined from outcrops to the west on I-84.

#### 4.1 Stop 2:

(15 min) Muscovite-rich mylonite zone in biotite-quartz-feldspar-gneiss near Dicktown shear zone (garnet zone).

(Pl. 2-B)

A 3-meter-thick muscovite-rich mylonite is exposed near the contact between biotite granitic gneiss and more plagioclase-rich biotite gneiss. This outcrop is situated within a zone of highly mylonitized and retrograded gneiss within the garnet zone of Paleozoic metamorphism. Tiny euhedral garnets, biotite, epidote and abundant muscovite characterize mylonite in this area. A short

distance west mylonite zones contain Proterozoic garnets retrograded to biotite. The mylonite strikes N.30° E. and dips 74° SE., and contains a strong east-plunging rodding. Exposures of minor shear zones at the west end of the cut show shear bands striking N 55° E. with a right-lateral sense of motion. Retort structures and card-deck displacements of fractured plagioclase crystals might be seen in vertical sections. Tracing of Middle Proterozoic units shows Paleozoic folds were limited to the ductile deformation zone and penetrative Paleozoic foliation is absent from the areas between shear zones.

West on I-84

- 4.5 Crops of rusty quartz-pyroxene and hornblende- plagioclase gneiss, amphibolite and other paragneisses.
- 6.5 Exit I-84 onto Taconic State Parkway headed south toward N.Y.
- 10.6 Hortontown Hill Road, mylonitic biotitic paragneiss and calc-silicate rock in Canopus shear zone.
- 12.7 Exit Rt. 301 west.
- 13.1 Entrance to Canopus Beach (log ends)

Stop 3: Dennytown fault zone and the town of Poughquag (20 min.).  
Quartzite and chlorite-grade phyllonitic rocks.

(Pl. 1A,B)

This exposure marks the westernmost and lowest- grade phyllonite zone that we have mapped. It essentially marks the start of the Paleozoic structural front in which mylonite-phyllonite retrogression becomes important. From this point south for about 2 km a series of small inliers of fault-bounded exposures of Poughquag Quartzite can be found. From this point chloritic, muscovitic phyllonite can be traced southwestward to near Peekskill, New York (Figs. 1 and 2). Phyllonite zones are exposed both east and west of the small syncline in the Poughquag. The unconformity with biotite-quartz-plagioclase gneiss is nearly exposed near the southwest end of the syncline. Muscovite, chlorite, epidote and locally actinolite are retrograde minerals in the shear zones where Proterozoic biotite, garnet, plagioclase and hornblende show complete retrogression to a fine-grained phyllonitic matrix. Plagioclase, microcline, pyroxene, and hornblende are microfaulted, and quartz forms excellent nonannealed ribbon structure.

- 13.1 Log resumes - turn west on Rt. 301.
- 13.4 Turn right onto T.S.P. southbound ramp, park at maintenance shed before parkway.

Stop 4: (15 min.) Chlorite-biotite zone shear zone in Canopus fault.

(Pl. 1-C) Small outcrops by the Taconic Parkway and in the woods to the south illustrate well the type of mylonite and phyllonite developed along the Ramapo-Canopus fault system. Garnet and biotite are altered to chlorite, K feldspar and plagioclase are altered to epidote and muscovite, and quartz is present in ribbon structure. Locally calc-silicate rocks form spectacular fine-grained calcite-rich mylonite in which chunks of diopside-hornblende calc-silicate rocks float. Actinolite is the common alteration mineral along borders of calc-silicate blocks and is present as fine needles in the ductilely deformed calcite matrix. Green muscovite-rich phyllonite like this is easily traced in mapping, forming resistant small ridges in lowlands or distinctive notches on hillsides. This zone extends southwestward to near Annsville, New York where evidence of Proterozoic sillimanite-grade ductile shearing as well as Mesozoic reactivation is present.

(turn around and return to Rt. 301)

- 13.5 Eastbound (right turn on 301)
- 16.2 Turn right on Sagamore Drive
- 16.4 Turn left onto unnamed dirt road.
- 16.6 Stop by small grey garage on left.

Stop 5: Garnet to biotite-grade mylonite in Dicktown fault, (Oscawana Lake quadrangle) (10 min.).

(Pl. 1-D) This small outcrop is the only one easily accessible to a group so we will have to settle for this. The Dicktown fault is traceable as a continuous fault zone expressed by a thickness of up to 100 meters of medium-grained biotite-rich mylonite, protomylonite and ultramylonite along the northwestern side of a large block of internally undeformed Proterozoic gneiss (Figs. 1 and 2). The southeastern boundary of the block is the California Hill shear zone from which the  $436 \pm 3 \text{ Ma } ^{40}\text{Ar}/^{39}\text{Ar}$  age was obtained. The Dicktown fault rocks dip vertically or steeply to the northwest. To the south garnet within this zone is retrograded to biotite. Locally, spears of biotite formed from garnet have aspect ratios of 15:1 in the mylonitic foliation. Microtextures support an up-from-the-northwest and right-lateral sense of motion consistent with the steep northwest-dipping planes shown in Figure 6. To the north tiny euhedral garnets have grown within mylonites of the Dicktown fault.

(Pl. 2-A)

In this outcrop bands of ultramylonite 1-2 cm thick bound layers of mylonite containing complex intrafolial folds. Exposures to the

northwest are of biotite granitic gneiss which is the protolith for this rock.

17.1 Stop sign, turn right on Route 301.

18.6 at Kent Cliffs

19.0 Park on left side of road opposite large roadcut.

Stop 6: Mylonitic fabric, shear zones and reactivation structures at ESERCO drill site (Fig. 11). (20 min).

At this roadcut, in Ygg biotite granitic gneiss, strongly developed mylonitic structures occur in spaced shear zones identified as M1 to M4. Near mylonite zones the K-feldspar becomes pinkish and muscovite more abundant. Augen gneiss is common. A 1 km deep drill hole was completed near the site (Fig. 11) in Nov. 1984 by ESERCO for in-situ stress determinations. The hole started in amphibolite gneiss and penetrated down into granitic gneiss. The log of the drill hole and core taken at spaced intervals reveal mylonitic structures similar to those seen in the roadcut. At 1680 feet a biotite-rich mylonite zone was cored that shows structures similar to zones M1-M3 (Fig. 11). Brittle fractures are spaced throughout the core and the mylonite at 1680 feet shows clear evidence of reactivation as a normal fault. Examination of the mylonites at M-3 and M-4 also reveal gouge and evidence of normal faulting. Triassic reactivation of Paleozoic and older mylonites is common in the Hudson Highlands, and the exposures here illustrate this well. Biotite from the California Hill fault zone resembling zones M<sub>1</sub> - M<sub>4</sub> yielded an  $^{40}\text{Ar}/^{39}\text{Ar}$  plateau age of  $436 \pm 3$  Ma. Zoback and others (1985) determined from hydrofracturing and bore-hole-breakouts that the principal horizontal compressive stress is located in the northeast quadrant.

21.4 Turn right on 301, cross viaduct, turn left and follow 301 to Carmel.

23.6 At light turn left on Rt. 52, follow 52 North.

24.2 Pull into shopping center and park behind buildings.

Stop 7: High grade mylonitic gneiss and blastomylonite in granitic gneiss in sillimanite and kyanite grade at Carmel, New York (Lake Carmel quadrangle).

(Pl.3-C)

Between the last stop and this point we have crossed into a zone in which reformation of the Proterozoic basement is intense. Mylonite zones up to Stop 6 are clearly defined zones of intense deformation that bound relatively undeformed areas of gneiss. In the area between Stop 6 and 7 shear zones contain biotite, garnet

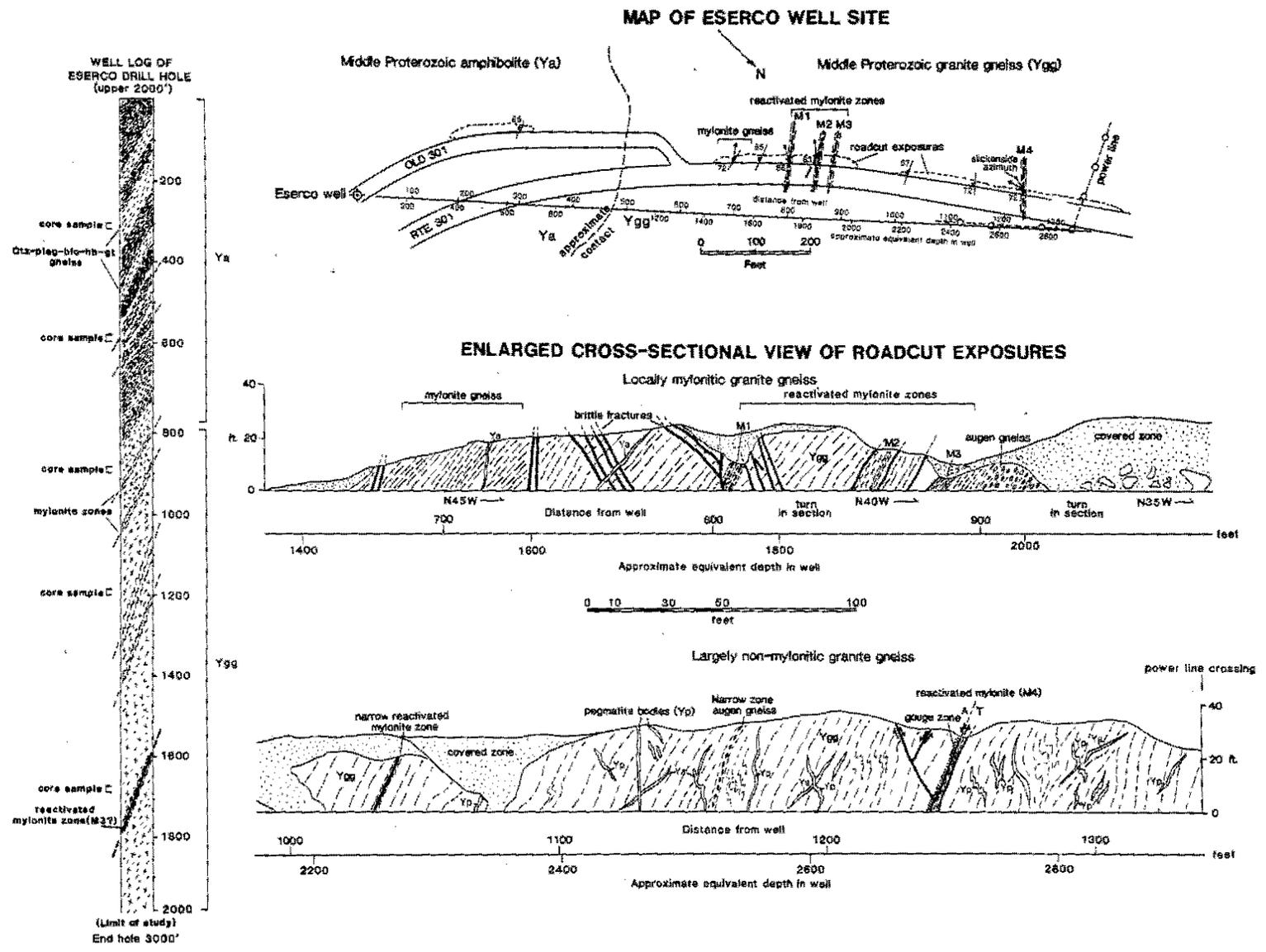


Figure 11 Cross section of roadcut in mylonitic granite gneiss at Stop 6 and partial log of ESERCO drillhole showing mylonite zones M<sub>1</sub> - M<sub>3</sub> sampled in core.

and kyanite and locally staurolite and kyanite, and deformation zones are still discrete features. At the stop and to the east shear zones become less continuous and internal deformation in the gneiss more fully developed, with complete refolding of Proterozoic structures.

The augen gneiss and blastomylonite seen in this outcrop can be traced northward to Stop 9. At this locality, coarse biotite-rich zones and anatectic (?) granitic material define the mylonitic fabric. A strong downdip rodding, retort-shaped augen, and microramp structures define fold-thrust fabrics similar to those seen at lower grades. Muscovite, biotite, microcline and plagioclase all show ductile deformation and evidence of dynamic recrystallization. No sillimanite has been found at this site but we are close to if not within the sillimanite zone.

From the parking lot walk up the slope to outcrops of well-rodded mylonitic gneiss, blastomylonite showing reclined folds with axes plunging N. 75° E., and excellent ramp structure showing an up-from-the-northeast sense of motion. Stringers of aplite and pegmatite appear in segregations in the mylonitic fabric.

- 24.2 Resume log: exit parking lot and turn right on Rt. 52.
- 26.8 Turn right on Rt. 311 S.
- 30.5 I-84--turn onto westbound lane.
- 31.1 Excellent outcrops of sillimanite schist of Walloomsac Formation near contact with gneiss. Sillimanite rods plunge down the dip of  $S_2$  (fault-related structure).
- 31.6 Contact with mylonite gneiss.
- 33.3 Exit 17--turn right at Ludingtonville exit.
- 33.4 Immediately turn left on unmarked road (Carey Rd.)
- 34.5 Stop sign--right onto Holmes Rd.
- 36.0 Rt. 292--turn left at Holmes Rd.
- 36.1 Entrance to Camp Henry Kaufman (log ends).

Stop 8: Contact of mylonitic gneiss with Walloomsac Formation at Whaley Lake thrust (sillimanite grade). (Optional Stop)

Park by superintendent's house in excellent mylonitic gneiss with strong NE-plunging fold-thrust lineation. Walk east 600 feet to small cliff exposure of Walloomsac exhibiting excellent slabby  $S_2$

fabric and biotite spears that plunge northeast parallel to the strong lineation in the underlying and highly folded gneiss. Intensely developed mylonite and fold-thrust fabric is present immediately beneath the Whaley Lake thrust in a zone up to 0.25 km thick.

The Walloomsac Formation and the underlying mylonite derived from gneiss contain fibrolite, garnet, biotite and locally hornblende growing within the thrust ( $S_2$ ) fabric.

From camp turn right (north) on 292.

37.9 Pear Tree Rd.--park on side street.

Stop 9: Augen gneiss at sole of Whaley Lake thrust.

(Pl. 3A-B) Outcrops in the railroad cut under the bridge exhibit excellent augen gneiss with N.  $35^\circ$  W.  $50^\circ$  NE mylonitic foliation. Augen of microcline show retort structure indicating up from the northeast side parallel to the northeast-plunging strong lineation. Outcrops to the north belong to Walloomsac Formation above the Whaley Lake thrust, and contain a strongly developed  $S_2$  fabric with spears of biotite that plunge downdip. From here follow leader into development to N. and park at circle- Note: You must have owner's permission to park here to get to RR tracks. When revisiting this outcrop utilize the RR access road from Rt. 292 at north end of Whaley Lake, not this entrance.

Stop 10: Whaley Lake thrust, carbonate sliver, fold-thrust folds and gneiss sliver.

Outcrops by the lake shore are Poughquag Quartzite with stacked isoclinal reclined folds that plunge N.  $86^\circ$  E. On the east side of the railroad tracks are exposures showing white dolostone (Wappinger) which are also folded into reclined folds. Near the top of the cut steeply-dipping beds on the limb of a recumbent synform are truncated by dark gray, mylonitic Poughquag Quartzite. The fault contact can be traced in the top of the outcrop where recumbent, reclined folds involving both rocks plunge due south to N.  $60^\circ$  E. down the dip of the axial surface of the  $S_2$  folds.

Walk east into the woods following the quartzite to its contact with mylonitic augen gneiss similar to rocks at Stop 9. The upper contact of the gneiss sliver with overlying mylonitic Walloomsac Formation is exposed at the break in slope to the east. This entire sequence is interpreted as slivers of quartzite, dolostone and gneiss interleaved in the Whaley Lake thrust. Near this locality syntectonic ( $S_2$ ) garnet, biotite, and sillimanite are present in the Walloomsac.

Log resumes at Pear Tree Rd. and Rt. 292, at Stop 9--turn left on 292 and follow north to Rt. 55 east.

- 41.3 Stop sign--take 292 to right.  
 42.3 Stop by large outcrop at west end of Walloomsac Formation.

Stop 11: Contact Walloomsac Formation and gneiss.

(Pl. 4A-B) These excellent exposures show the nearly exposed contact between mylonitic granitic gneiss and the Walloomsac at sillimanite grade. "C" and "s" fabrics show up-from-the-east and right-lateral sense of motion. The strong platy structure and isoclinal folds in the Walloomsac are  $S_2$ -generation structures that postdate the crystallization of coarse biotite, garnet and staurolite and locally kyanite. In exposures to the east biotite and kyanite are (Pl. 2D) kinked and form retort structures with syntectonic fibrolite forming the tails, indicating up from the northeast parallel to the elongation lineation. Thrusting and the  $S_2$  event were accomplished under sillimanite-grade conditions.

Biotite from this outcrop yield an  $^{40}\text{Ar}/^{39}\text{Ar}$  plateau age of  $418 \pm 5$  (Figure 10, Bence and McLelland, 1976). Turn around and follow 55 west through town of Poughquag, to Columbia County Rt. 9 and turn right.

- 45.0 Turn right onto Columbia County Rt. 9.  
 47.2 Intersection Columbia County Rt. 21. Continue north on Columbia 9 through Clove to small bog or pond on east side of road 800 feet south of Sweezy Creek. Get owner's permission and park by pond.

Stop 12:  $S_2$  fabric in Walloomsac Formation, aluminous Taconic rocks, and structures of Clove Valley in relation to isograds (30 min).

Park by pond and walk south and east through pasture to low cliffs 1000 feet southeast of pond.

(Pl. 4C,D) These exposures of muscovite-chloritoid-quartz schist contain excellent quartzite beds that outline intense reclined  $S_2$  folds with northeast-plunging axes. Locally, minute garnet is present.  $S_2$  structures like these dominate the fabric in the Walloomsac and Taconic rocks on East Mountain, above the ( $S_2$ ) Poughquag-Clove Valley fault. Locally zones of retrogression are intense and phyllonites contain muscovite and chlorite. In the Walloomsac a second generation of fine biotite and muscovite defines the  $S_2$  structure.

(Pl. 5 A, B, C, D) To the north along the slopes can be seen sheared Walloomsac with  $S_2$  structures and finally a large and spectacular outcrop of brecciated limestone conglomerate belonging to the Balmville Limestone, which forms the basal conglomerate at the base of the Walloomsac Formation.

In Figure 9 we have related the mineral data and the shear zone structures to the isograds. It is important to note that the garnet and higher grade rocks lie for the most part east of and above the Poughquag-Clove Valley fault that is expressed by retrogressive phyllonite. One locality containing garnet is present west of the fault zone, but a garnet-absent retrograded zone intervenes.

In our interpretation the metamorphic isograds have been folded and faulted during late-stage events so that the metamorphic mineral assemblages have been juxtaposed.

Turn right on 9 and follow leader north.

- 48.0 North Clove Rd.
- 49.0 Brush Hill Rd.
- 49.5 Outcrop of Walloomsac with strong  $S_2$  structure and no garnet, sharp turn to left.
- 50.0 Right-angle turn to left
- 50.5 Cross Beaver Creek and park by white barn-house on right side of road.

Stop 13: Green phyllonite in  $S_2$  shear zone.

Pl. 5 A, B) Well displayed small exposures of phyllonite derived from Taconic rocks that normally contain garnet and chloritoid, exposed by the owner's swimming pool and in small ledges to the east. Examination of Rosemary's thin sections from the Dover Plains and Amenia quadrangles to the northeast, on strike with this zone, all exhibit extensive  $S_2$  shearing and retrogression.

In the area west of Clove Valley, on Clove Mountain, chloritoid and biotite both exist sporadically as relict minerals of the  $M_1$  event and are commonly retrograded in the strong penetrative  $S_2$  foliation.

Our conclusion therefore is that the classical prograde metamorphic zonation in eastern Dutchess County is really a structurally complicated and partially retrograded sequence and not an intact package of prograde assemblages.

-End Trip-

Return on Rt. 9 via Rt. 55, to Ludingtonville exit. From here take I-84 east to Danbury and Hartford.

## AGE AND STRUCTURAL RELATIONS OF GRANITES, STONY CREEK AREA, CONNECTICUT

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University of Maryland

### INTRODUCTION

The Stony Creek gneiss dome of coastal Connecticut is a possible diapiric structure containing four distinct granitic (*sensu lato*) units. These show variably concordant to discordant contacts to associated Pre-Cambrian metasediments. The field trip will offer an opportunity to discuss the origins (intrusive, extrusive or anatectic) and ages (Pre-Cambrian? Acadian? Alleghanian?) of the granitic units, and their relations to the metamorphic and structural history of the area. In particular, discussion will center on:

- (i) possible polymetamorphism and associated melting of basement material, and
- (ii) possible relations between melt segregation and shear deformation.

This roadlog presents preliminary petrographic, structural and chemical data. Rb/Sr dating of the various granitic units is in progress, and results should be available at the time of the field trip.

A sketch map showing approximate locations of stops is shown in Fig. 1.

### REGIONAL SETTING

The Stony Creek gneiss dome lies at the intersection of two major structural trends (Fig. 2). It is the westernmost of a series of gneiss domes extending E-W along the Connecticut coast from just east of New Haven to the Rhode Island border. It also lies on the southernmost extension of the Bronson Hill Anticlinorium.

Rocks in the core of the Stony Creek dome have been mapped as inter-mixed Sterling Plutonic Group and Plainfield Formation (Dale, 1923). This interpretation implies correlation with rocks in the core of the Lyme dome to the east, and this in turn is correlated with the Hope Valley terrain of southwestern Rhode Island (Loughlin, 1910). The Hope Valley terrain is ~ 600m.y. in age (Hermes et al., 1981; Gromet and O'Hara, this volume). In support of possible correlation, Hills and Dasch (1972) obtained a 616m.y. Rb/Sr whole-rock age on the Stony Creek. However, the interpretation of this age presents some difficulty in view of the large error ( $\pm 78$ m.y.) and poorly-controlled sampling.

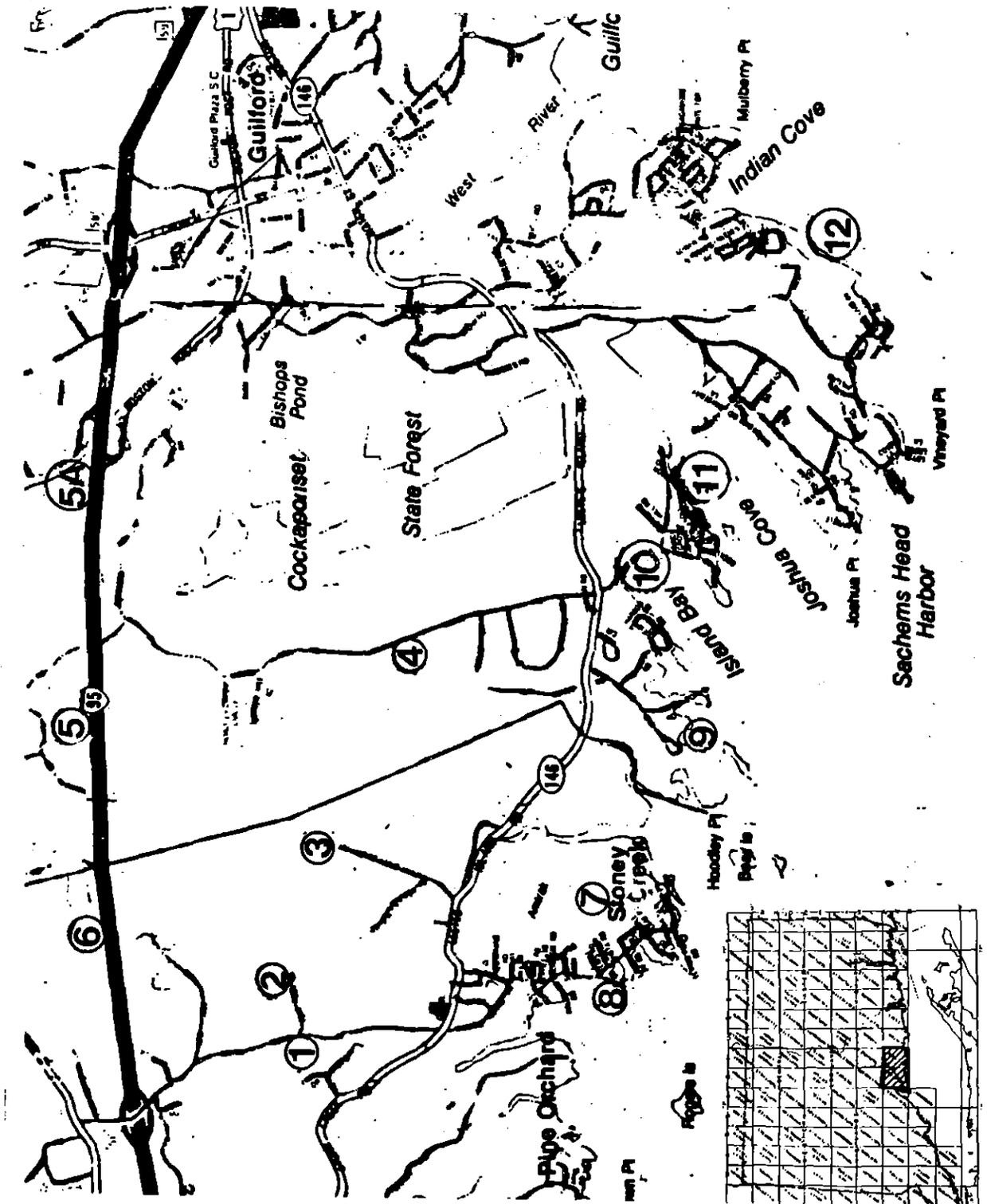


Figure 1. Location of field trip stops.

Rocks of the Lyme dome and Hope Valley terrain are separated from Siluro-Devonian rocks of Merrimack Synclinorium affinity by the Honey Hill Fault. The significance of this structure is unclear; movement on the fault is known to have occurred ~ 250m.y. ago (O'Hara and Gromet, 1983) but it is not known whether this movement was an Alleghanian suturing event or local basement-cover sliding on an existing (?Taconic ?Acadian) suture. Mylonites of the Honey Hill Fault are deformed by the Lyme dome (Lundgren and Ebbin, 1972) suggesting the Lyme dome to be a late Alleghanian feature. However, the Honey Hill Fault cannot be traced west of the Lyme dome, and no major tectonic discontinuity can be traced between the Stony Creek rocks and overlying Siluro-Devonian strata.

Hall and Robinson (1982) suggest correlation of units in both the Stony Creek and Lyme domes with the ~ 560m.y. (Naylor et al., 1975) Dry Hill Gneiss in the Pelham Dome of the Bronson Hill Anticlinorium. Domes in the Bronson Hill are generally considered to be of Acadian age (c.f. the Alleghanian Lyme dome) (Hall and Robinson, 1982). It has been suggested that ~ 600m.y. rocks of Pelham-Stony Creek-Lyme-Hope Valley affinity were already sutured to the North American continent by the time of Taconic (Hall and Robinson, 1982) or Acadian (Naylor, 1975; Osberg, 1978) orogeny. An attempt to reconcile apparently conflicting evidence in terms of successive docking of discrete fragments of a dispersed terrain is presented by Gromet and O'Hara (1983 and this volume).

There are no studies of the grade or timing of metamorphism in the Stony Creek area. The role of Alleghanian metamorphism in eastern Connecticut is very difficult to determine. Dallmeyer (1982) has reported ~ 250m.y. K/Ar and  $^{40}\text{Ar}/^{39}\text{Ar}$  ages across the Hope Valley terrane and into an area of eastern Connecticut traditionally viewed as having been subject only to Acadian events. Similarity of biotite (~248-240m.y.) and hornblende (~ 260-255m.y.)  $^{40}\text{Ar}/^{39}\text{Ar}$  ages argues for this being a discrete Alleghanian thermal pulse, rather than slow post-Acadian cooling. No data are yet available from as far west as Stony Creek.

In the interpretation of their Rb/Sr data, Hills and Dasch (1972) note localised remobilisation of Rb on a scale ~ 0.3m; on this scale data could be fitted to a ~ 250m.y. isochron. Hills and Dasch therefore suggested Alleghanian metamorphism as a possible explanation, but noted that "earlier redistribution of Rb and Sr cannot be disproven".

Two problems therefore arise:

- (i) can the Stony Creek rocks be correlated with any or all of the Pelham dome, Lyme dome or Hope Valley sequences?
- (ii) what was the timing of suturing of Stony Creek rocks to the North American continent--Taconic, Acadian or Alleghanian?

#### STRATIGRAPHIC FRAMEWORK

Some discussion of stratigraphy in the Stony Creek area may be found in Mikami and Digman (1957), Bernold (1962) and Sanders (1968). The litho-stratigraphic sequence proposed by the author is shown in Table 1, and the

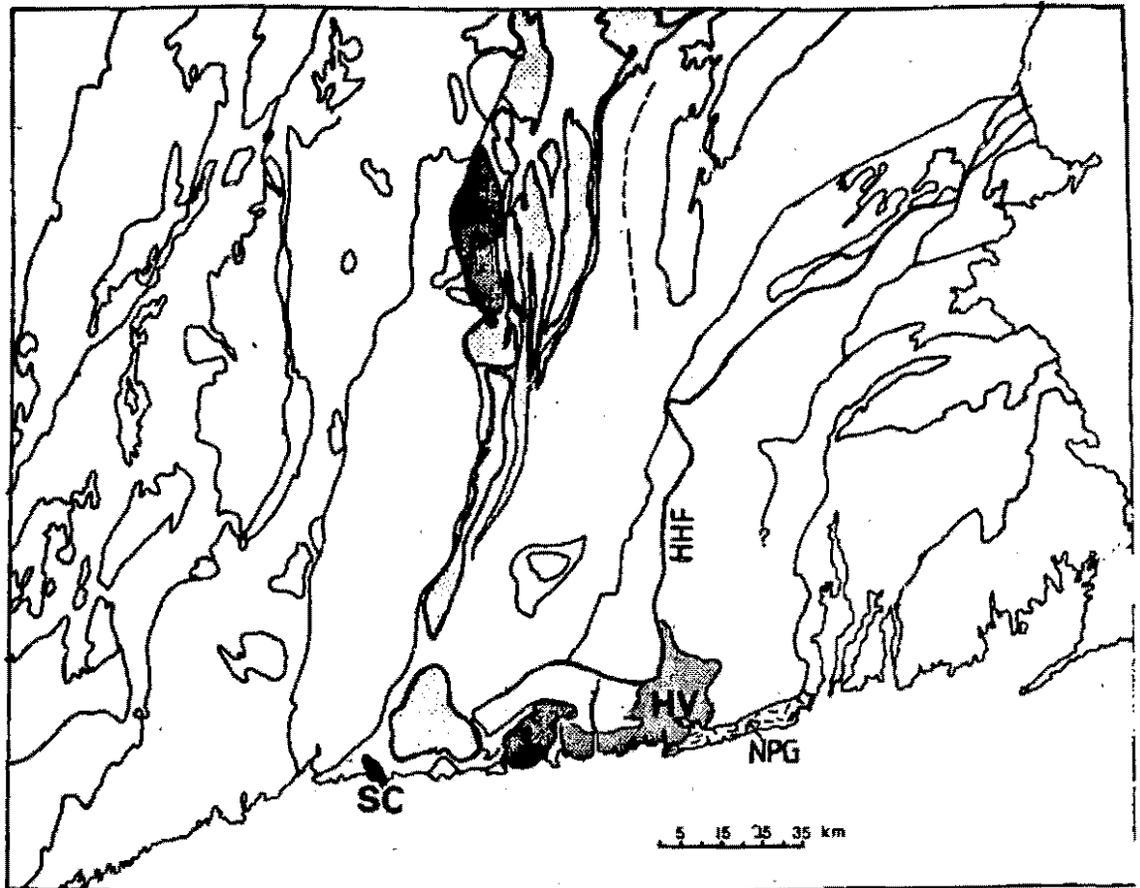
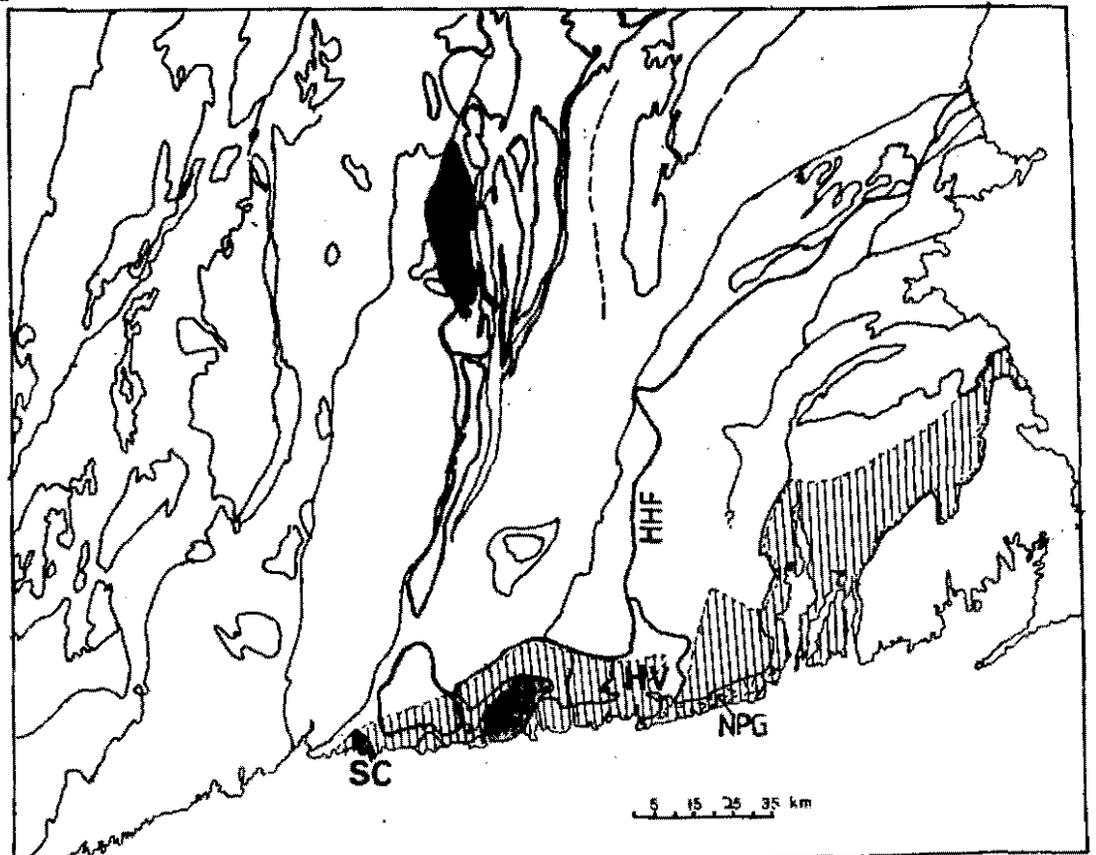


Figure 2. Geologic setting of the Stony Creek Dome, (a) Location of Stony Creek, Lyme and Pelham Domes, Hope Valley terrane and other major structures.



■ PreCambrian metamorphism    ▨ Alleghanian metamorphism  
 (b) Regions of PreCambrian and Alleghanian metamorphism, from Robinson (1984).

distribution of the various units is shown in Fig. 3. This sequence is similar to that of the above authors, and to that of Goldsmith (1967) for the Lyme dome, with the following additional information.

The core of the dome is dominated by gneisses and amphibolites, identified as part of the Lower Plainfield by the abundance of garnet-bearing lithologies. The amphibolites are typically coarse-grained and massive. They may have a spotty appearance in outcrop (e.g. STOP 12) due to the irregular distribution of plagioclase crystals, and they are usually extensively veined by subpegmatitic tonalite veins. The majority of the Lower Plainfield gneisses are dark bi-plag-qz-gt  $\pm$  amph gneisses, with very pronounced gneissic banding on a scale of  $\sim$  2cm. Garnet selvages are very common along the contacts to associated granites (STOP 11). The Middle Plainfield is very heterogeneous. Quartzites occur at several different levels within the Plainfield, but only the Upper Plainfield is dominated by quartzite and only at this level is the rock massive. This massive quartzite marks the top of the Plainfield.

Mamacoke and Monson gneisses are relatively easily distinguished by the abundance of very dark, hb-bearing lithologies in the latter, together with its more massive character. The Mamacoke is commonly layered on a scale 1-2cm., and often weathers with a distinctive red-and-green striped appearance.

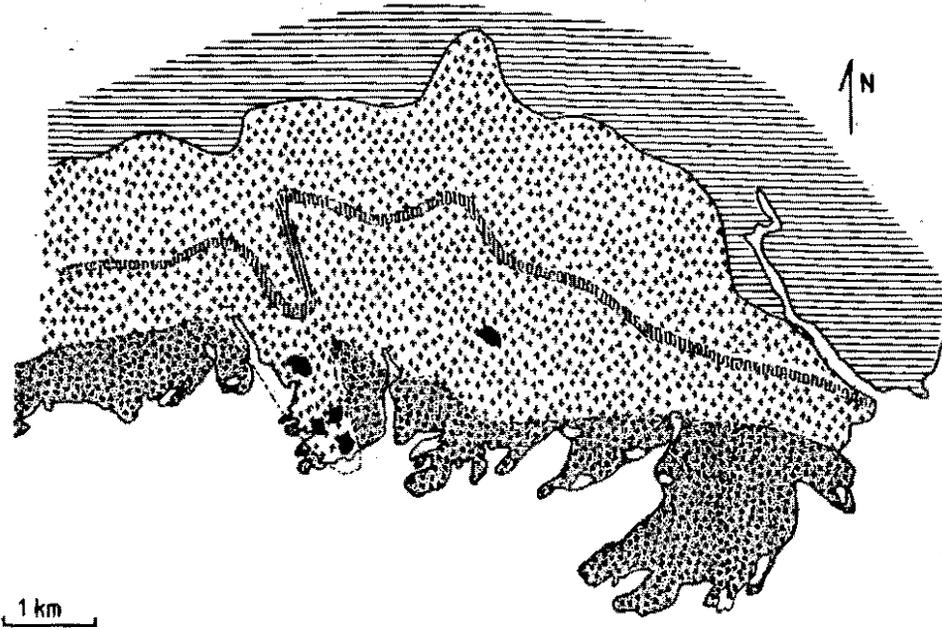
#### STRUCTURAL FRAMEWORK

Interpretation of the Stony Creek structure as a dome follows from the approximately concentric disposition of lithostratigraphy (Fig. 3) with a core of older units (Lower Plainfield) and a margin of younger units (Mamacoke and Monson). Basement domes in the southern and central Appalachians have been variously attributed to thrusts, diapirs and refolded nappes (Hatcher, 1984; Muller and Chapin, 1984). Gneiss domes in the Bronson Hill Anticlinorium, of which the Stony Creek dome is the southern extension, are considered diapiric (Hall and Robinson, 1982). The Lyme dome may be a basement ramp (Losh and Bradbury, 1984). The alternative possibilities can be evaluated by careful search for the characteristic thrusts, strain distributions and cross-folds.

No major thrusts have been recognized in the Stony Creek dome or at its contact to overlying strata. The mapped distribution of Upper Plainfield strata reveals the existence of a north-plunging overturned anticline, the Sachem's Head anticline (Sanders, 1968), the axis of which is almost coincident with one axis of the dome (Fig. 3). Other, more minor, folds parallel to this axis may also be recognized. However, the cross-folds which would be necessary to warp the linear anticline into the observed dome shape have not been observed.

Much of the evidence is suggestive of a diapiric origin. Foliation (phase layering in SCGI and biotite selvages in SCGII) is flat-lying in the core of the dome and steepens gradually outward to  $\sim$ 70 $^{\circ}$  at the margins (Fig. 4, STOP 6). Lineation (defined by feldspar shape aggregates in SCGI and SCGII and by feldspar augen and biotite stringers in SCGII) likewise shows a transition from flat-lying in the core to steeper in the margins (Fig. 4). This lineation is not strictly a downdip lineation--it is often

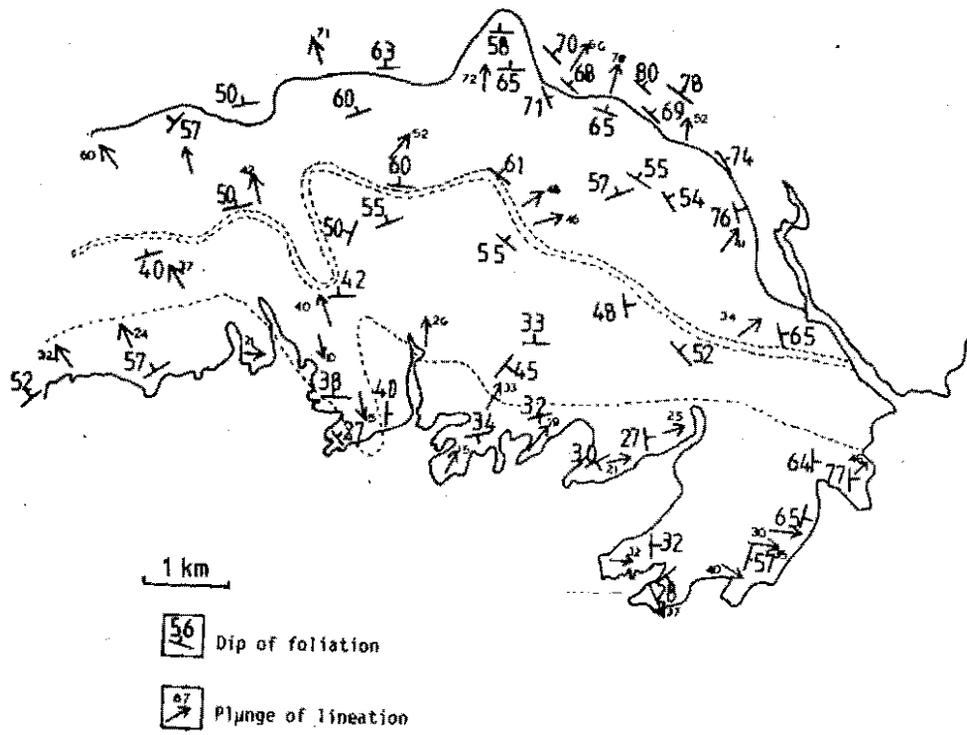
UNITS IN STONY CREEK DOME



- |   |                             |   |                        |
|---|-----------------------------|---|------------------------|
|  | Mamacoke Formation          |  | Stony Creek Granite I  |
|  | Upper Plainfield Formation  |  | Stony Creek Granite II |
|  | Middle Plainfield Formation |   |                        |
|  | Lower Plainfield Formation  |   |                        |

Figure 3. Geological sketch map of Stony Creek Dome.

FOLIATIONS AND LINEATIONS WITHIN THE DOME



- |   |                     |
|---|---------------------|
|  | Dip of foliation    |
|  | Plunge of lineation |

Figure 4. Foliations and lineations in Stony Creek Dome.

more steeply dipping than foliation, possibly suggestive of non-coaxial deformation. L-fabrics are strongly developed in the core of the dome, whereas margin fabrics are LS or S type. The concentric foliations, radial lineations and change outward from L to S fabrics are more suggestive of an origin by diapirism than by multiple folding.

An estimate of variations in strain across the dome is shown in Fig. 5(a). Strain estimates are from:

- (i)  $R_f/\phi$  analysis of grain aggregates in SCGI and SCGII (only  $R_f$  shown on Fig. 5(a))
- (ii) direct measurements of axial ratios (X:Z) of enclaves in SCGI.

Not surprisingly, enclave measurements suggest greater ellipticity (higher strains) than do grain aggregate measurements. This may be due to originally higher ellipticity ( $R_i$ ), indicated by the tendency of biotite schists to have higher ratios than do (presumably more isotropic) amphibolites or quartzites from the same area. Only values for quartzites are shown on Fig. 5(a), to allow for comparability. Alternatively and/or additionally, if SCGI is intrusive then the greater ellipticity of enclaves than of granite fabrics may indicate that enclaves are in fact xenoliths deformed more strongly than the enclosing granite, due either to syntectonic intrusion of SCGI or to preintrusion deformation of the country rock. Compatible with this latter suggestion is the truncation of folded fabrics in the enclaves at their contacts with SCGI (STOP 2). Comparison of the strains recorded in SCGI and SCGII is complicated by their spatial separation; it is not clear whether the differences in strain magnitude (SCGI more strained than SCGII) are due to real differences in strain history (SCGI predates SCGII) or are due to variations in strain across the dome.

On a Flinn diagram (Fig. 5(b)) data are clearly suggestive of near-plane strain in the core of the dome and increasing flattening in the margins. Plane strain near the core of the dome is compatible with the diapir models of Dixon (1975) which predict a neutral surface ( $K=1$ ) where vertical extension in the "stalk" passes into horizontal extension in the "cap" of a mushrooming diapir. Similar strain distributions have been reported by various workers (Holder, 1979; Ramsay, 1979) for plutons infilled by ballooning; however, the apparently diapiric strains extend well beyond the regions of possible intrusion (SCGI outcrops).

If the Stony Creek dome is indeed a diapir, this may imply a structural history more similar to the Pelham dome than to the Lyme. This, in turn, would by implication suggest Acadian-age diapirism, i.e. Stony Creek basement would have been already sutured to North America by the Acadian.

Both foliations and lineations are deformed by shear zones (Plate 8). These typically dip gently to the NW (STOP 3) or N (STOP 12). They are the last penetrative deformational event seen. SCGIII and SCGIV are closely associated with these shear zones and commonly are concentrated within them (Fig. 20).

Any proposed structural history for the Stony Creek dome must take account of the above, and must also answer the following:

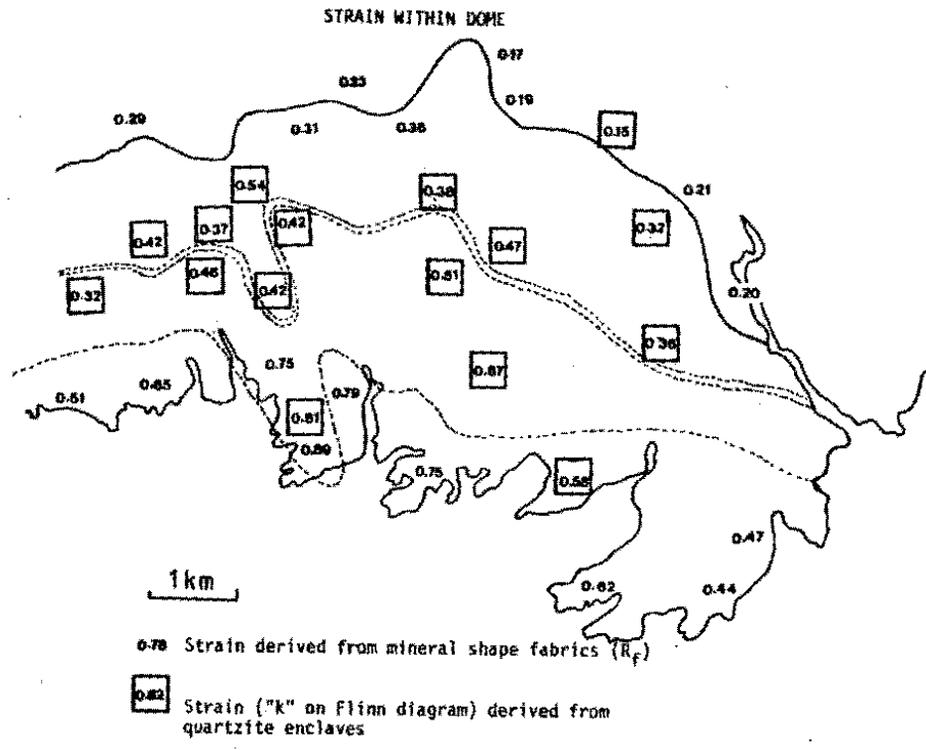
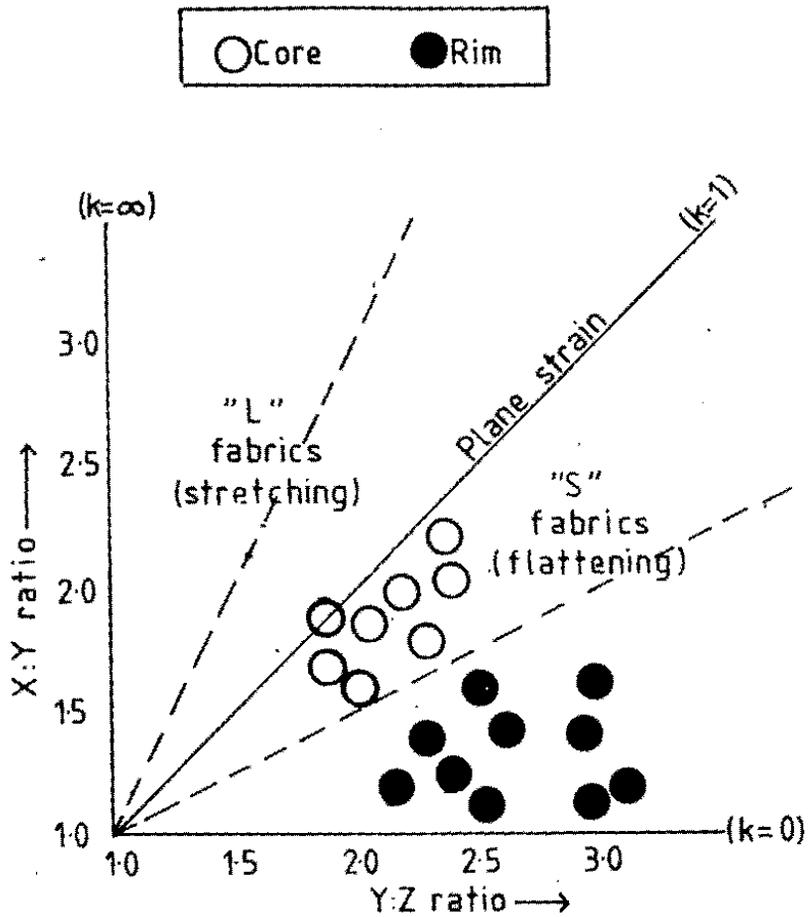


Figure 5 (a). Strain within the Stony Creek Dome.



(b) Flinn diagram of enclave measurements.

- (i) what is the timing of diapirism?
- (ii) what is the relative timing of SCGI and SCGII formation, and of this compared to formation of the Sachem's Head Anticline? Can SCGI develop in a pre-existing fold (possibly by lit-par-lit injection, STOPS 3 and 6)? Alternatively, could the fold develop on a large scale without producing small-scale fabrics in the granite?
- (iii) what is the timing of folding seen in enclaves of Plainfield within SCGI? Do these folds relate to the Sachem's Head Anticline, or are they an earlier, possibly PreCambrian (the Hills and Dasch (1972) date, if accepted) event?
- (iv) does the occurrence of SCGIII and SCGIV in shear zones indicate the segregation of melt into shear zones, or the localisation of strain in relatively low-strength melt areas? Why do SCGIII and SCGIV lack fabrics--was melt crystallisation post-tectonic, but melt-segregation syn-tectonic? If SCGIII is ~ 250m.y. (as implied by Hills and Dasch (1972)) then these shears must be of Alleghanian age.

#### METAMORPHIC FRAMEWORK

Only limited geobarothermometric data is presently available. Further work to detail (i) possible variations in P and T within the dome and (ii) changes in P and T with time is planned. Deduced metamorphic conditions are plotted on Fig. 6 together with curves for the beginning of melting of granite (Johannes, 1984) and of peraluminous granite (Clemens and Wall, 1981).

#### Ternary Feldspars

Compositions of coexisting feldspars are plotted in Fig. 7, where they are compared to the isotherms of Whitney & Stormer (1977) for low-structural state feldspars. The calculated temperatures are slightly P-dependent, and maximum recorded temperatures range from ~ 680°C at 5kb to ~ 720°C at 10kb. Most temperatures are lower than this due to re-equilibration. The maximum recorded temperatures are likewise under-estimates of maximum attained temperatures, due to possible re-equilibration.

#### Garnet-biotite

Due to the generally high Al and Ti contents of the biotites, the formulation of Ferry and Spear (1977) gave anomalously high temperatures. Recalculation using the equation of Indares and Martignole (1985) suggests temperatures in the range 730°C to 770°C. These temperatures are for garnet-biotite pairs from the Lower Plainfield and work is in progress to obtain thermal data from other units.

#### Qz-Plag-As-Gt

Using the equation of Ghent (1976) a pressure of 8.1 was obtained from this geobarometer; this estimate appears too high in view of the absence of kyanite from the Stony Creek rocks (although this may be a bulk composition

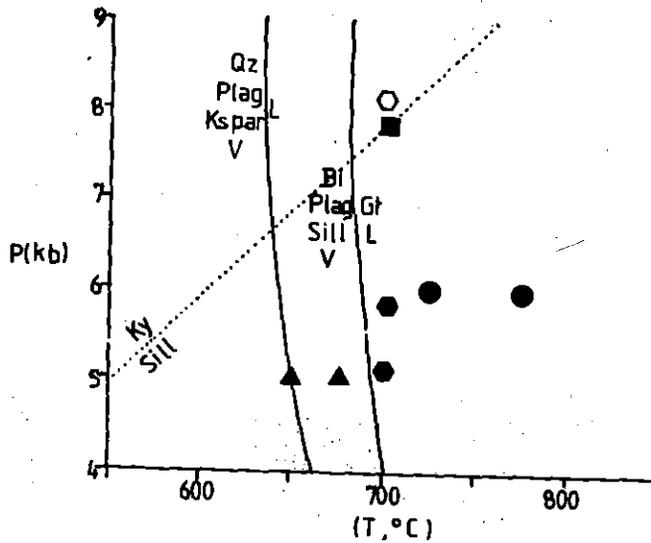


Figure 6. Metamorphic conditions in the Stony Creek Dome. Temperatures: ● from garnet-biotite geothermometry (Indares and Martignole, 1985), ▲ from feldspar geothermometry (Whitney and Stormer, 1977). Pressures from qz-plag-gt T=700 (Ghent, 1976), ● (Newton and Haselton, 1981), ■ GRAIL T=700 (Bohlen, 1984). Curves for granite melting are from Johannes (1984) and for garnet-granite melting from Clemens and Wall (1981).

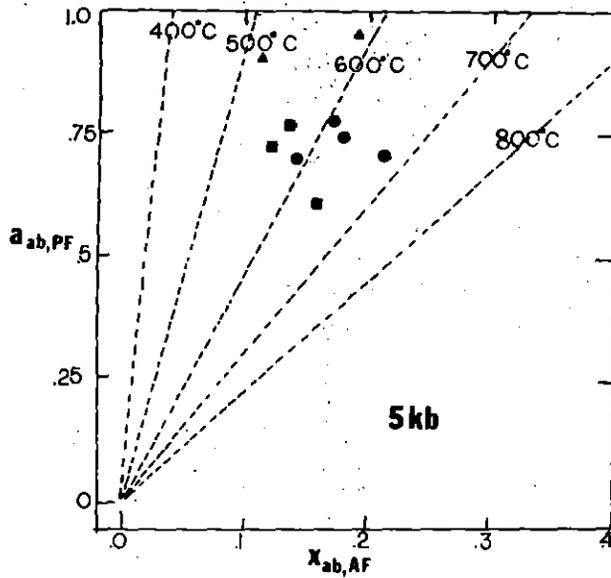
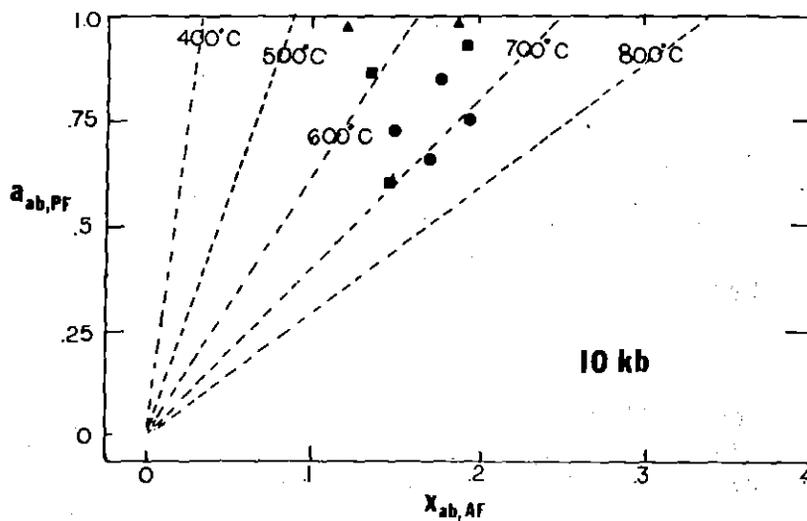


Figure 7. Compositions of ternary feldspars. Isotherms are from Whitney and Stormer (1977).



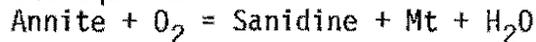
problem). The method of Newton and Haselton (1981) gives pressures of 5.2kb to 5.8kb from the core and rim of a garnet respectively. These values may be low, but the relative increase in pressure from core to rim appears real.

### GRAIL

One sample from the Lower Plainfield Formation contains rutile in the core of the garnet and ilmenite in the rim (note that this implies P decrease during garnet growth, the opposite to that noted above). Using the data of Bohlen (1984), the assemblage gt-rutile-sill-ilm-qz in this rock implies  $P \sim 7.8\text{kb}$ . This appears high in view of the lack of kyanite.

### Bi-Kspar-Mt

Using the equilibrium:



and assuming  $T \sim 700^\circ\text{C}$ ,  $a_{\text{H}_2\text{O}} = 1$ , we obtain a value of  $\log f_{\text{O}_2}$  for samples of SCGIII ranging from -20 to -22 (NNO to QFM) (Eugster & Wones, 1962).

### Possible polymetamorphism

Complex garnet zoning (Fig. 8) may be interpretable in terms of polymetamorphism; further work is needed to constrain possible pressure variations noted above. Evidence for polymetamorphism would clearly be of great significance for the tectonic evolution of the Stony Creek dome.

## GRANITIC UNITS

In addition to the stratigraphic sequence outlined above a variety of granitic (s.l.) units can be recognised (Fig. 3, Plates 1-3, 5-8). Modal and bulk chemical analyses of three of these units can be found in Tables 2 and 4. The amount of granitic material is greatest in the core of the dome, where outcrops of homogeneous granite occur, and decreases outward. Granite is very much more abundant in the Plainfield Formation than in the overlying Mamacoke and Monson, where it is typically restricted to cross-cutting pegmatites (Fig. 9). Four distinct granitic units (Stony Creek Granite I-IV (SCGI-IV)) can be recognised on the basis of mineralogy and relations to structural chronology.

### Field Relations

The first of these units, referred to as Stony Creek Granite I (SCGI) is most abundant in the northern part of the dome. It was the subject of the local quarrying industry and is the type referred to in the literature as "Stony Creek Granite". It is a medium-grained (3-5mm), granular-weathering biotite-granite. A garnet-bearing facies of SCGI is developed locally (STOP 7) (Plate 3) where it appears gradational to the more normal type. This rock type is also found to the west of Stony Creek where it has been mapped as the Branford Quartz Monzonite (Mikami and Digman, 1957). SCGI can be seen to wedge into disrupted country rock (STOPS 3 and 5) and to assimilate xenoliths of Plainfield Formation (STOPS 3 and 5). Large blocks of resistant quartzite from the Upper Plainfield form a series of

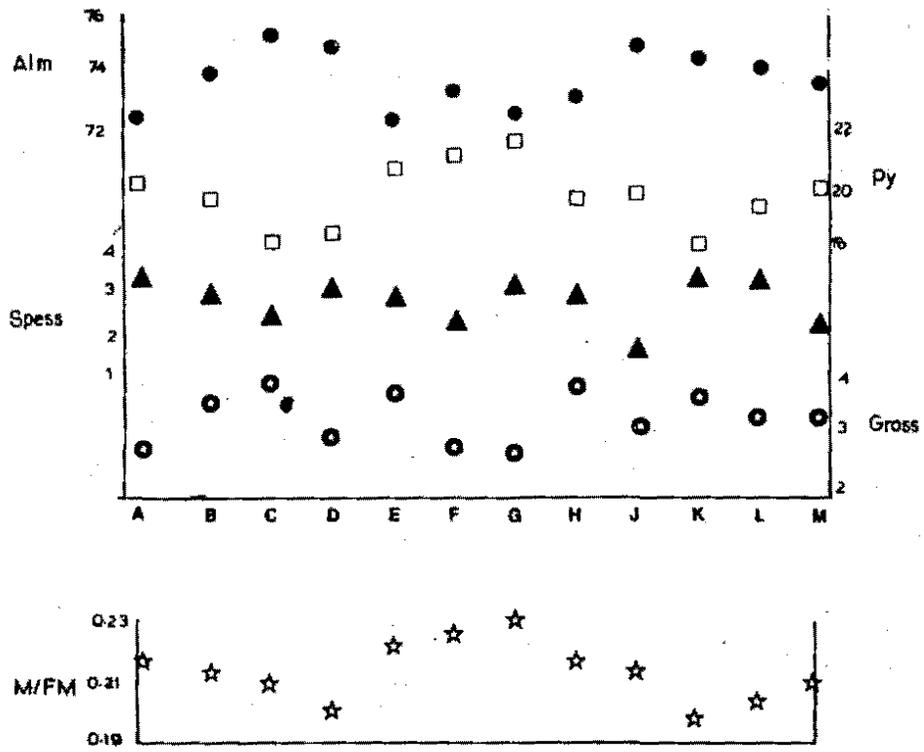


Figure 8. Zonation in country rock garnet.

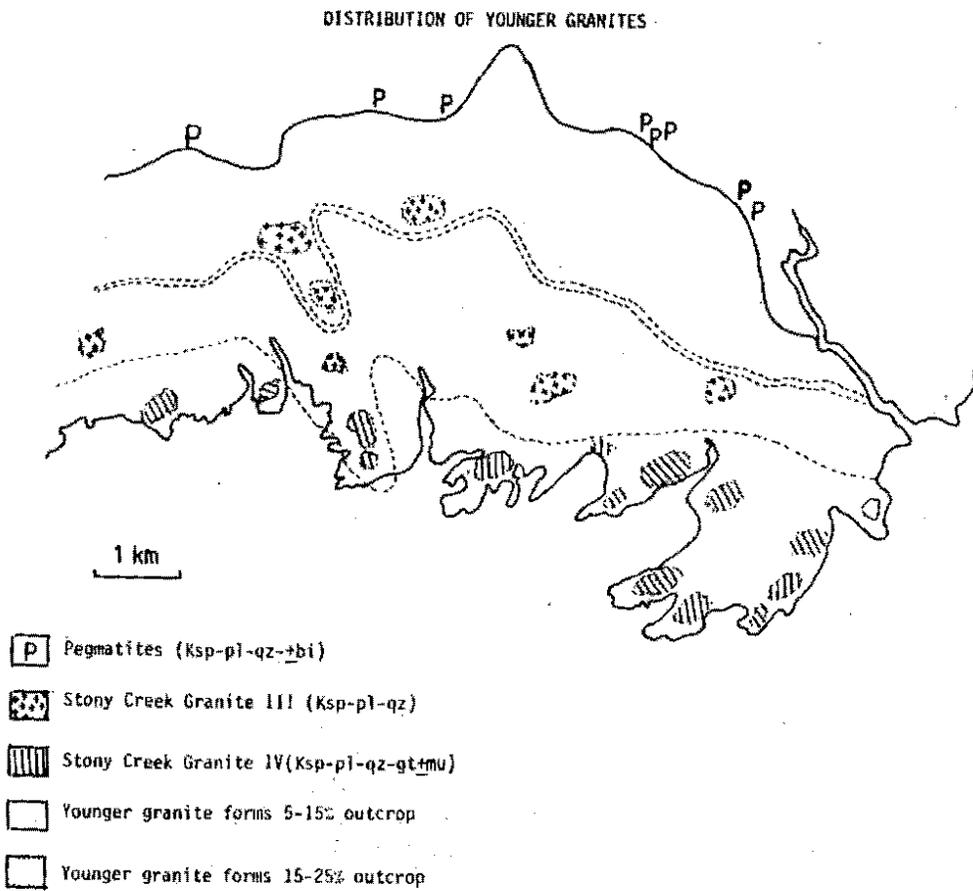


Figure 9. Distribution of younger granites, SCGIII and SCGIV.

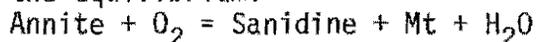
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In addition to the stratigraphic sequence outlined above a variety of granitic (s.l.) units can be recognised (Fig. 3, Plates 1-3, 5-8). Modal and bulk chemical analyses of three of these units can be found in Tables 2 and 4. The amount of granitic material is greatest in the core of the dome, where outcrops of homogeneous granite occur, and decreases outward. Granite is very much more abundant in the Plainfield Formation than in the overlying Mamacoke and Monson, where it is typically restricted to cross-cutting pegmatites (Fig. 9). Four distinct granitic units (Stony Creek Granite I-IV (SCGI-IV)) can be recognised on the basis of mineralogy and relations to structural chronology.

### Field Relations

The first of these units, referred to as Stony Creek Granite I (SCGI) is most abundant in the northern part of the dome. It was the subject of the local quarrying industry and is the type referred to in the literature as "Stony Creek Granite". It is a medium-grained (3-5mm), granular-weathering biotite-granite. A garnet-bearing facies of SCGI is developed locally (STOP 7) (Plate 3) where it appears gradational to the more normal type. This rock type is also found to the west of Stony Creek where it has been mapped as the Branford Quartz Monzonite (Mikami and Digman, 1957). SCGI can be seen to wedge into disrupted country rock (STOPS 3 and 5) and to assimilate xenoliths of Plainfield Formation (STOPS 3 and 5). Large blocks of resistant quartzite from the Upper Plainfield form a series of

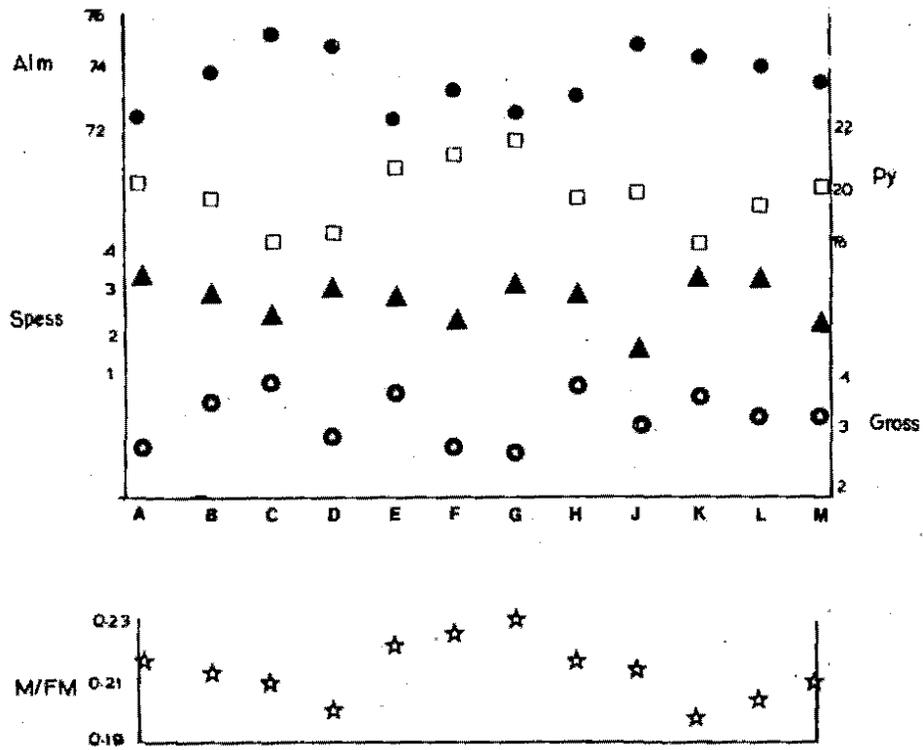


Figure 8. Zonation in country rock garnet.

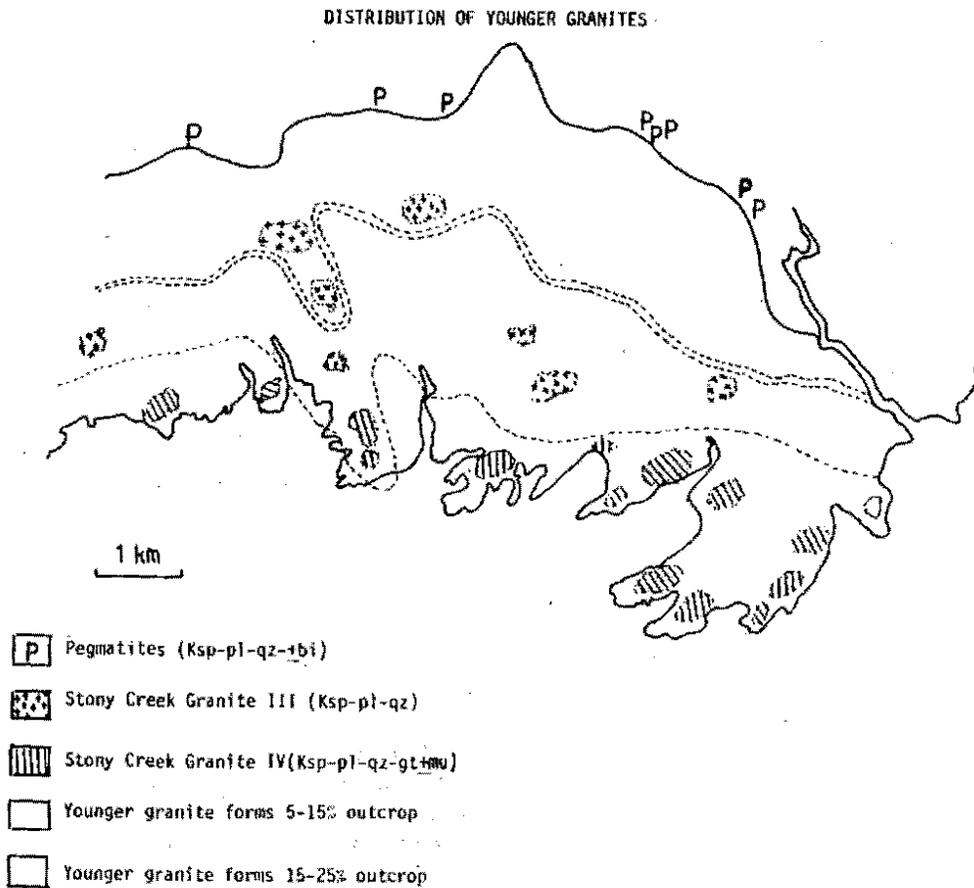


Figure 9. Distribution of younger granites, SCGIII and SCGIV.

raft trains near the outer margin of the body. Xenoliths are most common near the outer edges of the dome. There is a gradual transition in granite: host proportions over a distance ~ 400m. from country rock with veins (STOP 1) to masses of SCGI with occasional xenoliths (STOP 2). Fabrics within xenoliths are truncated at xenolith margins (STOP 2). SCGI shows pronounced phase layering, marked by the development of Kspar-rich and qz+plag-rich bands. This phase layering is parallel to locally abundant biotite schlieren (STOPS 3 and 6) (Plate 1). Both features were attributed by Mikami and Digman (1957) to filter pressing during melt crystallization.

The second granitic unit, SCGII, is a relatively fine-grained (1-3mm.) garnet- and locally muscovite-bearing granite. It is found in lit-par-lit association with the Lower Plainfield Formation. Layers of SCGII may be very thick, up to 10m. across; many outcrops contain several such units separated by thin layers of Lower Plainfield (STOP 9). Fragments of Lower Plainfield can be traced out into surrounding SCGII as biotitic clots and stringers of tiny garnets (STOPS 9 and 11). SCGII shows a well-developed lineation (Plate 5) formed both by biotite and garnet stringers (Plate 6) and by elongate aggregates of qz and feldspar. Kspar augen are common (STOPS 9 and 12).

The third granitic unit (SCGIII) is typically very closely associated with SCGI (STOPS 3 and 4). Rocks apparently correlative to SCGIII can be found as pegmatitic veins cutting the Plainfield, Mamacoke and Monson Formations (STOPS 5a and 6). Where associated with SCGI, SCGIII may form 5-15% of the outcrop and may occur either as veins (~ 3-20cm. wide) cutting layering in SCGI, or as subpegmatitic to pegmatitic (grain size 5-10mm.) patches. Such patches range in size from ~ 3x3cm. to ~ 40x40cm. Pegmatitic patches of SCGIII merge into host SCGI by a decrease in grain size and by a decrease in the proportion of Kspar. The high Kspar content of those patches renders them more distinctly pink than the host. Veins are more sharply bounded than patches and often have a ~ 1cm. wide selvedge rich in bi + plag. Neither veins nor pegmatitic patches have an internal fabric (Plate 2). Coarse garnet-bearing veins in the garnetiferous facies of SCGI (STOP 7) may be a garnetiferous facies of SCGIII and will be treated as such below.

SCGIV occurs as pods and veins at a high angle to the lineation in SCGII. Proportions of SCGIV to SCGII in outcrops vary from ~ 5:95 to 20:80 (STOPS 9 and 12). SCGIV is often concentrated in shear zones (Plate 8) within SCGII (STOP 12); such pods can be traced several m. across the outcrop. SCGIV lacks the fabric of the enclosing SCGII, and contains garnets which are both coarser and more abundant than those in SCGII.

### Petrography

Modes of units SCGI-III are shown in Table 2 and Fig. 10. With the exception of a few samples of SCGII (which are classified as quartz monzonites) samples of all units lie in the granite field of Streckeisen (1976). Shown on Fig. 10 for comparison are modes of granitic rocks from possibly correlative Sterling Plutonic Group sequences, i.e. the Hope Valley alaskites of Rhode Island and alaskites and gneisses from the Lyme dome. Data from the Stony Creek samples are more Kspar-rich than the field of Hope Valley alaskite samples. Material from the Lyme dome is represented by only 2 samples; however, these samples lie in the center of the field defined by the Stony Creek

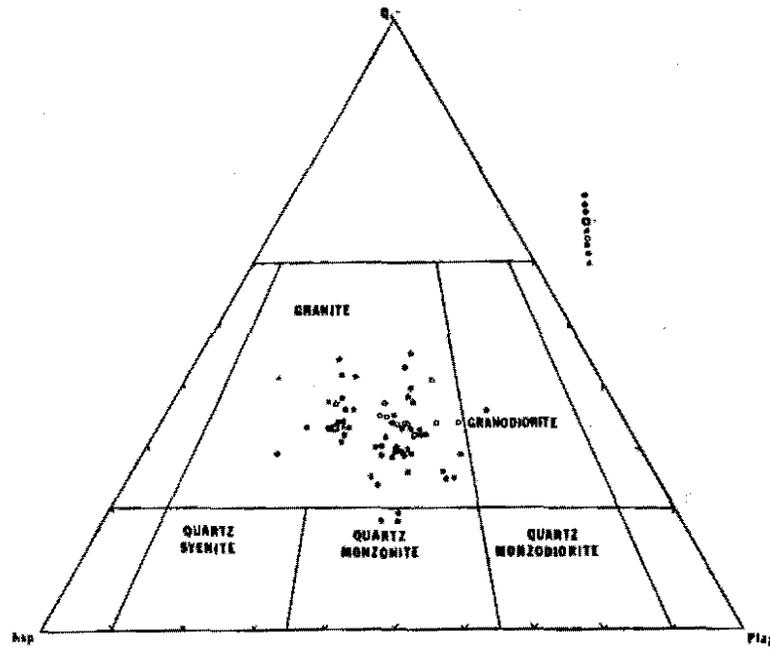
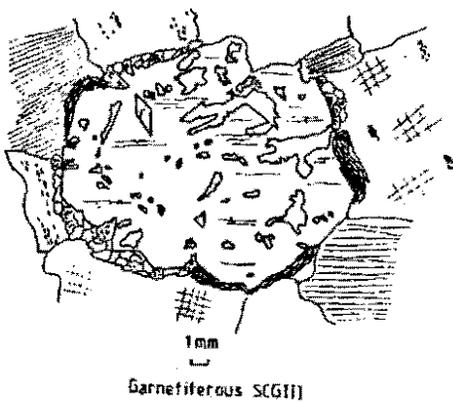


Figure 10. Modes of SCGI■, SCGII●, SCGIII▲, in Qz-Ab-Or. Other data: Westerly★ and Narragansett Pier Granites✱ from Hermes et. al. (1981) and Quinn (1972); Hope Valley alaskites✶ and Ten Rod Granite Gneiss○ from Day et. al. (1980); Lyme alaskite⊕ and biotite-granite-gneiss✧ from Goldsmith (1967).



Garnetiferous SCGIII

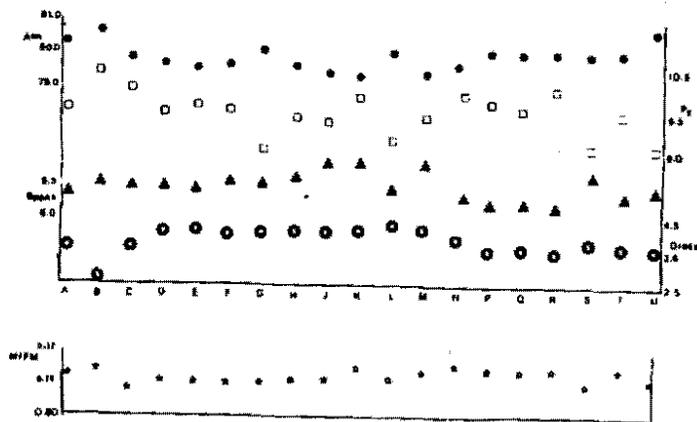


Figure 11. Garnet from SCGIII sample 7, showing (a) qz inclusions and rim of sillimanite, together with (b) zoning.

data. Also shown on Fig. 10 are modes of the Alleghanian-age Narragansett Pier and Westerly Granites. Coarse material of SCGIII type has previously been interpreted as pegmatites of Narragansett Pier Granite; however, Fig. 10 shows that the fields of SCGIII and of Narragansett Pier Granite are different.

SCGI samples in thin section show very highly strained quartz; grain contacts are extensively sutured and subgrains are abundant. In some samples local production of smaller, more equant, strain-free quartz grains has occurred. The samples have therefore been deformed at temperatures where recovery and recrystallisation was possible ( $\sim 400^{\circ}\text{C}$ ). Work to characterise quartz c-axis fabrics in terms of temperature and strain-rate is planned. No relict igneous textures can be recognised; all Kspar is microcline. SCGI samples contain abundant accessory sphene, zircon and magnetite.

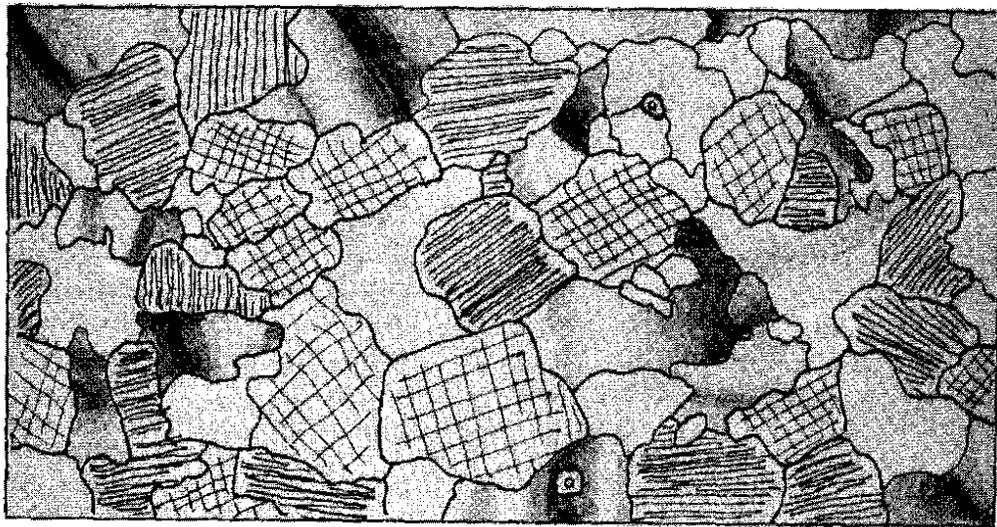
The garnetiferous facies of SCGI (STOP 7) looks texturally like the garnet-free SCGI, with evidence of recovered and recrystallised quartz. Biotite flakes define a layering.

SCGII samples in thin section show textures very similar to SCGI, with no relic igneous textures and with abundant recovered and recrystallised quartz grains. Ribbons of sutured quartz are very common, and elongate aggregates of feldspar (interpreted as the products of recrystallisation of highly deformed feldspar) occur. Both shape fabrics and phase layering are more obvious in SCGII than in SCGI. Apatite is an abundant accessory in SCGII; magnetite and zircon occur sporadically.

SCGIII samples in thin section are highly inequigranular, with Kspar much coarser than the other species. Kspar grains are often perthitic, and often contain plagioclase and quartz inclusions. Myrmekitic intergrowths of plagioclase and quartz often invade Kspar. Some Kspar is microcline. Kspar grains often show brittle fractures, which locally displace perthite lamellae and/or microcline twins. Plagioclase is occasionally Carlsbad-twinned, and may show discontinuous zoning to more sodic rim compositions. Quartz occurs as strained grains with abundant fluid inclusion trails. These fluid inclusions will be examined in a later stage of this project. There is no sign of recrystallisation in quartz. Accessory minerals are represented by large grains of magnetite.

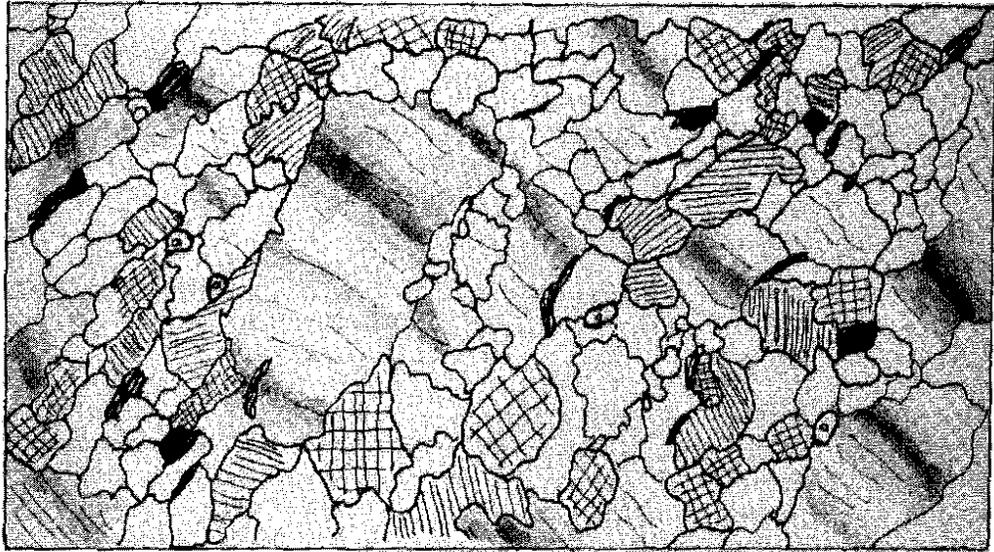
Veins of garnetiferous SCGIII are texturally similar to non-garnetiferous SCGIII, with large and ragged Kspar containing quartz inclusions and being invaded by myrmekite. Garnets in these veins are often partially atollated (see also Fig. 11), with concentric inclusions of quartz. These garnets are mantled by sillimanite  $\pm$  biotite, now partially altered to muscovite and chlorite. Myrmekitic intergrowths of plagioclase and quartz, together with tiny anhedral grains of Kspar, surround the sillimanite  $\pm$  biotite rim (Fig 11). Quartz in the garnetiferous facies of SCGIII is highly strained but not recrystallised.

Diagrammatic sketches to illustrate the textures of SCGI-III are presented in Fig. 12. The above descriptions indicate that SCGIII (both garnet-bearing and garnet-free) contains relict igneous textures which have been destroyed by high-temperature deformation in SCGI and SCGII. This fits with field observations (above) that SCGIII cross-cuts fabrics in SCGI and SCGII.

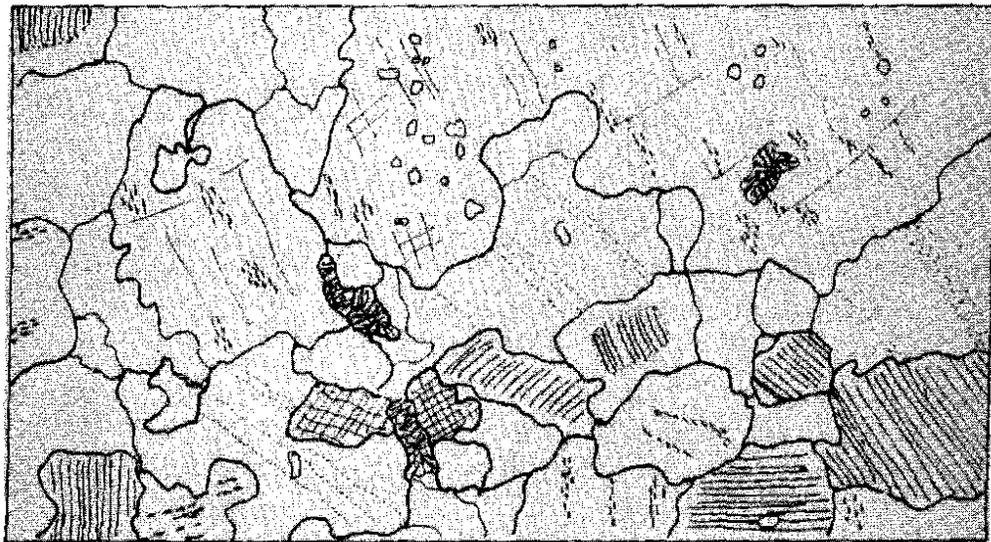


SCGI ┘

Figure 12. Schematic thin section sketches (a) SCGI, (b) SCGII, (c) SCGIII.



SCGII ┘



SCGIII ┘

## Mineral Chemistry

### Feldspars

Representative feldspar compositions (obtained by microprobe analysis with a 20 $\mu$  beam) from samples of SCGI-SCGIII are shown in Table 3 and plotted on Fig. 7. Ternary feldspar isotherms on Fig. 7 are from Whitney & Stormer (1977) for low structural state feldspars. Peak temperatures (obtained from the most sodic Kspar in a sample) range from ~ 680°C (at P = 5kb) to ~ 720°C (at P = 10kb). These temperatures are minimum estimates due to the possibility of re-equilibration. In general, peak temperatures from SCGIII samples (including the garnetiferous SCGIII sample no. 7) are ~ 100°C higher than the peak temperatures from SCGI and SCGII samples. This difference is probably more apparent than real; rather than indicating any real temperature gradient between the different groups (which is physically implausible) it probably reflects variations in the extent of post-peak re-equilibration. All of the analysed samples show some evidence of resetting to lower temperatures (~ 500°C), as shown by the occurrence of grains of less sodic Kspar. The difference in peak temperatures between groups SCGI, SCGII and SCGIII probably relates to more extensive solid-state homogenisation in the former. This interpretation is supported by the textural differences noted above, whereby SCGI and SCGII have solid-state deformation textures whereas SCGIII retains igneous textures.

Consistent zoning of plagioclase was noted only in samples V and 3 of group SCGIII, where zoning ranged from Ab<sub>78</sub> to Ab<sub>94</sub> (core to rim) and from Ab<sub>74</sub> to Ab<sub>86</sub> (core to rim) respectively.

### Garnet

The typical composition of garnets in garnetiferous SCGIII veins is Py<sub>9.5</sub> Alm<sub>80</sub> Gr<sub>4</sub> Sp<sub>6.5</sub>; for comparison, garnet in metagreywackes of the Plainfield Formation averages Py<sub>20</sub> Alm<sub>75</sub> Gr<sub>3</sub> Sp<sub>2</sub> (Table 3). Thus garnets in the veins are notably richer in CaO and MnO than those in the country rock. Vein garnets have considerably lower M/FM (0.106) than country rock garnets (0.211).

Comparative zoning profiles of vein and country rock garnets are shown in Figs. 7 and 11. Country rock garnets have very complex zoning whereas vein garnets are essentially unzoned. This cannot be an effect of diffusive homogenisation which would preferentially homogenise the smaller (country rock) garnets, contrary to observation. Thus it appears that vein garnets are not xenocrysts of country rock.

Allen and Clarke (1981) separate xenocrystic garnets from magmatic garnets in the South Mountain Batholith of Nova Scotia by composition, Py<sub>9.5</sub>Alm<sub>80</sub>Gr<sub>1.3</sub>Sp<sub>6.6</sub> vs. Py<sub>6</sub>Alm<sub>76</sub>Gr<sub>2</sub>Sp<sub>16</sub>. Similarly spessartine-rich compositions (Py<sub>3.5</sub> Alm<sub>61.5</sub> Gr<sub>2.5</sub> Sp<sub>32.5</sub>) characterise magmatic garnets in the Ruby Mountains (Kistler et al., 1981). Not all magmatic garnets are so spessartine-rich; for example, samples from the Kinsman Quartz Monzonite of New Hampshire (Clark, 1977) average Py<sub>22</sub> Alm<sub>73</sub> Gr<sub>2</sub> Sp<sub>2</sub>. Thus spessartine content alone may not discriminate magmatic and metamorphic garnets, and the relatively low spessartine content (6.5%) of sample 7 garnets does not necessarily imply a non-magmatic origin.

The abundant quartz inclusions in the SCGIII garnets are compatible with synchronous crystallisation of garnet and quartz from a melt. Garnet-biotite tie-lines for sample 7 are shown in AFM projection in Fig. 13 in comparison to tie-lines for garnet-biotite pairs in surrounding metamorphic rocks. The crossing tie-lines for rocks of similar grade are suggestive of very different modes of origin, i.e. magmatic vs. metamorphic.

Abbott and Clarke (1979) present a variety of AFM liquidus topologies to allow crystallisation of garnet from granitic liquids; these are shown below as Fig. 14. Topologies A and B of Fig. 14 allow crystallisation of both garnet and sillimanite from liquid, and may allow the observed mantling of garnet by sillimanite in sample 7. Both topologies require  $T > 700^{\circ}\text{C}$ ; A (cordierite-absent) is favoured over B at  $P > 7\text{kb}$ . These conditions ( $P > 6\text{kb}$ ,  $T > 700^{\circ}\text{C}$ ) were apparently attained in the Stony Creek area (see "Metamorphic Framework" section). It is therefore plausible that the garnets in SCGIII crystallised from a melt.

### Biotites

Representative biotite compositions from SCGII and SCGIII are shown in Table 3. The high (~ 3.2%)  $\text{TiO}_2$  contents suggest that even in strongly deformed SCGII much original igneous mineral chemistry is retained. Particularly striking is the very low M/IM of biotites from sample 7 (which contains garnet interpreted as igneous). Biotites of peraluminous granites are characterised by tetrahedral Al contents of 2.5-2.8 per 22 oxygens, and by  $\text{MgO}/\text{MgO} + \text{FeO} \sim 0.6-0.8$  (Clarke, 1981). Only biotite from sample 7 (SCGIII group) fits both criteria.

### Bulk Chemistry

XRF analyses of samples from units SCGI-SCGIII are presented in Table 4, together with published analyses of Narragansett Pier Granite, Hope Valley alaskite and Pelham dome gneiss (Dry Hill Gneiss) for comparison. Also shown for comparison are published analyses of a peraluminous garnet-granite from the Ruby Mountains and of a metamorphosed rhyolite from the Adirondacks. Previously published analyses of Stony Creek Granite (here = SCGI) and Branford Quartz Monzonite (= garnetiferous SCGI) are also shown in Table 4.

There is very little difference between SCGI samples and SCGII samples; marginally higher  $\text{Na}_2\text{O}$  in SCGII may be more apparent than real. Some SCGI samples have very high  $\text{K}_2\text{O}$  (~7%). To test proposed correlations of Stony Creek basement with Pelham Dome basement and Hope Valley basement, published analyses of these are shown. Only partial analysis of the Dry Hill Gneiss was reported (Hodgkins, 1983). The Dry Hill Gneiss is clearly richer in  $\text{Fe}_{\text{total}}$  and poorer in  $\text{CaO}$  (and possibly  $\text{Na}_2\text{O}$ ) than Stony Creek samples. Hope Valley alaskites (this study and Day et al., 1980) are typically higher in  $\text{SiO}_2$  and lower in  $\text{Al}_2\text{O}_3$ ,  $\text{CaO}$  and  $\text{Fe}_{\text{total}}$  than SCGI and SCGII. Similarly the Adirondack metarhyolite (which is chemically similar to Hope Valley alaskite) is higher in  $\text{SiO}_2$  and lower in  $\text{Al}_2\text{O}_3$ ,  $\text{CaO}$  and  $\text{Fe}_{\text{total}}$  than SCGI and SCGII.

Surprisingly, the garnetiferous SCGI sample no. 6 shows very little difference from garnet-free SCGI. Published analyses of Branford Quartz Monzonite, however, show markedly higher  $\text{Al}_2\text{O}_3$ : ( $\text{Na}_2\text{O} + \text{CaO} + \text{K}_2\text{O}$ ) consistent with the presence of garnet. Analyses of these show similarities to the garnet-2-mica granite from the Ruby Mountains.

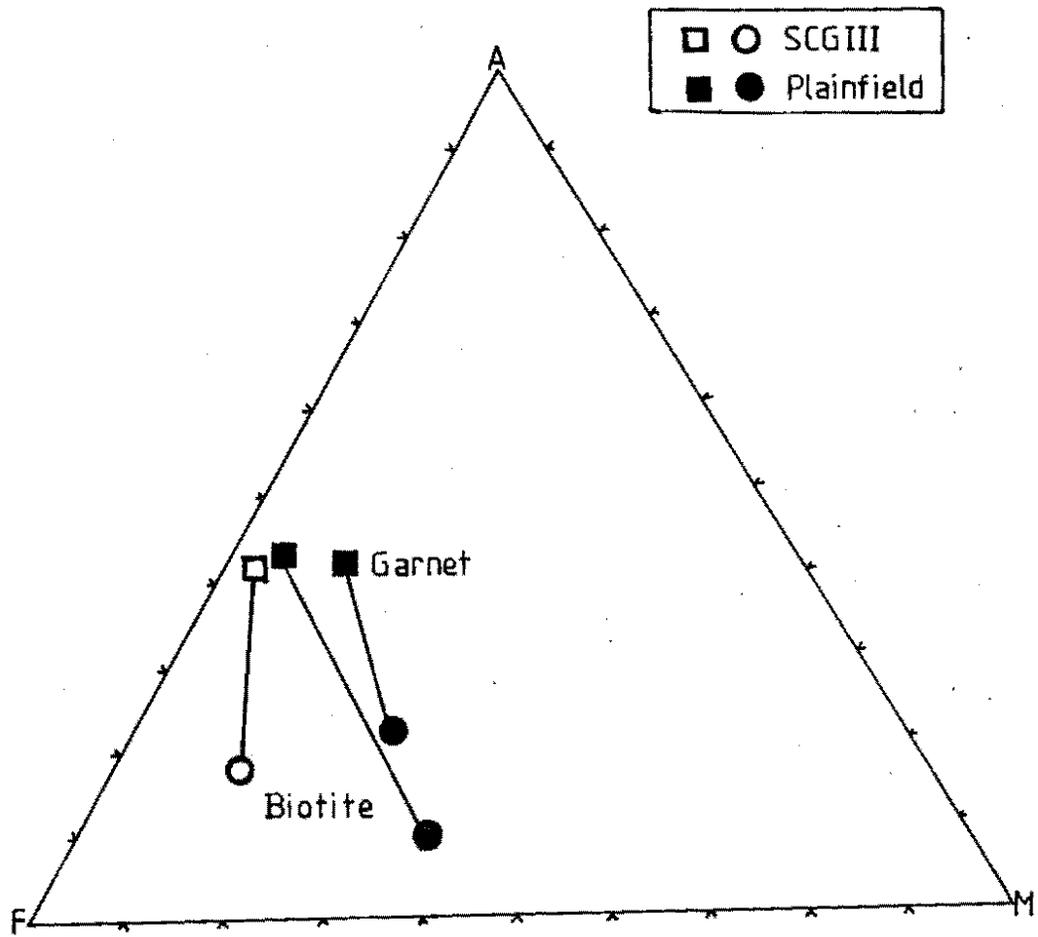


Figure 13. AFM projection of garnet-biotite tielines for 2 country-rock samples and for SCGIII sample 7.

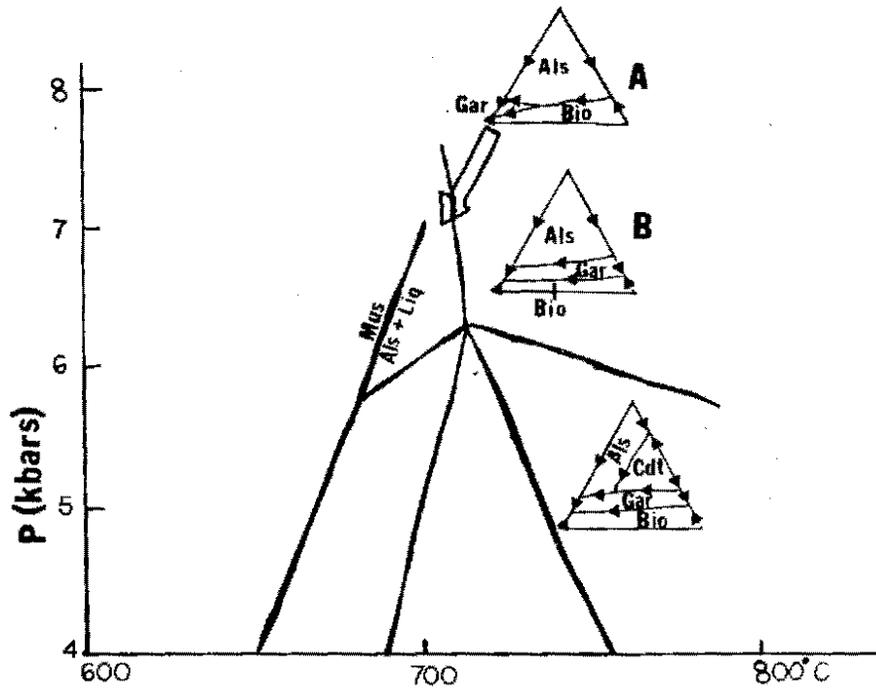


Figure 14. AFM liquidus topologies for garnet-bearing melts, from Abbott and Clarke (1979).

Samples of SCGIII are typically richer in FeO and MgO than are SCGI and SCGII. The very high FeO content of sample 5V correlates with very low SiO<sub>2</sub>; the reason for this anomalous analysis is unknown but may reflect sampling of a very magnetite-rich region. SCGIII samples appear more variable in composition than those of other groups. It is therefore difficult to compare them with possibly-correlative Narragansett Pier Granite; both groups occupy similar ranges of Fe<sub>total</sub>, CaO, Na<sub>2</sub>O and K<sub>2</sub>O. However, Narragansett Pier Granite has higher Al<sub>2</sub>O<sub>3</sub>:(Na<sub>2</sub>O + CaO + K<sub>2</sub>O) ratios.

### Phase Relations

Modal analyses of groups SCGI-SCGIII are plotted onto a series of Qz-Ab-Or phase diagrams in Fig. 15. None of the samples lie close to the eutectics (at 5 or 10kb) for the An-free system. However, with the exception of sample V whose unusual chemistry has already been noted, samples of SCGIII show a clustering around the eutectic for the system with Ab/An = 3 at 5kb. The low-temperature valley in the Qz-Ab-Or system runs along the quartz-Kfeldspar cotectic (as shown by the liquidus isotherms for the An-free system at 5kb in Fig. 15). Samples from SCGIII generally lie along this Kfeldspar-quartz cotectic and thus, although not minimum-melt compositions, SCGIII samples probably represent relatively low-temperature melts. SCGII samples appear to occupy the positions of relatively higher-temperature melts further from the eutectic and cotectic. SCGI samples occupy a relatively wide area of the diagram.

Liquidus isotherms for the Ab/An = 3 system at P = 5kb are unknown; temperatures are anticipated to be slightly higher than for the An-free system. Note that all samples lie within the 740°C isotherm for the An-free system at P = 5kb; temperatures in the An<sub>25</sub> system are unlikely to be more than about 25°C higher (Johannes, 1985). It seems likely therefore that groups SCGI-SCGIII all lie within the 760°C isotherm; this is near the upper limit of temperatures recorded by metamorphic assemblages in the area. The data is consistent with, but does not demand, occurrence of SCGIII as melts at metamorphic temperatures where SCGI and SCGII were not melts. This interpretation is open to question, as discussed below.

Firstly, identification of SCGIII samples as relatively low-melting compositions depends on correct identification of their relatively higher plag:Kspar ratios. In view of coarse grain size (and consequent sampling problems) and the use of unstained sections, this may need further revision; analysis of additional, stained sections is planned. Secondly, the locations of eutectics, cotectics and isotherms are less well-known than Fig. 15 implies. Johannes (1985) suggests that most published experimental data are for disequilibrium melting and so may be of limited value for petrogenetic interpretations. Thirdly, the accuracy of geothermometers is inadequate to resolve the ~ 20°C difference in melting temperatures between group SCGI + SCGII and group SCGIII proposed in Fig. 15. As noted earlier ("Petrography") SCGIII retains textural evidence of melting which has been lost during deformation in SCGI and SCGII. This is compatible with the suggestion above that SCGIII were molten in preference to SCGI and SCGII. However, if metamorphic temperatures of 770°C are real, samples of all groups would have been molten during metamorphism, as all lie within the 760°C isotherm of Fig. 15.

This suggests that factors other than composition in terms of Qz:Ab:An:Or were important in determining the behaviour of the various granitic units.

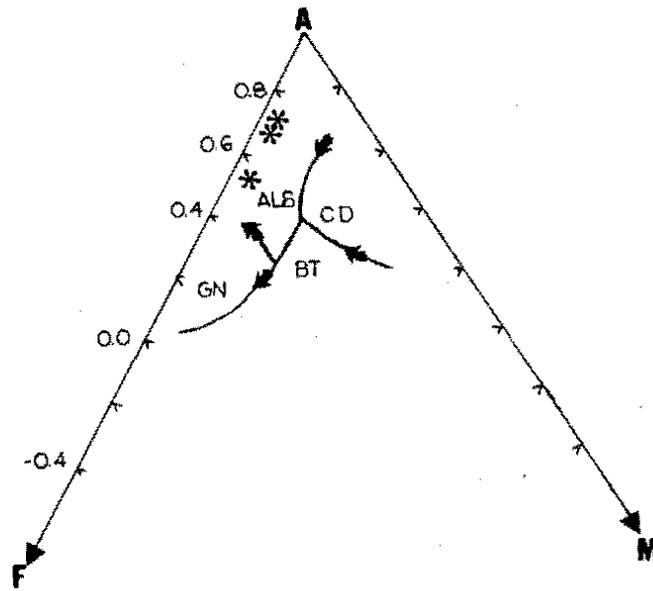


Figure 16. AFM projection of SCGIII samples compared to liquidus fields or sill. and gt. from Abbott and Clarke (1979).

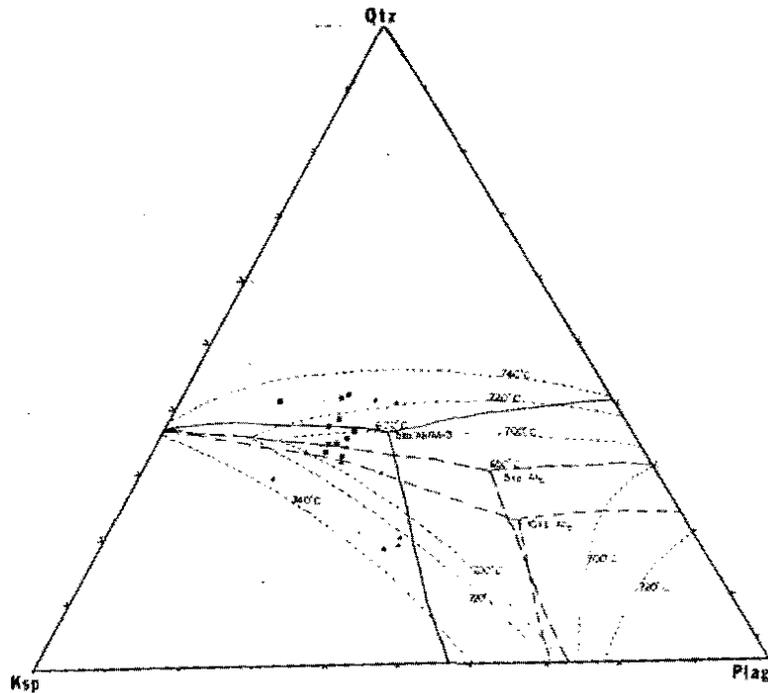


Figure 15. Modes of SCG1-SCGIII in Qz-Ab-Or compared to experimental data. Eutectics for An at 5 and 10 kb from Johannes (1985); eutectic for Ab/An = 3 from Winkler (1976). Liquidus contours for 5 kb from Luth and Tuttle (1964).

SCGIII occurs both as pegmatites outside the envelope of Plainfield and in association with SCGI inside it. The wide range of compositions may argue for a hydrothermal-vein origin, possibly derived from SCGI. SCGIV is restricted to association with SCGII and may form by (hydrothermal or anatectic) remobilisation of it.

#### SPECULATIONS

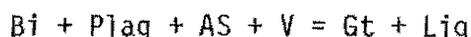
The reader will have noticed that, in the Stony Creek dome, at the present stage of investigation, the method of multiple hypothesis may be more aptly re-named "the method of multiple confusion". Undeterred by conflicting data, or indeed the lack of data, we present for amusement the following speculative history. The Stony Creek dome represents a far-travelled slice of Hope Valley--correlative basement, subject to deformation, metamorphism (with possible anatexis-SCGII) and intrusion (SCGI) in the Pre-Cambrian. This basement was sutured to North America before or during the Acadian (gneiss dome tectonics). Subsequent Alleghanian effects included ductile shearing and remobilisation (hydrothermal or anatectic) of the existing granitic units to form SCGIII and SCGIV. We look forward to animated discussions!

Possible factors favouring melts of SCGIII include:

- (i) higher MgO and/or FeO--this has been shown to lower the solidus by ~ 30°C (Naney, 1983)
- (ii) higher B or F (Manning, 1981)
- (iii) higher P<sub>2</sub>O<sub>5</sub> (Johannes, 1985)--although P<sub>2</sub>O<sub>5</sub> was present in only trace amounts, and apatite is more abundant in SCGII than in SCGIII.

Bulk analysis of one SCGIII garnetiferous vein and of published analyses of the Branford Quartz Monzonite are shown in AFM projection in Fig. 16. Also shown are Abbott and Clarke's (1979) schematic liquidus fields for P > 5kb, T > 700°C. All samples contain garnet, none contain Al<sub>2</sub>SiO<sub>5</sub> (although sample no. 7, which is otherwise identical to the plotted sample, does contain sillimanite). All samples plot within the Al<sub>2</sub>SiO<sub>5</sub> liquidus field. Boundaries in Fig. 16 are not well-constrained, so the significance of this is unknown.

Sample no. 7 contains textures suggestive of reaction of the form Gt + Liq → Sill ± Bi; this is incompatible with Abbott and Clarke's topologies. However, Clemens and Wall (1981) note that at temperatures ~ 700°C the reactions:



and

$$\text{Bi} + \text{Plag} + \text{AS} + \text{Qz} = \text{Gt} + \text{Kspar} + \text{Liq}$$

can occur. Such reactions, operating during melt crystallisation, could generate the observed textures. Reaction temperatures lie well below peak temperatures for the area (Fig. 6).

#### Possible Origins of Units SCGI-SCGIV

Field evidence (the cross-cutting relations and apparent assimilation of xenoliths) strongly suggest that SCGI was intrusive. No explanation can be given for the local development of a more garnetiferous facies (STOP 7). This is restricted to a level in the dome where Middle Plainfield assimilation might be anticipated, and so local assimilation of pelites and consequent raising of the Al<sub>2</sub>O<sub>3</sub>: (Na<sub>2</sub>O + CaO + K<sub>2</sub>O) ratio of the magma might be postulated. Sample 6 however gave a surprisingly low Al<sub>2</sub>O<sub>3</sub>: (Na<sub>2</sub>O + CaO + K<sub>2</sub>O ratio). If the Branford Quartz Monzonite is a correlative of garnetiferous SCGI, the hypothesis of preferential assimilation breaks down; the Branford masses occur at the level of the Mamacoke Formation.

SCGII is chemically very similar to SCGI, but is always layer-conformable and is of limited vertical extent. In field relation, it is very similar both to the alaskites of the Lyme dome (Lundgren, 1967) and to alaskites of the Adirondacks (Carl and van Diver, 1975). The latter are interpreted as metamorphosed rhyolites. Chemically, SCGII is distinct from these (lower SiO<sub>2</sub>, higher Al<sub>2</sub>O<sub>3</sub>, CaO and Fe<sub>total</sub>). The association with migmatized lower Plainfield Formation and the occurrence of possibly restitic garnet stringers is consistent with formation by anatexis. This will be investigated further.

TABLE 1  
STRATIGRAPHY

MONSON GNEISS

Massive or, less commonly, layered gneisses. Dark grey, medium-grained qz-pl-bi-hb gneisses, probably intrusive.

MAMACOKE FORMATION

Massive or layered, fine-grained, light grey, qz-pl-bi±hb gneisses. "Striped" appearance in outcrop. Rare garnetiferous gneisses.

PLAINFIELD FORMATION

Upper Well-bedded massive quartzite, minor calc-silicates and amphibolites.

Middle Foliated grey pelitic gneisses, often migmatitic. Very variable; minor quartzite.

Lower Highly migmatitic garnetiferous gneisses, garnetiferous amphibolites.

TABLE 2

MODAL ANALYSIS

SCGI Samples

	SCGI/1	SCGI/2	SCG/1	SCG/2	7/3
Qz	38	28	33	35	41
Plag	20	24	23	24	28
Kspar	33	38	40	39	25
Bi	8	9	3	1	6
Mt	1	1	tr.	1	tr.
Ap	tr.	tr.	1	tr.	tr.
Others	tr.	tr.	tr.	tr.	-

SCGII Samples

	19/1	19/2	20/1	20/2
Qz	31	34	35	39
Plag	21	21	21	14
Kspar	35	39	33	43
Bi	9	4	5	2
Mt	3	2	4	2
Ap	1	tr.	2	tr.
Others	tr.	tr.	tr.	tr.

SCGIII Samples

	2	3/1	3/2	6/1	6/2	6/3	V/1	V/2	V/3	7/1	7/2
Qz	28	32	28	32	36	39	18	17	19	27	26
Plag	18	25	18	23	22	22	39	37	40	23	28
Kspar	51	39	39	42	36	31	37	40	37	35	32
Bi	2	3	10	3	4	6	3	2	2	3	3
Mt	1	1	3	-	1	tr.	3	3	2	tr.	tr.
Ap	tr.	tr.	2	-	tr.	tr.	-	1	tr.	tr.	tr.
Others	tr.	-	tr.	-	(Gt)1	(GT)1	-	-	tr.	12	11

A3-24



TABLE 4

## BULK CHEMISTRY

## SCGI Samples

	SC	32C	SCGI	6
SiO <sub>2</sub>	68.68	74.23	71.24	70.25
Al <sub>2</sub> O <sub>3</sub>	16.20	14.49	14.43	15.11
Fe <sub>2</sub> O <sub>3</sub>	1.64	2.19	2.29	2.22
MgO	0.09	0.10	0.22	0.01
CaO	1.55	1.33	2.36	1.63
Na <sub>2</sub> O	4.20	2.29	4.67	5.09
K <sub>2</sub> O	7.31	4.97	4.36	5.11
TiO <sub>2</sub>	0.28	0.28	0.32	0.28
MnO	0.01	0.02	0.03	0.06
Al/Alk	1.24	1.68	1.27	1.27

## Possible correlatives to SCGI

	SCG <sup>1</sup>	TRG <sup>2</sup>	BQM <sup>3</sup>	BQM <sup>3</sup>	RM <sup>4</sup>
SiO <sub>2</sub>	70.19	73.00	74.29	72.47	73.60
Al <sub>2</sub> O <sub>3</sub>	15.66	13.60	14.74	14.78	14.00
Fe <sub>2</sub> O <sub>3</sub>	0.72	2.98	0.20	0.57	0.80
FeO	0.81		0.65	2.00	1.40
MgO	0.24	0.56	0.15	0.34	0.18
CaO	1.23	1.51	0.80	1.27	0.92
Na <sub>2</sub> O	2.87	3.21	2.70	4.03	2.90
K <sub>2</sub> O	7.41	4.47	4.62	4.53	4.70
TiO <sub>2</sub>	0.15	0.31	0.10	-	0.06
MnO	0.01	0.07	0.06	0.48	0.01
Al/Alk	1.36	1.48	1.82	1.50	1.64

## SCGII Sample

	20
SiO <sub>2</sub>	70.71
Al <sub>2</sub> O <sub>3</sub>	14.48
Fe <sub>2</sub> O <sub>3</sub>	1.89
MgO	0.07
CaO	1.63
Na <sub>2</sub> O	5.29
K <sub>2</sub> O	5.34
TiO <sub>2</sub>	0.41
MnO	0.12
Al/Alk	1.18

## Possible correlatives to SCGII

	HVA <sup>5</sup>	DH <sup>6</sup>	A <sup>7</sup>	HVA <sup>8</sup>	HVA <sup>8</sup>	HVA <sup>8</sup>	HVA <sup>8</sup>
SiO <sub>2</sub>	73.61	n.a.	76.51	76.08	76.70	76.10	72.85
Al <sub>2</sub> O <sub>3</sub>	15.25	n.a.	10.60	11.09	11.06	12.99	13.12
Fe <sub>2</sub> O <sub>3</sub>	0.87	3.42	2.71	0.62	0.47	1.03	1.02
FeO				1.58	1.34	0.57	0.56
MgO	0.04	0.08	0.18	0.10	0.07	0.04	0.32
CaO	1.08	0.67	0.62	0.79	0.60	0.75	0.89
Na <sub>2</sub> O	3.98	2.73	3.53	2.96	2.98	3.98	3.56
K <sub>2</sub> O	4.98	4.42	4.40	5.28	5.41	5.03	4.83
TiO <sub>2</sub>	0.14	n.a.	-	0.21	0.17	0.12	0.29
MnO	0.06	n.a.	0.01	0.07	0.05	0.02	0.05
Al/Alk	1.53	n.a.	1.24	1.23	1.23	1.33	1.41

## SCGIII Samples

	2	5V	32V	61V	V
SiO <sub>2</sub>	70.83	57.38	71.32	70.35	71.66
Al <sub>2</sub> O <sub>3</sub>	14.57	15.89	14.78	16.20	17.57
Fe <sub>2</sub> O <sub>3</sub>	2.27	7.81	1.54	2.54	2.45
MgO		0.09	0.11	0.10	0.08
CaO	1.78	1.55	1.24	1.66	2.63
Na <sub>2</sub> O	5.10	3.77	4.53	3.69	1.57
K <sub>2</sub> O	5.00	9.13	5.97	5.02	3.65
TiO <sub>2</sub>	0.42	0.57	0.25	0.31	0.35
MnO	0.03	0.05	0.02	0.09	0.04
Al/Alk	1.22	1.09	1.26	1.56	2.24

## Possible correlatives to SCGIII

	NPG <sup>9</sup>	NPG <sup>10</sup>	NPG <sup>10</sup>	NPG <sup>10</sup>	NPG <sup>11</sup>	NPG <sup>11</sup>
SiO <sub>2</sub>	70.37	71.96	70.45	70.59	74.30	74.89
Al <sub>2</sub> O <sub>3</sub>	15.75	14.26	14.09	15.76	15.18	14.71
Fe <sub>2</sub> O <sub>3</sub>	2.04	0.86	0.79	1.13	0.39	0.01
FeO		1.29	0.96	0.69	0.26	0.21
MgO	0.13	0.49	0.34	0.42	0.18	0.07
CaO	1.77	1.58	1.18	1.14	1.05	0.77
Na <sub>2</sub> O	4.62	3.37	3.55	4.08	3.27	3.75
K <sub>2</sub> O	4.93	5.06	5.13	4.99	4.35	4.88
TiO <sub>2</sub>	0.39	0.45	0.24	0.29	0.11	0.08
MnO	0.03	0.03	0.02	0.03	0.01	0.01
Al/Alk	1.39	1.42	1.43	1.54	1.75	1.56

1. Stony Creek Granite from Mikami & Digman (1957)
2. Ten Rod Granite Gneiss from Day et al. (1980)
3. Branford Quartz Monzonite from Mikami & Digman (1957)
4. Ruby Mountains 2-mica-granite granite from Kistler et al. (1981)
5. Hope Valley Alaskite, this study
6. Dry Hill Gneiss (Pelham Dome) from Hodgkin (1981)
7. Adirondack alaskite from Carl & von Diver (1980)
8. Hope Valley Alaskite, from Day et al. (1980)
9. Narragansett Pier Granite, this study
10. Narragansett Pier Granite, Hermes et al. (1981)
11. Muscovite-garnet Narragansett Pier Granite, Hermes et al. (1981)

PLATE CAPTIONSPLATE 1

SCGI from STOP 3, showing phase layering and biotite schlieren.

PLATE 2

Close-up of SCGII from STOP 3, showing lack of internal fabric.

PLATE 3

Veins of garnetiferous (?) SCGII in garnetiferous(?) SCGI, STOP 7.

PLATE 4

Migmatized Lower-Plainfield from STOP 8. Note marginal selvages, very open folds and cross-cutting pegmatite.

PLATE 5

SCGIII from STOP 9, showing very strong L fabric; flattened aplitic layer parallels this.

PLATE 6

Close-up of SCGIII at STOP 9, showing garnet + biotite stringers.

PLATE 7

SCGIII from STOP 12, with feldspar augen forming strong L fabric, warped into shear zone.

PLATE 8

SCGIII from STOP 12, deformed in shear zones; SCGIV concentrated in these.

PLATE 9

Migmatized Lower Plainfield with pegmatite, possibly equivalent to SCGII, STOP 12.

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## ROAD LOG AND STOP DESCRIPTION

Assembly time: 8:00 a.m.

Assembly point: Branhaven Plaza, Branford.

If coming from New Haven and west, take exit 53 from I-95 (Connecticut Turnpike). Turn right on U.S. Rte 1 west - Branhaven Plaza is approximately 0.25mi on left.

If coming from New London, Rhode Island or most points east of New Haven, take exit 55 from I-95 (Connecticut Turnpike). Turn right at stop sign onto U.S. Rte 1 west, drive 2.8mi. Turn left at lights into Branhaven Plaza.

There will be an opportunity to buy lunch supplies between STOPS 4 and 5. Lunch is scheduled for the scenic attractions of STOP 9, but may be brought forward to STOP 7 (by a consensus of opinion) if hunger intervenes. Please consolidate vehicles as much as possible; many of the stops have extremely limited parking. We will be returning to Branhaven Plaza at the end of the trip to collect vehicles left there.

<u>Mileage</u>		
<u>Interval</u>	<u>Total</u>	
	0.0	Turn right out of plaza on U.S.1E and follow signs for I-95N.
0.3	0.3	Bear left: follow signs for I-95N.
0.8	1.1	Turn left at light for I-95N.
0.1	1.2	Turn right for I-95N (signed for "Rhode Island and East").
		Several outcrops of Branford Granite.
2.9	4.1	Take exit 56 off I-95N (signed for "Stony Creek").
0.3	4.4	Turn right at stop sign onto Thimble Island Road.
0.8	5.2	<u>STOP 1</u>
		Turn left down Flat Rock Road and park. Walk back up Flat Rock Road, cross Thimble Island Road to low flat outcrops. We are here near the northwestern margin of the Stony Creek dome. The outcrop is dominated by units of the Upper Plainfield Formation, particularly the well-bedded quartzite member. Rafts of this,

several meters in length, are surrounded by anastomosing veins of granite belonging to the SCGI unit. The granite locally reaches 40 volume percent, and locally contains coarser subpegmatitic patches. Relict stratigraphy is preserved here.

The SCGI here shows a mineral shape fabric which wraps around enclaves of Upper Plainfield. Flattening strains here (discussed in the introduction) are attributed to ballooning during emplacement of SCGI.

Walk 0.3mi down Flat Rock Road past vehicles and up gentle hill. After 9th house on left, road forks; bear left and turn into first driveway on left. Walk to low flat outcrops behind house.

#### STOP 2.

Northwestern margin of Stony Creek dome. Enclaves of Upper Plainfield in SCGI. Upper Plainfield here is more varied than at STOP 1 and includes biotite-schists, amphibolites and semi-pelites together with quartzite. Granite: enclave ratio varies from 20:80 to 60:40. Some parts of the outcrop show relict stratigraphy as at STOP 1, and here foliations can be traced continuously from enclave to enclave. Elsewhere in the outcrop, however, internal fabrics cannot be traced continuously between enclaves; relative rotations of up to 60° between adjacent blocks can be documented (see sketch below). Such rotation can apparently occur without disrupting the larger-scale stratigraphy. Towards the base of the outcrop both fabric and stratigraphy are disrupted, with various lithologies chaotically jumbled in apparently random orientation (see sketch). The extent of stratigraphic disruption does not seem to be solely controlled by the relative volumes of host and enclave.

Amphibolitic and biotitic enclaves contain aplitic layers which commonly show biotite-rich selvages. These aplitic layers are folded, and the folds themselves are truncated at the margins of the enclaves (see sketch). Amphibolitic layers are commonly sub-equant (axial ratios ~ 1:1:1) whereas semi-pelitic enclaves are more tabular (axial ratios ~ 2:5:1:1). This is presumably due to variations both in original lithology and in competence contrast with the host.

The host is granite of SCGI type, here ~ 5mm in grain-size and without pronounced fabric. Locally coarser patches occur. Enclaves are commonly cut by small (~ 3cm.) veins of this coarser, undeformed SCGI.

We expect debate to focus on possible relations between the aplitic layers in the enclaves and the granite surrounding the enclaves. If these are gradational, why do the aplitic layers appear to define folds whilst the granite appears undeformed? Are the aplites mimetic after an earlier folded fabric? What is the significance of the biotite selvedge?

We also wish to discuss whether the enclaves are of tectonic origin (i.e. are boudins) or of magmatic origin (i.e. are xenoliths). Do the folds indicate intrusion of magma into previously-deformed rocks? Are folding and boudinage successive stages in a single, non-coaxial deformational event, or do they represent very different deformational events?

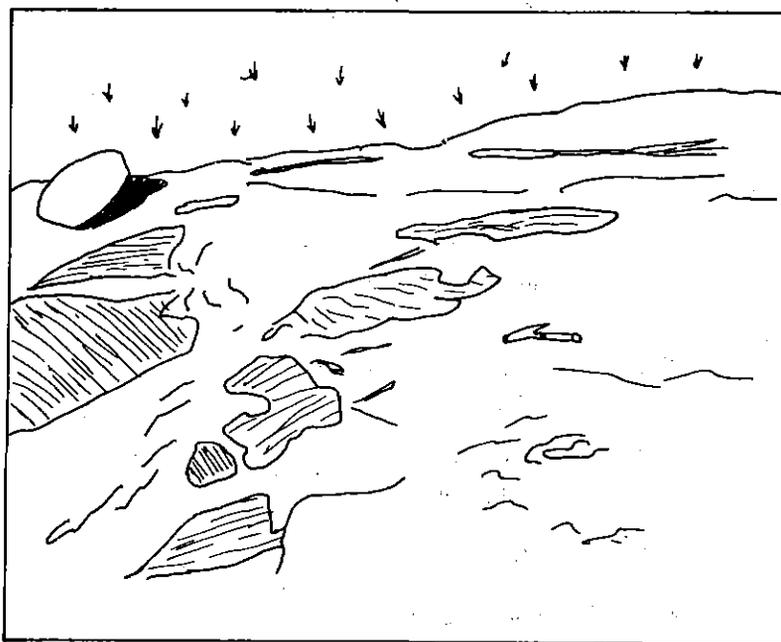


Fig. 17 STOP 2. Chaotic mixing of Upper Plainfield lithologic types in host of SCGI.

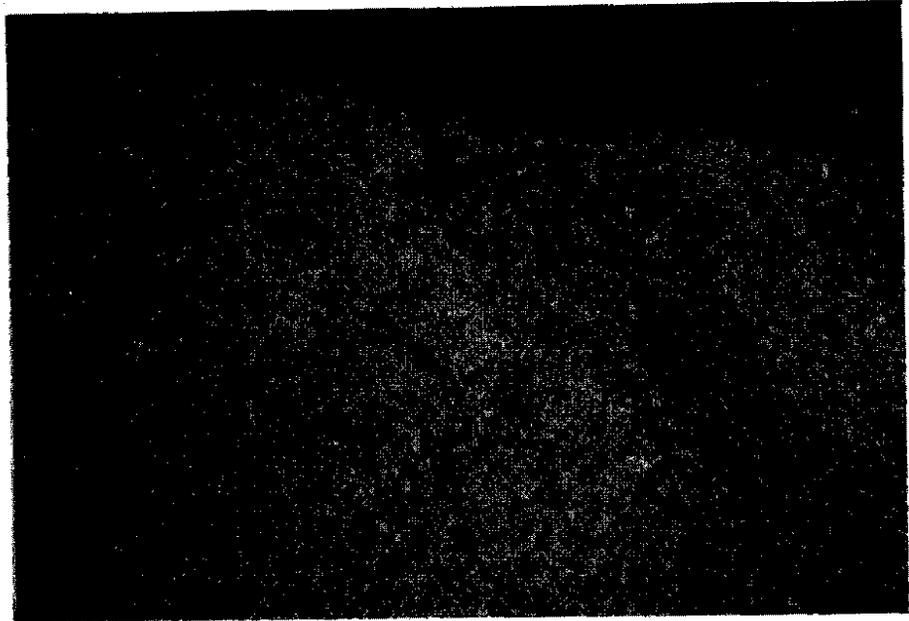


Plate 1 STOP 3. Biotite schlieren in SCGI from quarry.

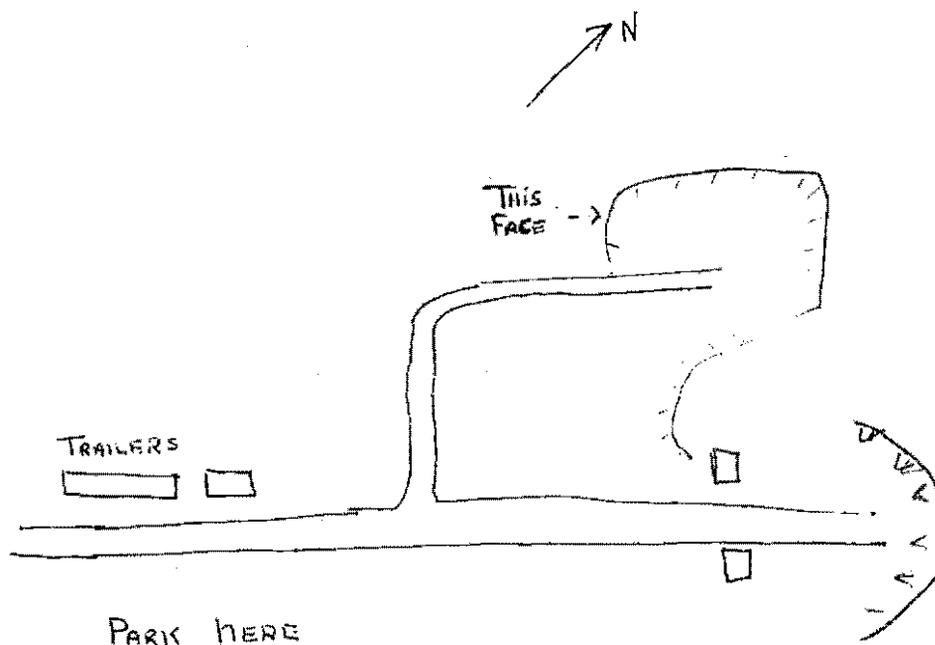
Return to vehicles

- |     |     |  |
|-----|-----|--|
| 0.0 | 5.2 | Turn left out of Flat Rock Road onto Thimble Island Road.  |
| 0.8 | 6.0 | Stop sign. Junction of Thimble Island Road and Conn. Rte. 146. Turn left onto Conn. Rte. 146.  |
| 0.6 | 6.6 | Turn left on Quarry Road   |
| 0.6 | 7.3 | Road forks; bear left through stop sign. (Gate here is closed at 3 p.m. daily.)  |
| 0.1 | 7.4 | Large metal gate; closes at 3 p.m. daily. Continue through.  |
| 0.1 | 7.5 | Park vehicles across from trailers. (If following this trip later, check in with quarry manager in large blue trailer.) Do not enter quarry without supervision, |

i.e. after 3 p.m. or at weekends. Do NOT climb quarry faces and exercise great care when walking on loose blocks--not all of these are stable.

STOP 3.

The main working area of the quarry is to your left; follow track ~ 80m. towards working face. Examine the closer of the two faces striking NW-SE (see sketch).



One wall of the quarry shows numerous partially digested blocks of Upper Plainfield quartzite, biotite schist and amphibolite. It was by tracing such relict stratigraphy that Sanders (1968) was able to map out the Sachem's Head Anticline. This wall also shows the well-developed phase layering of SCGI marked by alternating quartz- and feldspar-rich layers, attributed by Mikami and Digman (1957) to filter-pressing.

Sub-parallel to this are schlieren ~ 2" thick of more mafic material, predominantly biotite + plagioclase. These may either be the products of flow segregation

or highly flattened and assimilated xenoliths. If interpreted as xenoliths, their great lateral continuity would require extreme flattening: the blocky nature of observed enclaves may argue against such extreme flattening. However, such biotite + plagioclase-rich lithologies can be seen in the process of formation during assimilation of amphibolite xenoliths elsewhere in the quarry (Plate 1 and sketch below). This may support a xenolithic origin for the schlieren.

The layering in SCGI can be seen to be deformed by a series of evenly-spaced (~ 6" separation) shear zones. These dip at a low angle and are suggestive of dextral (top to NW) shear. Large (up to 18") veins of coarsely crystalline, pink-weathering SCGIII cut phase layering in SCGI (sketch below). The margins of these veins show signs of flexure in the shear zones; however, the veins have no internal fabric. The veins commonly have a biotite + plagioclase-rich selvage. In addition to its occurrence as veins, SCGIII also occurs as irregular patches within SCGI, gradational into SCGI by a decrease in %Kspar and in grain size (see Plate 2).

Phase layering and successive stages in the assimilation of enclaves can be clearly seen in the gigantic blocks near the trailers. Foliation in SCGI (Plate 1) can be clearly seen to be a composite of biotite schlieren, phase layering and mineral aspect ratio.

If time permits, we will walk down the path to the right of the working area (the continuation of the entrance road) to the abandoned face. This face shows steeply dipping wedges of SCGI in the Upper Plainfield. This lit-par-lit structure, together with the assimilation so well displayed in blocks here, is suggestive of emplacement of SCGI by stoping which in turn implies a thermal contrast between granite and country rock and/or high-level intrusion. However, the flattening strains seen at the margin are more suggestive of diapirism, i.e. no thermal contrast and/or deep-level emplacement.

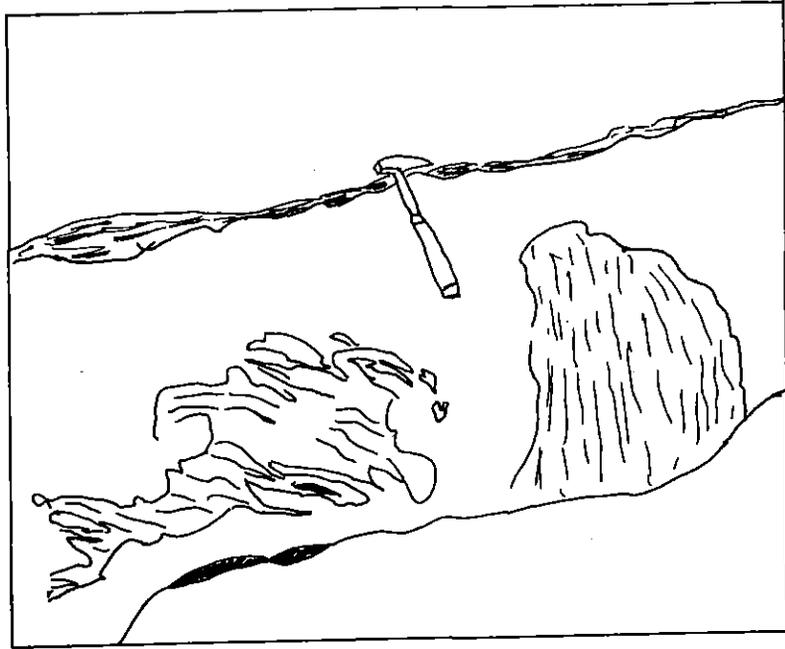


Figure 18 STOP 3. Enclaves of amphibolite of Upper Plainfield showing rotation of internal fabric and partial assimilation to form bi + plag-rich borders.

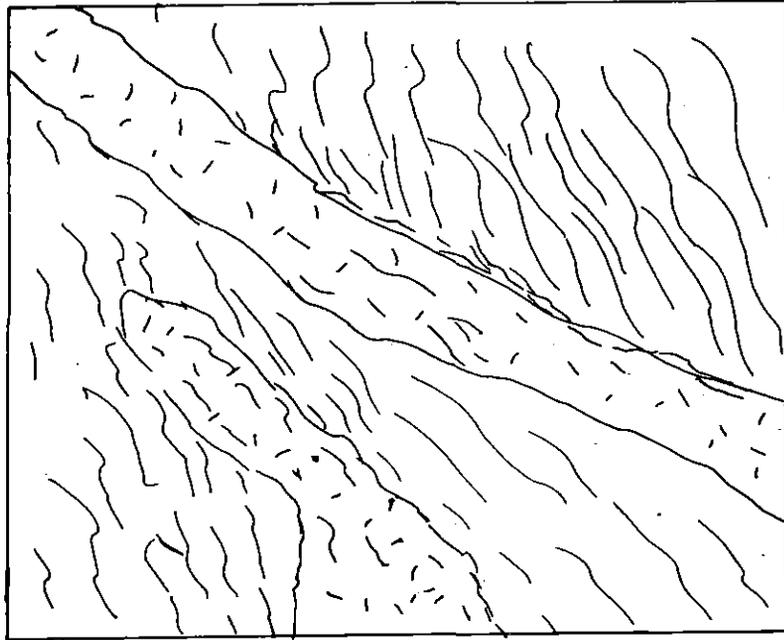


Figure 19 STOP 3. Veins of SCGIII cut phase layering in SCGI. Note selvedges.

Return to vehicles.

- |     |      |  |
|-----|------|--|
| 0.0 | 7.5  | Drive back down quarry road through gates.   |
| 0.9 | 8.4  | Junction of Quarry Road and Conn. Rte. 146 (Leetes Island Road). Turn left onto Conn. Rte. 146.  |
| 0.6 | 9.0  | Outcrops of homogeneous SCGI on left are similar to those we will see at the next stop.  |
| 0.4 | 9.4  | SLOW DOWN! Very sharp curve through railroad underpass, road often slippery. Oncoming traffic often occupies full width of road.   |
| 0.8 | 10.2 | Moose Hill Road on left, Shell Beach Road on right. Turn left on Moose Hill Road.  |
| 0.4 | 10.6 | Dromora Road on left; continue on Moose Hill Road.   |
| 0.4 | 11.0 | Top of hill. Park at side of road, across from walled-in sheep pasture with conifers. Walk down drive at left and enter pasture by gate on left. (If doing this subsequent to NEIGC, please check with owner first.) |

#### STOP 4

(If following this roadlog in winter, you will be able to detect the outcrops even through several inches of snow, by the simple expedient of kicking one's toes against them. Local inhabitants are accustomed to the sight of geologists brushing away snow in this pasture.)

Core Facies of SCGI. The granular appearance on weathering is typical of outcrops of homogeneous SCGI in the core of the dome. Note the lack of enclaves, in contrast to STOPS 1, 2 and 3. The core of the dome generally consists of such relatively clean, massive granite; strains in this area are much lower than at the margins of the dome (cf. grain shape fabrics here and at STOP 3). Diffuse patches of coarser-grained granite may be local pegmatitic segregations of SCGI, or may be transitional to SCGIII as seen at STOP 3. Isolated very coarse (up to 10cm.) Kspar crystals occur; these may be boudinaged pegmatites, or local segregations.

Return to vehicles.

- |     |      |   |
|-----|------|---|
| 0.0 | 11.0 | Continue along Moose Hill Road.                         |
| 0.8 | 11.8 | Barker Hill Drive on left; continue on Moose Hill Road. |

- 0.7 12.5 Stop sign. Continue straight. Note change of name of road to Peddler's Road. Do not turn left on Moose Hill Road.
- 0.1 12.6 Outcrops of Mamacoke Formation at left are similar to those we will see at optional STOP 5A. Continue on Peddler's Road.
- 1.4 14.0 Junction of Peddler's Road and U.S. Rte. 1 (Boston Post Road). Turn right at stop sign.
- 0.4 14.4 Shopping mall on right is stop to buy lunch. "Dairy Mart" provides basic supplies and the "Grog Shop" is self-explanatory. The "Bishop Hill Farm Shop" which you passed at the junction of U.S. Rte. 1 and Peddler's Road is also a possible source of supplies.
- 0.0 14.4 Leave mall, turn left on U.S. Rte. 1 back past the junction with Peddler's Road.
- 0.4 14.8 Junction of U.S. Rte. 1 and Peddler's Road--continue straight on U.S. Rte. 1, following signs for I-95S.
- 0.8 15.6 Turn right onto I-95S (signed for "New Haven and west").

0.2 15.8 OPTIONAL STOP 5A.

Park on right at edge of turnpike. WATCH FOR TRAFFIC. We will have a police escort for this section of the trip. Pegmatitic veins of SCGIII (marked on the State map as "Narragansett Pier Granite") cut light-coloured gneisses of the Mamacoke and darker-coloured gneisses of the Monson. Pegmatitic veins of SCGIII are very common on the northern margin of the dome.

Continue on I-95S.

STOP 5.

- 0.5 16.3 Park on right at edge of turnpike. WATCH FOR TRAFFIC. Lit-par-lit structure between SCGI and Upper Plainfield (see sketch). Again note relatively steep radial dips at the dome margins. Veins of SCGIII sharply cross-cut both SCGI and Upper Plainfield. Some pegmatitic patches carry (possibly retrograde) muscovite. Biotite schist layers contain qz+plag+Ksp+gt±sulfide veins--how do these relate to the other granites?

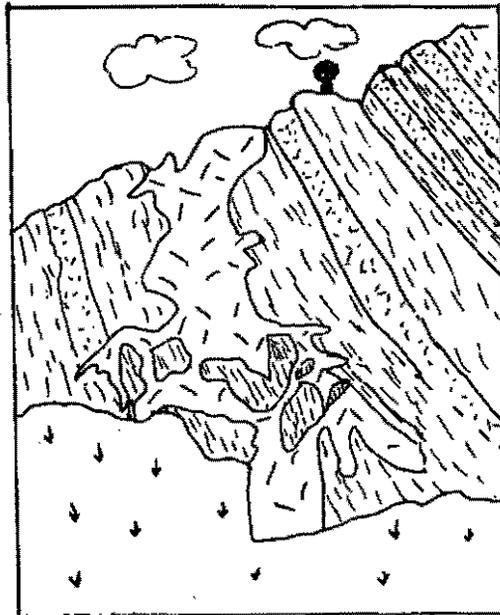


Figure 20 STOP 5. Lit-par-lit structures of SCGI and U.P1.

Continue on I-95S.

1.3 17.6 STOP 6.

Park on right at edge of turnpike. WATCH FOR TRAFFIC. Walk ~ 100m. examining outcrop. Outcrop is predominantly SCGI with diffuse patches of SCGIII. This stop offers another opportunity to examine the assimilation of xenoliths and the biotite schlieren seen at STOP 3. Several very mafic blocks ~ 1m. in length show varying degrees of assimilation. These blocks appear to define a relict stratigraphy which is at an angle to the phase layering in SCGI. Plagioclase + biotite rims develop around the blocks at the contacts to granite. Local concentrations of mafics in SCGI appear to parallel the relict stratigraphy defined by the enclaves, and are likewise at an angle to phase layering. Moving W. along the outcrop (i.e. moving in from the dome margin) the dip of phase layering in SCGI can be seen to increase.

0.0 17.6 Return to vehicles and continue on I-95S.

1.3 18.9 Take exit 56 off I-95S. Turn left at stop sign onto Thimble Islands Road.

Continue on Thimble Islands Road past STOPS 1 and 2 to the junction of Thimble Islands Road and Conn. Rte. 146.

- 1.9 20.8 Junction of Thimble Islands Road and Conn. Rte. 146. Straight across on Thimble Islands Road, following sign for "Town Dock".
- 0.6 1.4 Bear left on Thimble Islands Road, following signs for "Town Dock".
- 0.2 21.6 Bear left on Thimble Islands Road.
- 0.1 21.7 Tennis courts on right.
- 0.1 21.8 Large Victorian mansion on right, Wallace Road on left. Don't take this entrance to Wallace Road.
- 0.1 21.9 Wallace Road on left; Long Point Road on right. Turn left on Wallace Road. Almost immediately there is a small green house with a white picket fence on the left; road forks just beyond this. Left fork goes up hill--do not take this. Take right fork into woods.
- 0.2 22.1 Track bends sharply to right through marsh. Park in turning circle on left. Follow main track around to right hand side of quarry--continue to far wall (east side) of quarry.

STOP 7.

DANGER--DO NOT CLIMB QUARRY WALLS.

Two old quarries near core of dome contain a garnetiferous facies of SCGI intimately associated with the more typical garnet-free SCGI. Strain is low in both types (cf. STOP 3). The garnetiferous facies is irregularly distributed on a scale ~ 10m; it does not show clear contact relations to SCGI. Enclaves of semi-pelitic material here are assigned to the Middle Plainfield; they show no consistent relation to the development of garnets in SCGI. Both garnet-bearing and garnet-free SCGI are cut by small veins of coarser, random-textured material. The veins in garnet-free SCGI appear to be of SCGIII type, and commonly show a well-developed selvedge (as was seen at STOP 3). Garnetiferous SCGI carries veins which are rich in Kspar+garnet; these veins typically lack a selvedge. Garnets in the veins are up to 5x the grain diameter of those in the host (see sketch). There is no systematic pattern to the orientation of the veins (Plate 3).

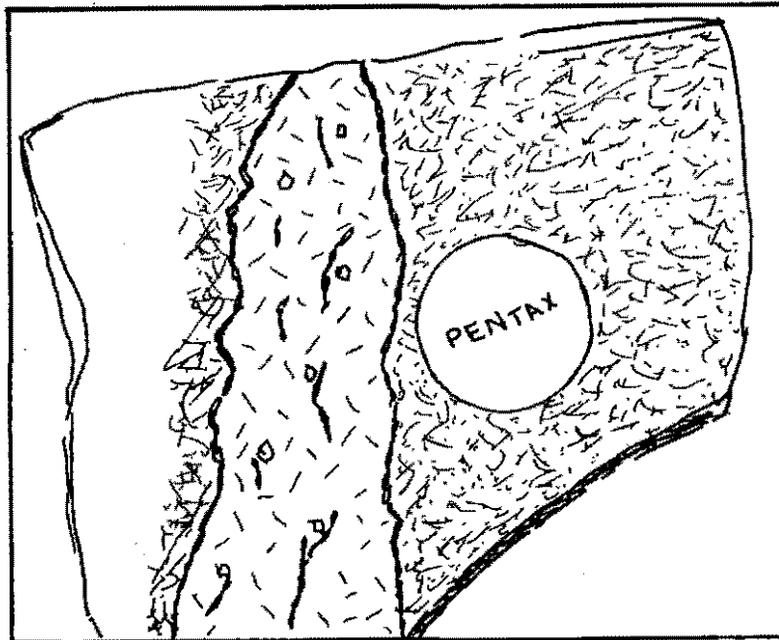


Figure 21 STOP 7. Garnetiferous (?) SCGIII vein in garnetiferous (?) SCGI.

The garnetiferous facies of SCGI has not been found elsewhere in the dome. Samples from these quarries were among those used by Hills and Dasch (1972) in dating the so-called Stony Creek Granite.

Return to vehicles.

- |     |      |   |
|-----|------|---|
| 0.0 | 22.1 | Drive back down track to Wallace Road.  |
| 0.2 | 22.3 | Turn left onto Wallace Road, then almost immediately right onto Thimble Islands Road at stop sign.          |
| 0.5 | 22.8 | Thimble Islands Road Tennis Courts. Park by tennis courts, walk through motel grounds to outcrops on shore. |

#### STOP 8.

(At the time of writing, we are unclear whether we will be able to obtain permission to visit this site.)

Outcrops below Stony Creek Pier show extensively migmatized and folded Lower Plainfield Formation (Plate 4 and sketch below). Biotite-gneisses predominate, often carrying veins rich in garnet. Highly aluminous layers consist of biotite + garnet (up to 25% gt)-- these do not appear to be restitic, but appear to be

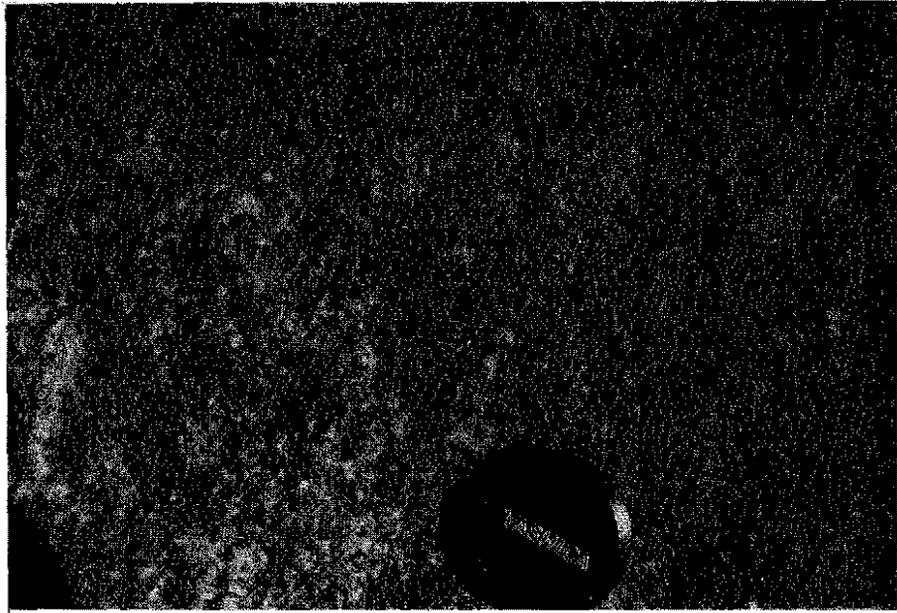


Plate 2 STOP 3. Close-up of SCGIII.



Plate 3 STOP 7. Garnet veins in SCGI.

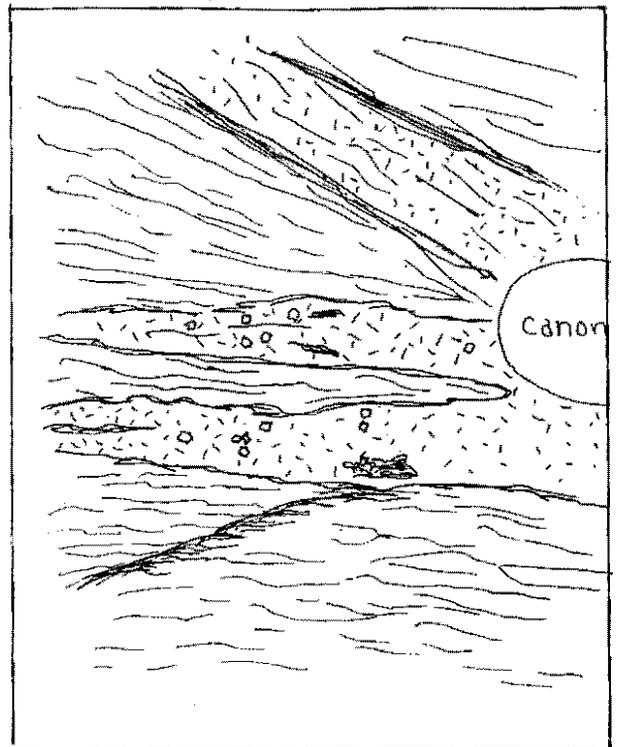


Figure 22 STOP 8. Migmatized L.PI. with garnet bearing leucosomes.

original compositional layering. There are rare amphibole-rich lenses. Garnet-biotite geothermometry suggests  $T \sim 770^{\circ}\text{C}$  here. Migmatite lenses in the Lower Plainfield are warped into open, upright folds; these folds are locally cut by small shears. Many leucosomes are trondhjemitoid, with tiny garnets enclosed in plagioclase crystals. The outcrop is cut by a large vein of massive pegmatite; at first glance this resembles SCGIII pegmatites such as seen at STOP 5, but it seems to be more quartz-rich, Kfeldspar poor. Locally it may contain biotite schlieren, and locally has garnet  $\pm$  muscovite.

This outcrop is near the core of the dome, and is therefore presumed to be low-strain. At STOP 12 we will see very similar lithologies to this, but showing more sign of strain. This variation in strain across the dome, and consequently variable fabric, makes correlation of the granitic units very difficult.

Return to vehicles.

- |     |      |  |
|-----|------|--|
| 0.0 | 22.8 | Turn right at stop sign onto Thimble Islands Road.   |
| 0.7 | 23.5 | Turn right at stop sign onto Conn. Rte. 146. This takes us back past the entrance to STOP 3.   |
| 1.7 | 25.2 | SLOW--DANGEROUS BRIDGE AGAIN! Wave to those field trip participants stranded at first attempt at this bridge.                            |
| 0.3 | 25.5 | Road bends to right with New Quarry Road on left, and, immediately afterward, Old Quarry Road on right. Turn right down Old Quarry Road. |
| 0.6 | 26.1 | Old Quarry Road ends in circle. Park here. Walk between stone gate posts into "Yale University--Peabody Museum Field Station".           |

Walk past house to outcrops on shore. (If doing this subsequent to NEIGC, ask permission at the house.)

LUNCH STOP

STOP 9.

BE CAREFUL ON STEEP ROCKS!

Good exposures of SCGII, showing strong compositional banding defined by stringers of biotite + garnets, also strong shape fabric in quartz and feldspar (Plates 5, 6). Locally this fabric wraps augen of Kfeldspar.

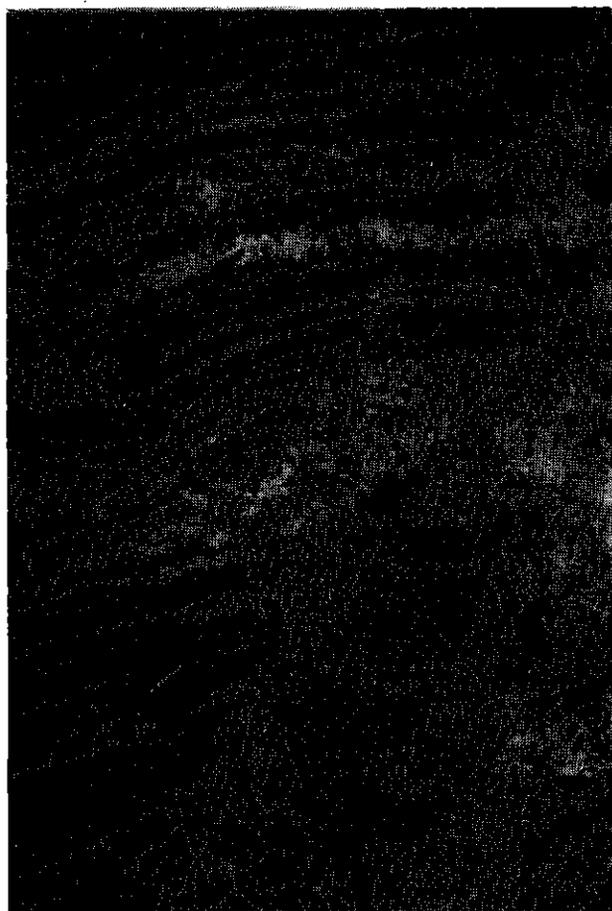


Plate 4 STOP 8. Migmatized L.PL. with selvages. Note flexures and cross-cut pegmatite veins.

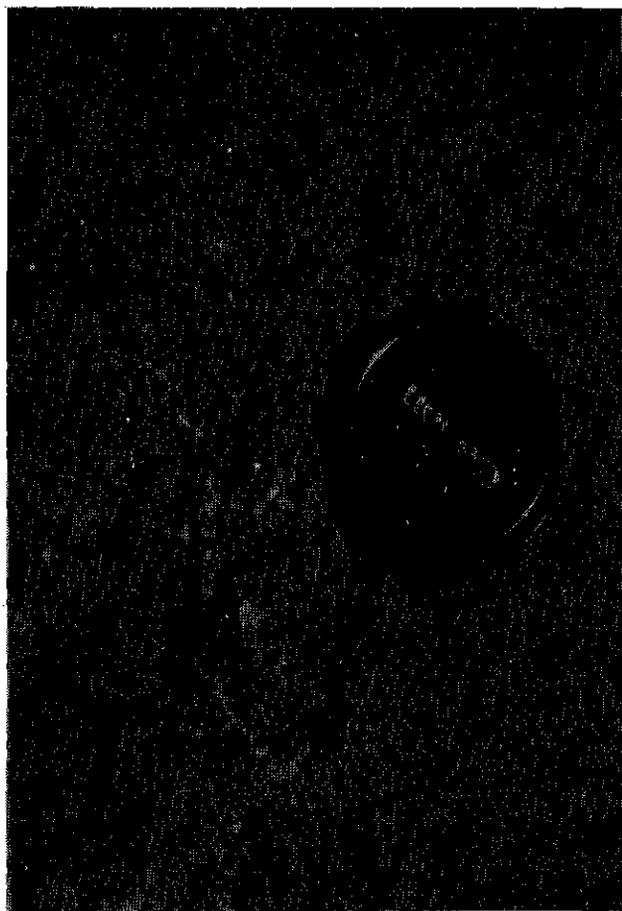


Plate 6 STOP 9. Close up of garnet and biotite stringers in SCGII.



Plate 5 STOP 9. Strongly lineated SCGII. Note flattened aplite.

SCGII is typically associated with Lower Plainfield Formation, but this is rare here. Biotite-rich enclaves of Lower Plainfield are locally rich in garnet; wisps of this material can be traced out into enclosing SCGII, where they form strands of tiny garnets. The biotite + garnet layers may be interpreted as highly strained layers of restite, or as the result of deformation of an original biotite + garnet-bearing granite.

An aplitic unit can be seen within the SCGII, at a low angle to layering; this may be a vein rotated into near-parallelism with layering during deformation. Both the layering in SCGII and the aplitic unit are deformed by shear zones identical to those seen at STOP 3.

- 0.0 26.1 Return to vehicles and retrace Old Quarry Road back towards Conn. Rte. 146.
- 0.5 26.6 Bear left, following "one-way" sign. At stop sign turn right on Conn. Rte. 146.
- 0.4 27.0 Moose Hill Road (leading to STOP 4) on left, Shell Beach Road on right. Turn right down Shell Beach Road.
- 0.2 27.2 Sign on left reads "Private Beach, Leetes Island Members Only". Park at right, by beach. Obtain permission if necessary. Walk to outcrops on right of beach. DO NOT ENTER GARDEN.

#### STOP 10.

Migmatized Lower Plainfield, SCGII and SCGIV. Biotite schist horizons of the Lower Plainfield carry very large plagioclases; are these porphyroblasts growing in situ by a metamorphic/anatectic process, or are they porphyroclasts indicating preferential deformation of originally coarser areas? Note the occurrence of isolated very coarse Kfeldspar elsewhere in the outcrop. Garnet-biotite temperatures here are ~ 730°C. As seen at STOP 8, the Lower Plainfield is deformed by early open folds and late shears.

SCGII is apparently interlayered with the migmatized Lower Plainfield, and shows the same shear deformation. This is presumably contemporaneous with that at STOPS 3 and 9. Biotite schlieren are common within masses of SCGII. SCGIV is developed as diffuse patches within shear zones in SCGII, and can be distinguished by its lack of fabric (it cross-cuts layering in SCGII) and by the presence of abundant garnets. These garnets are coarser than those in SCGII.

The outcrop is cut by a huge pegmatite vein (crystals up to 6cm.) bordered by huge biotite crystals. The relationship of this pegmatite to that at STOP 8, and in turn to pegmatitic SCGIII at STOP 5, is unknown. Such pegmatites are very common in this part of the dome, and show similarities to the pegmatites described at "Narragansett Pier Granite of Blackhall type" further east in Connecticut (Lundgren, 1967).

Return to vehicles.

0.0 27.2 Continue along Shell Beach Road.

0.4 27.6 Shell Beach Road bears left; do not take right fork (Point Road--Dead End). Immediately afterwards turn right into Rockledge Circle (one-way) and park.

Walk back down Rockledge Circle and turn right onto Beach Road. After 0.1 mile road forks; Boulder Road to left and Great Harbour Drive to right. Bear right; first house on right (no 72) sits low by water. Cross garden of no. 72 (obtain permission) and descend steps to water's edge.

#### STOP 11.

This again shows SCGII and Lower Plainfield, as at the previous stop, but here the contacts between the two are marked by selvages. Enclaves of very coarsely crystalline bi + plag + qz can be found in SCGII. Are these xenoliths of Lower Plainfield caught up in intrusive SCGII and subsequently deformed, or are they restites produced complementary to SCGII during migmatization of Plainfield?

This outcrop shows a very strong fabric in SCGII. Note pods of musc + gt-bearing aplite elongate parallel to foliation--this material is similar to the SCGIV developed in shear zones at STOP 10 and which will be seen again at STOP 12.

Walk round outcrop to front of house. This shows a possible contact between SCGI and SCGII. Granite here has the strong lineations typical of SCGII but lacks the characteristic garnet. Veins here are of the SCGIII type normally found in association with SCGI (as at STOPS 3 and 4). Is this rock transitional between SCGI and SCGII, or can a contact between two distinct types be traced? Could it simply be a high-strain facies of SCGI?

Look at outcrop by seawall. This shows the concordant contacts between Lower Plainfield and SCGII. Migmatized Plainfield has abundant coarse garnet in leucosomes.

Return to vehicles.

- 0.0 27.6 Drive round Rockledge Circle back to Shell Beach Road. Turn left on Shell Beach Road and drive back past stop 10 to Conn. Rte. 146.
- 0.6 28.2 Junction of Shell Beach Road and Conn. Rte. 146. Turn right onto Conn. Rte. 146.
- 1.3 29.5 Conn. Rte. 146 swings sharp left under railroad bridge.
- 0.4 29.9 Turn right on Mulberry Point Road.
- 0.4 30.3 Junction of Mulberry Point Road and Chaffinch Island Road; continue straight on Mulberry Point Road.
- 0.7 31.0 Follow Mulberry Point Road round to left.
- 0.1 31.1 Park on right.

Turn around, walk 0.1 mile back along Mulberry Point Road to junction with Daniel Avenue. Turn left onto Daniel Avenue.

Walk 0.4 mile up Daniel Avenue; follow the "one-way" sign. Turn right at stop sign onto Indian Cove Road. Walk 0.1 mile on Indian Cove Road, turn left onto Spencer Avenue.

At junction of Spencer Avenue and Prout St., go left down grass slope towards sea, keeping the red house on your right. This is a right-of-way. Turn right at shore and walk ~ 0.5 mile along outcrops.

TAKE CARE ON STEEP SLIPPERY ROCKS.

#### STOP 12.

The first set of outcrops shows a sharp contact between SCGII and amphibolites of the Lower Plainfield Formation. Folds and shears are well-developed (Plate 7). Garnetiferous leucosomes (SCGIV) occur in shear zones (Plate 8). None of these leucosomes can be traced for more than 3m. across the outcrop; their contacts to SCGII are generally sharp. The leucosomes trend sub-parallel to shears. Note the very strong lineation at this outcrop, marked by extreme elongation of quartz and feldspars.



Plate 7 STOP 12. SCGII showing strong augen bent into shear zone at left.



Plate 9 STOP 12. Migmatized L.PL. with pegmatite (?)=SCGIII.



Plate 8 STOP 12. SCGIV concentrated in shear zones in SCGII.

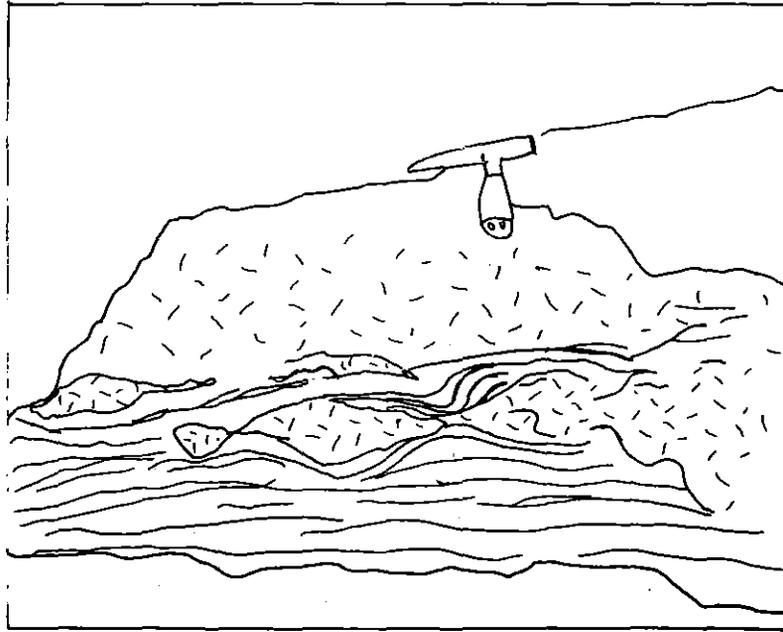


Figure 23 STOP 12. Drag folds around pegmatic boudins in L.PL.

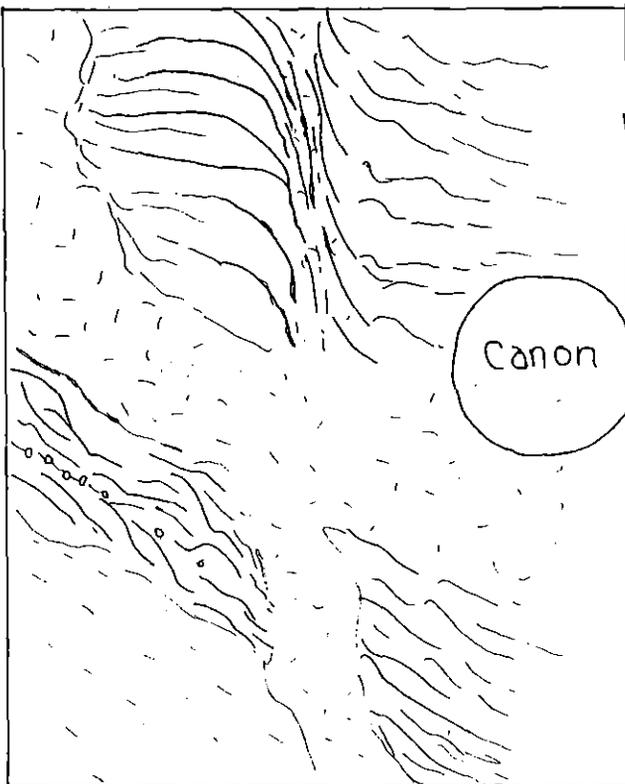


Figure 24 STOP 12. SCGIV concentrated in shear zones in SCGII.

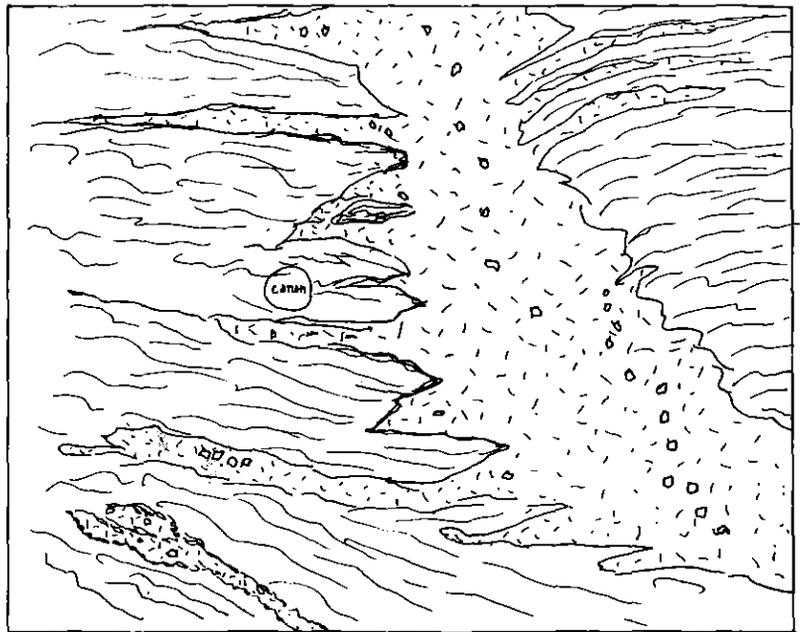


Figure 25 STOP 12. SCGIV forms diffuse patches within SCGII.

After ~ 0.25 mile we reach a very large shear zone just before the house on the point. A pegmatitic granite of the type seen at STOP 10 (possible NPG) occurs in this shear zone; it makes a sharp contact to an aplite which in turn cuts SCGII. A boudinaged vein in the Lower Plainfield here (see sketch) shows drag folds suggestive of top-to-S.E. motion.

Round the point, just below the cairn. Diffuse patches of SCGIV in SCGII have abundant coarse garnet.

50 meters further on, we return to the amphibolite seen at the first outcrop of this stop, veined by coarse Kspar + bi pegmatite of possible NPG-Blackhall type (see Plate 9).

Return to vehicles.

THIS IS END OF TRIP.

We will return to Branhaven Plaza to collect vehicles.

- |     |      |   |
|-----|------|---|
| 0.0 | 31.1 | Drive back down Mulberry Point Road.  |
| 0.1 | 31.2 | Junction of Daniel Avenue and Mulberry Point Road.<br>Turn right onto Mulberry Point Road.  |
| 1.1 | 32.3 | Junction of Mulberry Point Road and Conn. Rte. 146.<br>Turn left onto Conn. Rte. 146.   |
| 0.7 | 32.9 | SLOW-DANGEROUS BRIDGE   |
| 1.7 | 34.6 | Junction of Conn. Rte. 146 and Thimble Islands Road.<br>Turn right on Thimble Islands Road.   |
| 1.5 | 36.1 | Turn left onto I-95S (signed for "New Haven and West").   |
| 1.0 | 37.1 | Take exit 55 off I-95S. On reaching stop sign, turn right at U.S. Rte. 1 west. Drive 2.8 mile along Rte. U.S. 1 west; turn right at lights into Branhaven Plaza. There is no exit 53 in this direction. |

FINAL NOTE: Excellent polished samples of granite from STOP 3 can be found facing the General Post Office in downtown New Haven.

Acknowledgements:

Our thanks to Eric Miller who bravely battled the January snows to aid in the preparation of this roadlog, and whose efforts in sweeping the snow from outcrops will be long remembered. My thanks also to Bob Tracy,

who introduced me to these rocks as part of a more general introduction to the U.S., and who has since offered boundless encouragement and advice. Preparation of this roadlog was partially supported by NSF Grant #EAR83-19673 to Tracy. Thanks also to Vince La Piana for help in sample collection, Tony Creamer for the XRF analyses and to Jeanne Martin for rapid typing of this M.S. Probe data were collected at the Geophysical Laboratory, Carnegie Institute, Washington, D. C.; my thanks for the use of that facility, and to David George for his patient instruction in the use of the probe.

I also wish to thank the wonderful people of Stony Creek, who welcomed me to their gardens, homes and hearts. In particular, I wish to thank Ann Shimuzi, Sonny Wallace, William van Wie, Alison Page, the Leete Family, Mr. & Mrs. Hall, Shirley Belden, Mr. Carter and John Barnes.

## BEDROCK GEOLOGY OF THE DEEP RIVER AREA, CONNECTICUT

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### Introduction

The geology of the Deep River area is critical to understanding regional structural relationships in southeastern New England because at least four major terranes: the Avalon, Nashoba, Merrimack and Bronson Hill (see Zartman and Naylor, 1984) converge here in a 1 km wide zone between Chester and Centerbrook, just west of the Connecticut River (Fig. 1). Mapping in the Deep River area by Lundgren (1963, 1964) has led him (Lundgren, 1962) and Dixon and Lundgren (1968) to propose that a major recumbent syncline of regional significance joins what would now be called the Bronson Hill with the Avalon and the Merrimack with the Nashoba terranes. Because of the important constraints that the geology of the Deep River area places on these regional interpretations, the narrow belt of rocks between the Bronson Hill anticlinorium and the Avalonian terrane, known informally as the Appendix, has been restudied. Probably the most important structural interpretation proposed by Lundgren (1962) in the Deep River area is the Chester syncline, which he believed to be overturned to the west. The evidence for this interpretation stemmed from (1) the correlation of the pelitic units, Putnam gneiss and Brimfield formation (Lundgren, 1963, p. 28) and (2) the correlation of three belts of plagioclase gneiss in the Deep River quadrangle (i.e., Turkey Hill, Hadlyme and Cedar Lake) as a single stratigraphic unit (Lundgren, 1962, p. 14). These correlations suggested that Hebron Formation (Table 1) occupies the core of the Chester syncline, with Putnam gneiss and Hadlyme plagioclase gneiss forming the overturned east limb, and Brimfield formation and Turkey Hill plagioclase gneiss forming the normal west limb. Once established in the Deep River area, the concept of an overturned syncline was extended farther to the north and east, and has become the dominant structure in interpreting structural and stratigraphic relationships in eastern Connecticut (Dixon and Lundgren, 1968).

Since 1959 when Lundgren completed field mapping in the area much new work has been undertaken that bears on structural and stratigraphic interpretations. New geological work includes the subdivision of the Putnam gneiss into nine stratigraphic units (Dixon, 1964), the completion of quadrangle mapping in most of eastern Connecticut (Rodgers, 1985), the recognition of major lithotectonic terranes in New England (e.g., Zartman and Naylor, 1984) and detailed geologic mapping in the Deep River area (Wintsch, 1979; Wintsch and Kodidek, 1981; Wintsch, unpub.). Copies of the detailed bedrock geologic map of the Deep River area (Wintsch, in prep.) will be available through the Connecticut Geological and Natural History Survey. New geochemical work includes a large number of major-element chemical

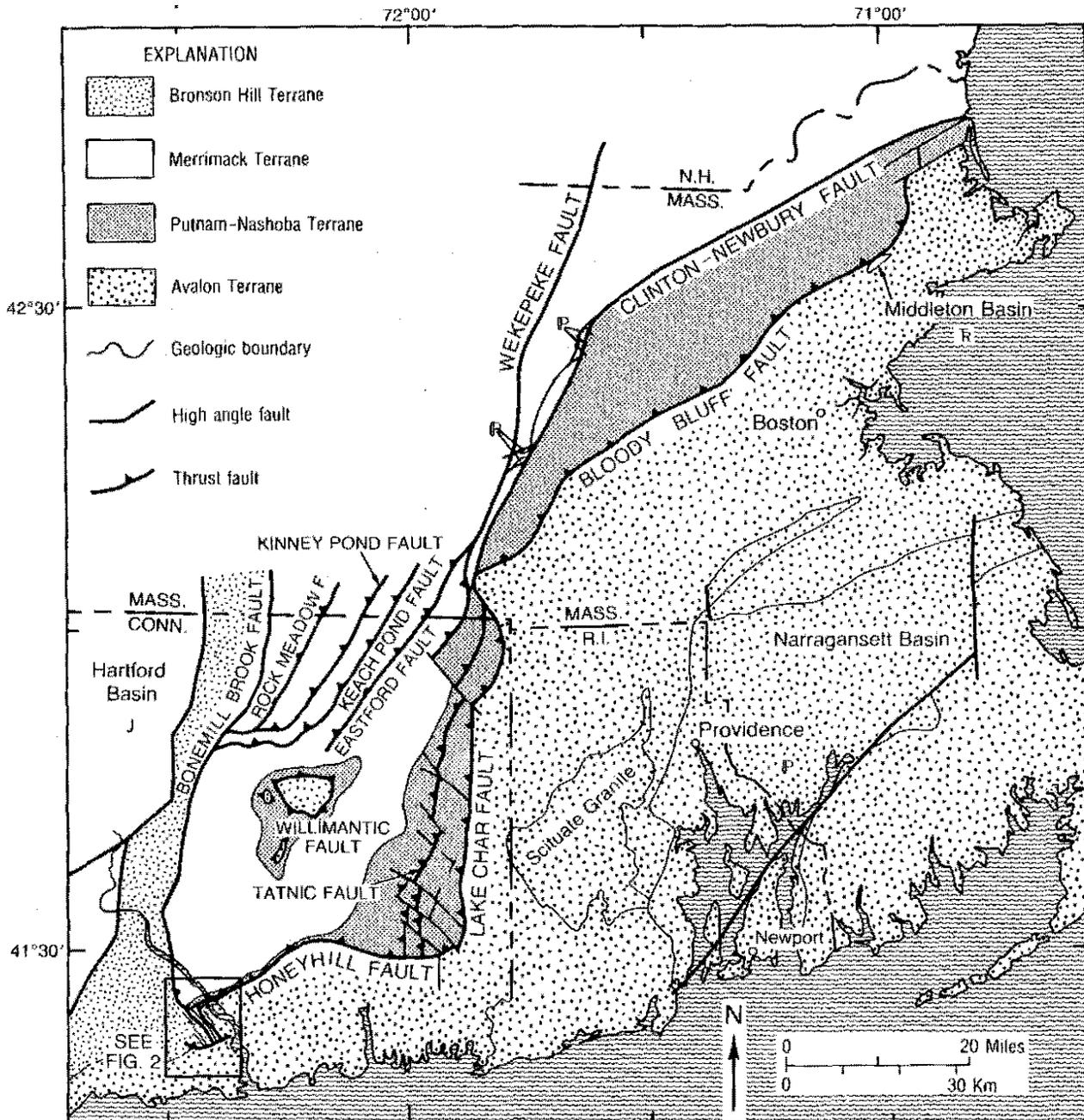


Fig. 1. Tectonic map of Southeastern New England, showing the distribution of the four major terranes as outlined by Zartman and Naylor (1984). All four terranes meet in the Deep River area of south-central Connecticut (Fig. 2) and the geology of that area is thus critical to understanding the relationships among these terranes.

analyses made from samples of the abundant plagioclase gneisses in the Bronson Hill and Avalon terranes (Wintsch and Grant, 1980; Webster and Wintsch, 1984; Leo et al., 1984) and the large number of Pb-U age determinations measured from zircons collected from all terranes (Leo et al., 1984; Zartman and Naylor, 1984; Hermes and Zartman, 1985; Zartman and Leo, 1985). These new data have led to subsequent interpretations that include the correlation of rocks south and east of the Honey Hill fault with the Avalon terrane (Goldsmith, 1980; Rodgers, 1985), and the correlation of the Putnam gneiss with the Late Proterozoic Nashoba terrane (Zartman and Naylor, 1984).

This trip is based on these new field and laboratory data that also form the basis for a reinterpretation of the Chester syncline. The most important result of this study is the identification in the Deep River area of all the terranes recognized on a regional scale (e.g., Zartman and Naylor, 1984), and the identification of major ductile faults at terrane boundaries. In the southeastern portion of the mapped area the plagioclase gneisses of Hadlyme (Zwh) and Essex (Zwe) intruded by a granite (Zsph) and locally overlain by an amphibolite-rich plagioclase gneiss (Zwc) constitute the northwestern portion of the late Proterozoic (Stop 2a) Avalon terrane (see Fig. 2). The northern and western boundaries of the terrane are defined by ductile faults (Honey Hill, Stop 7; Falls River, Stop 2; and unnamed, e.g. Stop 4). In the western portion of the mapped area lie the Ordovician Monson Gneiss and Middletown Formation of the Bronson Hill terrane. The eastern and southern boundaries of this terrane are defined by the ductile Bonemill Brook and Falls River faults (Fig. 2x). Pinched between these Avalon and Bronson Hill terranes in the Appendix lie fault slivers of Hebron Formation of the Merrimack terrane. Rocks of the Putnam-Nashoba terrane (Ztay, Ztal) are not found between Chester and Centerbrook and must be cut out by the fault separating O?h from Zwc (Stop 4).

This interpretation of several discrete terranes in the Deep River area is not consistent with the interpretation of the Chester syncline, which requires the correlation of map units Zwe and Zwh with Qm (See Table 1 and Lundgren, 1962). The differences in age (Stop 2a) and in chemical composition (Wintsch and Grant, 1980; Webster and Wintsch, 1984) make such a correlation doubtful, and supports the interpretation that the two structural blocks do constitute discrete terranes.

### Stratigraphy

The correlation and age assignment of stratified rocks in the Deep River area made in this study differ in some important ways from those made by Lundgren (1963; 1964), as outlined in Table 1. The most significant difference is that in this study units are grouped by terranes and no correlations across terrane boundaries are made. This differs from the treatment of Lundgren who correlated the Brimfield Formation with the Putnam gneiss and assigned three geographically distinct belts of plagioclase gneiss to the single unit: Monson Gneiss (Table 1). In the present study rocks of the Putnam gneiss of Lundgren (1963; 1964) are subdivided into the Yantic and lower members of the

Table 1. Comparison of mapped units and of correlation of units by Lundgren (1963, 1964) and Wintsch (this study)

Lundgren (1963;1964)		Wintsch (this study) <sup>B</sup>			
SINGLE TERRANE <sup>A</sup>	BRONSON HILL TERRANE	MERRIMACK TERRANE	NASHOBA TERRANE	AVALON TERRANE	
Hebron Formation <sup>E</sup> (h)	Granite <sup>I</sup> (Pp)			Granite (Pp) <sup>I</sup>	
Canterbury Gneiss (c)		Canterbury Gneiss (Dc)			
Brimfield <sup>C</sup> Putnam <sup>D</sup> Formation Gneiss (bm) Pelitic (pm) Calc- Silicate <sup>E</sup> (pc)	Granite Gneiss (O?gg)	Hebron Formation (O?h)		Sterling gneiss <sup>J</sup> (Zsph)	
Middletown Formation <sup>F</sup> (mig)	Middletown <sup>F</sup> Formation (Om)		Tatnic Hill Fm <sup>D</sup>	Waterford Group <sup>H</sup>	
Monson Gneiss (m)	Monson Gneiss <sup>G</sup> (Omo)		Yantic Member (Ztay)	Gneiss of Chester <sup>D</sup> (Zwc)	
Turkey Hill belt <sup>G</sup> (m <sub>TH</sub> )	Mylonitic Gneiss (Omom)		Lower Member (Ztal)	Gneiss of Hadlyme <sup>H</sup> (Zwh)	
Hadlyme belt <sup>H</sup> (m <sub>H</sub> )				Gneiss of Essex <sup>H</sup> (Zwe)	
Cedar Lake belt <sup>G</sup> (m <sub>CL</sub> )				Mylonitic gneiss <sup>H</sup> (Zwem)	
New London Gneiss <sup>I</sup> (n)				Quartz-Biotite Gneiss of Centerbrook <sup>E</sup> (Zc)	
Sterling Gneiss <sup>J</sup> (sgb)				Amphibolite Gneiss <sup>F</sup> of Plains Road (Zp)	

44-4

- A All rocks are considered by Lundgren (1962; 1963; 1964) to belong to a single terrane.
- B Rocks are divided into four terranes and no correlations across terrane boundaries are made.
- C Brimfield Formation not mapped along eastern margin of Middletown Formation. Local rare sillimanite-bearing schist mapped here as mylonitic Hebron Formation in the ductile Bonemill Brook fault (see text).
- D Putnam gneiss of Lundgren (1963) subdivided into Yantic and lower members of Tatnic Hill Formation (Dixon, 1964) and gneiss of Chester (Zwc).
- E Calc-silicate unit in Putnam gneiss mapped here as part of Hebron Formation except south of Falls River fault where Lundgren's (1964) quartz-biotite-plagioclase gneiss (pgg) variety of Putnam gneiss and his Hebron Formation are mapped together as gneiss of Centerbrook (Zc).
- F Middletown Formation of Lundgren (1963, 1964) unchanged north of Falls River fault, but mapped as Amphibolite gneiss of Plains Road south of Falls River fault.
- G Turkey Hill and Cedar Lake belts mapped together as Monson gneiss.
- H Hadlyme belt of Lundgren (1962, 1963) equated with the Rope Ferry Gneiss of Goldsmith (1980) and subdivided into gneisses of Chester, Hadlyme, Essex and mylonitic gneiss.
- I New London gneiss of Lundgren (1963, 1964) not mapped, but may be equivalent in part to Granite (Pp).
- J Sterling Gneiss is unchanged, but mapped area is expanded.

Tatnic Hill Formation, and the amphibolite-rich plagioclase gneiss of Chester. The Tatnic Hill Formation is correlated with the Nashoba terrane (e.g. Zartman and Naylor, 1984) which is late Proterozoic in age (Olszewski, 1980). The gneiss of Chester is considered part of the Avalon terrane because of its gradational lower contact with the plagioclase gneiss of Hadlyme (Table 1).

The Monson Gneiss of Lundgren is subdivided here on the basis of mineralogy, the presence of amphibolite xenoliths, and on major element chemical trends, and age assignments are made in view of new Pb-U dates by Zartman and Naylor (1984) Leo, et al. (1984), Zartman and Leo (1985) and Aleinikoff (pers. comm., Stop 2a). In the present study, the Hadlyme belt with relatively high concentrations of  $TiO_2$ ,  $Al_2O_3$ , MgO and CaO is correlated with the Rope Ferry Gneiss of the Waterford Group of Goldsmith (1980) of the Avalon terrane. The Cedar Lake and Turkey Hill belts, with relatively high  $SiO_2$ ,  $K_2O$  and  $Na_2O$  concentrations, are undivided in the Bronson Hill terrane, although there is some geochemical evidence (Webster and Wintsch, 1984) that they may be different metavolcanic units.

### Rock Units

Three important types of metamorphic rocks are recognized in the Deep River area: meta-intrusive orthogneisses, stratified meta-sedimentary and metavolcanic schists and gneisses, and blastomylonitic schists. Many gneisses in the Deep River area are igneous in bulk composition, but very high grade metamorphism, and locally intense ductile deformation have given these rocks a very strong foliation that often obliterates primary structures and contact relationships. Thus some rocks mapped here as metavolcanic (Om, Omo, Zwc, Zwh) could be orthogneisses, but in the absence of evidence to the contrary, are indicated as metavolcanics. Metasedimentary rocks include both pelitic and calcareous schists and gneisses (Omom, Zwem). Rocks whose textures, outcrop scale structure, and perhaps also mineralogy have been strongly influenced by deformation and metasomatism are indicated as tectonic or blastomylonitic schists and gneisses. Some of the units in the mapped area (Fig. 2) have not been described before, and informal names are adopted pending formal definition of these units. Rock units are described briefly below without regard to terrane. For correlation of mapped units, see Table 1.

#### Stratified Rocks

Om Middletown Formation. A dark gray to black weathering hornblende gneiss and amphibolite unit (Lundgren, 1963), often finely layered on a grain scale. In fault contact with O?h.

Omo Monson Gneiss (plagioclase gneisses of Turkey Hill and Cedar Lake, Lundgren, 1963). A light to medium gray weathering quartz-plagioclase-biotite anthophyllite gneiss (Stop 10) of high  $SiO_2$  (> 70 wt. %) composition (Stop 5, 9, 10) of Ordovician age (Zartman and Naylor, 1984). Interlayered contact with overlying Omm.

O?h Hebron Formation (Lundgren, 1963). Olive green to gray weathering plagioclase-quartz-biotite schist commonly inter-layered on a cm scale with diopside and/or epidote bearing schists (Stops 3, 4). This unit includes most of Lundgren's (1963) Hebron Formation, calc-silicate unit in Putnam gneiss. Rare, thin sillimanite bearing biotite schists mapped by Lundgren (1963) as Brimfield Formation are also included. The interlayered lower contact with Ztay suggests that this unit may also be late proterozoic in age, but Rodgers (1985) indicates it as SOh.

Ztay Yantic Member, Tatnic Hill Formation (part of Putnam gneiss of Lundgren, 1963). Pale to medium gray weathering quartz-plagioclase-muscovite-biotite schist characteristically mixed at a 5 cm scale with quartz-plagioclase microcline granofels. Rocks usually contain conspicuous 1-2 cm plagioclase augen and up to 20% retrograde muscovite and are probably late Proterozoic in age by correlation with Ztal. Gradational upper contact with Hebron Formation.

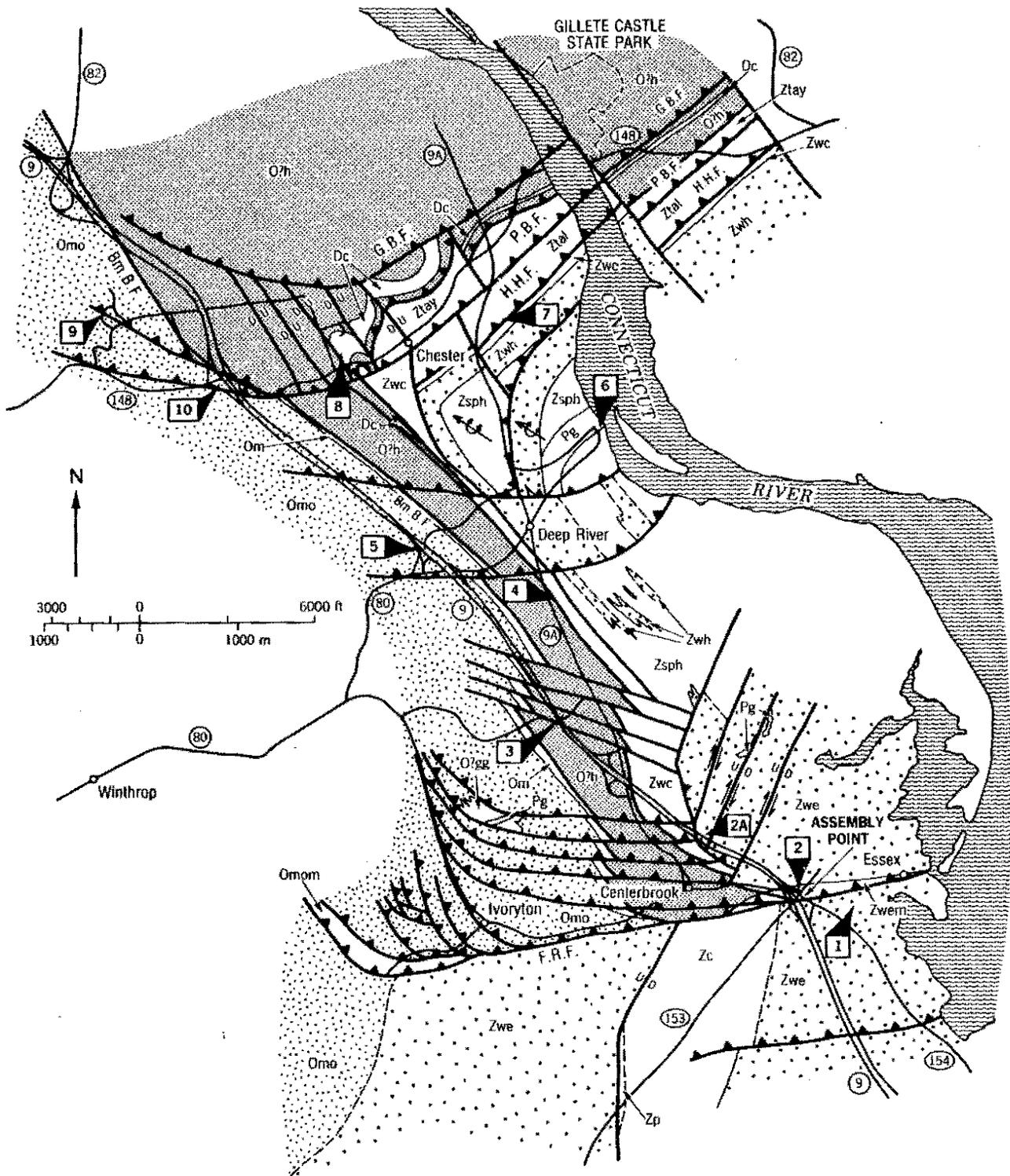
Ztal Lower member, Tatnic Hill Formation (part of Putnam gneiss of Lundgren, 1963). Brown to buff weathering, garnet and/or sillimanite bearing pelitic schists (Stop 7) of probable late Proterozoic age by correlation with the Nashoba block (Olszewski, 1980; Zartman and Naylor, 1984). Upper and lower contacts are faults.

Zwc Amphibolite-rich plagioclase gneiss of Chester (part of Putnam gneiss of Lundgren, 1963, 1964). A mixed unit, approximately equal amounts of plagioclase gneiss of similar composition to Zwh and of amphibolite of olivine tholeiitic to alkali basaltic composition (Stop 7). The lower contact is interlayered with Zwh, and the upper contact is defined by the Honey Hill fault. The interlayering with Zwh suggests a mixed metavolcanic protolith.

Zwh Plagioclase gneiss of Hadlyme (Hadlyme belt of Monson Gneiss of Lundgren, 1963). Light gray weathering plagioclase-quartz-biotite-hornblende-magnetite gneiss of low SiO<sub>2</sub> (<70%) dacitic composition, correlated with the Late Proterozoic Waterford Group of Goldsmith (1980).

Zc Quartz-biotite gneiss of Centerbrook mapped as both quartz-biotite-plagioclase gneiss variety of Putnam Gneiss (pgg) and as Hebron Formation by Lundgren (1964). Gray to buff weathering quartz-plagioclase-biotite granofels in 5 cm thick layers. Lacks cm scale layering and calc-silicate beds typical of Hebron north of Centerbrook. No contact relationships are exposed, but late Proterozoic age assumed by association with gneisses of the Waterford Group.

Zh A dark gray to black weathering hornblende gneiss unit occurring between Zc and Zwc south of Centerbrook. Contact relationships not exposed, but a late Proterozoic age is assumed by association with gneisses of the Waterford Group.



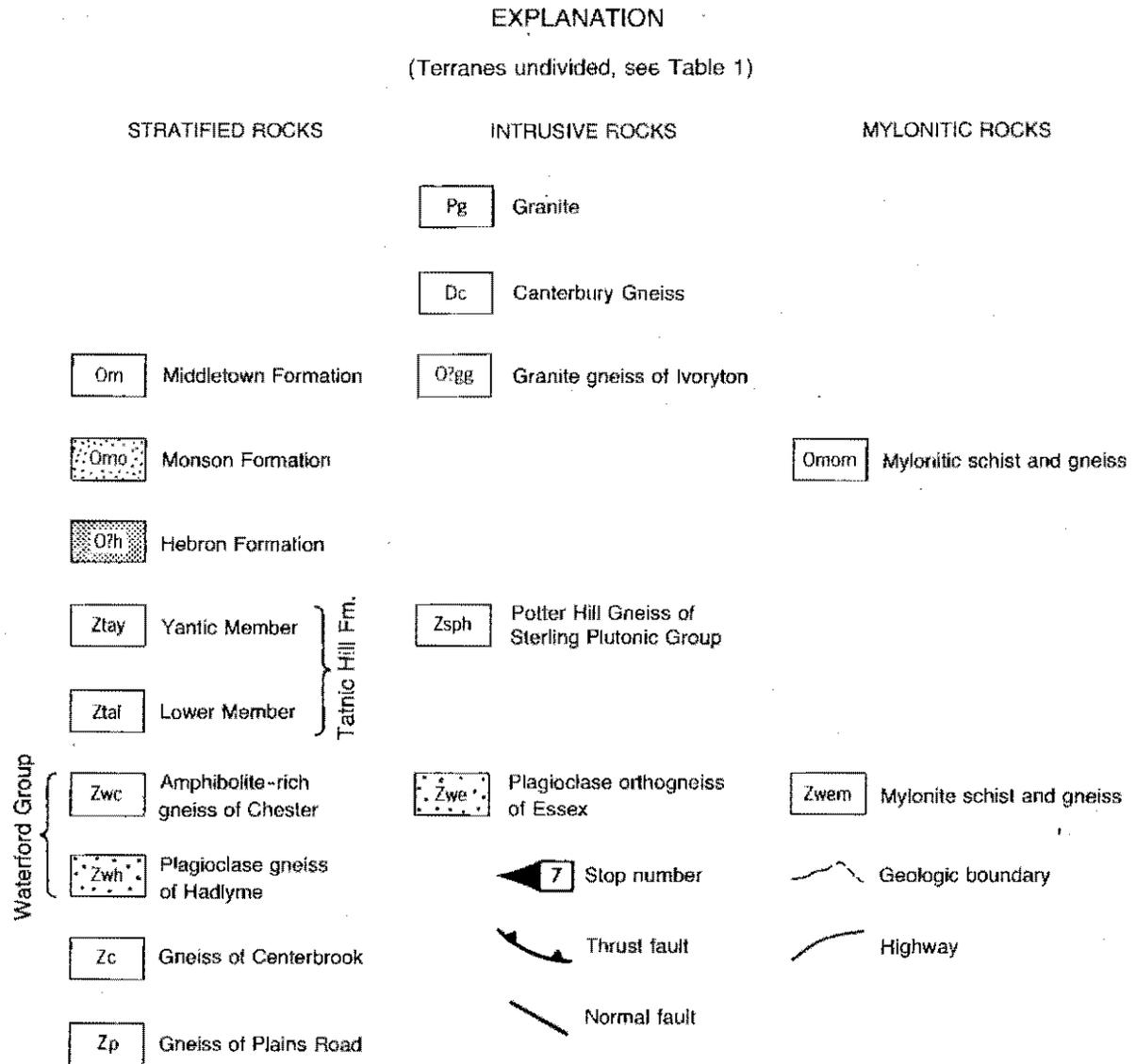


Fig. 2. Preliminary geologic map of the Deep River area, Connecticut, based on detailed mapping (scale 1:12,000, Wintsch, 1979; Wintsch and Kodidek, 1981; Wintsch, unpub.). Geology in Gillette Castle area is based on drill core data. Correlations of units based, in part, on chemical and isotopic data, are shown in Table 1. Abbreviations = G.B.F., Great Brook fault; P.B.F., Pattaconk Brook fault; H.H.F., Honey Hill fault; Bm.B.F., Bonemill Brook fault; F.R.F., Falls River fault.

Intrusive Rocks (All terranes, see Table 1 for correlation of units)

Pp Pegmatite. Pegmatite dikes are ubiquitous in all rocks of the Deep River area. Pre- to syntectonic foliated dikes tend to be rich in plagioclase and quartz, whereas post tectonic undeformed dikes tend to be syenitic to monzonitic in composition (Stop 2a). Larger pegmatites are very common in ductile fault zones and may be internally sheared with the foliation parallel to the surface of the fault in which they intrude (Stop 9). Not mappable at the scale of Fig. 2.

Pg Granite at Deep River. A tan to buff weathering unfoliated even textured quartz-K-feldspar-plagioclase-magnetite-biotite granite. Locally contains conspicuous red spots around magnetite crystals. Unit includes a porphyritic phase with 1 cm euhedral K-feldspar phenocrysts. Thin dikes of this granite (1-4 m thick) cut Zwe on Book Hill. Zircons from the buckled dike exposed at STOP 2A yield an age of about 270 Ma (J. Aleinikoff, pers. comm.), indicating a late Paleozoic age for the late deformation in the Essex area which deforms both the dike and the host gneiss. The unit may be equivalent, in part, to New London gneiss of Lundgren (1963).

Dc Canterbury Gneiss. A grey to buff weathering quartz-K-feldspar-plagioclase-biotite moderately to well foliated gneiss as described by Lundgren (1963). Evidence that the gneiss is intrusive is not found in the mapped area, but is based on its irregular outcrop pattern across the state (see Rodgers, 1985), and its younger age (395 Ma, Zartman and Naylor, 1984) than the age of metamorphism of the Yantic Member of the Tatic Hill Formation ( $\sim$ 440 Ma) as established by Rb-Sr whole rock analysis by Pignolet et al. (1980) in the vicinity of Yantic, Conn. The well developed foliation does show, however, that much deformation is younger than 395 Ma.

O?gg Granite Gneiss of Ivoryton included with Monson gneiss of Lundgren (1964). A small body of quartz-K-feldspar-plagioclase-biotite gneiss mapped only in Omo northeast and east of Ivoryton. An age younger than Ordovician is possible.

Zsph Potter Hill gneiss of the Sterling Plutonic Group of Rodgers (1985). Buff to cream weathering, medium to coarse grained quartz-K-feldspar-plagioclase gneiss. No evidence was found that the gneiss is an orthogneiss, but locally cross-cutting relationships farther east (Goldsmith, 1966) suggest that it is.

Zwe Orthogneiss of Essex. Light to dark gray weathering equigranular quartz-plagioclase-biotite-hornblende-magnetite-sphene gneiss. Inclusions of amphibolite are present in most outcrops (e.g. Stop 2A), and are locally abundant (Stop 1). Their shape is highly dependant on strain, and may have aspect ratios from 2.5:1 to 100:1. Inclusions of layered calc-silicate granofels occur with amphibolite on Rt 153, 4 km south of Centerbrook suggesting that all inclusions are xenoliths, and that this gneiss is an orthogneiss. Zircons from Stop 2A have been dated as approximately 620 Ma by J. Aleinikoff, which establishes a late Precambrian age for this gneiss.

### Mylonitic Rocks

Omom Blastomylonitic schist and gneiss mapped only along the Falls River fault west of Ivoryton and included by Lundgren (1964) with the Monson gneiss. Mylonitic schists dominate the unit, and the occurrence of a small (30 x 30 m) ultramafic pod and a single outcrop of calc-silicate gneiss suggest complicated tectonic mixing in the unit as well.

Zwem Blastomylonitic Schist mapped as part of the Putnam gneiss by Lundgren (1964). Mylonitic schist and augen gneiss mapped only along the Falls River Fault in Essex. This rock grades north into Zwe and chemical analyses of the gneiss layers between the schists are similar to less deformed gneiss, confirming that the gneiss could be derived from Zwe by local metasomatism.

### Structural Geology

The most important structure identified in this study is the 'S' shaped ductile fault zone separating the Avalon terrane from the terranes to the north and west (most obvious on Fig. 1). This fault zone has been repeatedly faulted, which leads to the complex distribution of units shown on Fig. 2. The upper loop of the 'S' structure between Chester and Deep River is defined by an anticline overturned to the west which folds the gneiss of Hadlyme (Zwh). The same anticlinal structure is defined by the Hebron Formation, although the outcrop pattern is complicated by later complex folding and faulting in Chester. The large scale structure is thus a homoclinal sequence folded around the Selden Neck fold of Dixon and Lundgren (1968), an interpretation also considered by Lundgren (1962), but rejected because of the correlations of Monson gneiss with the gneiss of Hadlyme.

The lower loop of the 'S' shaped structure is defined in part by the Falls River fault. The fault is not well exposed except at its eastern and western ends where mylonitic and blastomylonitic schists and gneisses are locally well exposed. This fault juxtaposes both the Bronson Hill and Merrimack terranes to the north with the Avalon terrane to the south. However, the fault also cuts the orthogneiss of Essex, as well as all units and structures in the Appendix, and thus the fabrics in this zone probably reflect Permian reactivation of this major tectonic boundary originally established in pre-Permian (Variscan?) times.

One of the most striking features of the geology of the Deep River area is the complexity of faulting. Almost all lithologic boundaries in the Appendix are occupied by ductile faults, and many faults of small displacement also cut lithologic boundaries at high angles (Fig. 2). Deformation occurred primarily under amphibolite or granulite facies metamorphic conditions, and fault zones are commonly characterized by strong metamorphic fabric (almost every stop this trip) and locally also by the syntectonic intrusion of pegmatites or granites.

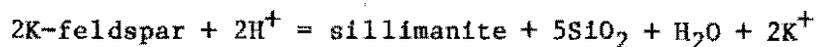
The relative timing of this faulting is revealed by crosscutting

relationship. The Honey Hill fault, the oldest fault in the area, is cut by the fault between Zwh and Zwc (Fig. 2). This fault and all faults between Zwh and Omo including the fault tentatively correlated with the Bonemill Brook fault of Pease (1982), are cut by the Pattaconk Brook and Falls River faults. Both of these are fully ductile and contain syntectonic pegmatite intrusions (eg. Stops 2 and 9) but deformation along the Falls River fault occurred under upper most amphibolite facies conditions, while that along the Pattaconk Brook fault was lower amphibolite facies. Because the faults appear to have operated at the same time, this difference in metamorphic grade probably reflects the difference in crustal levels during their operation. This is consistent with the higher grade of the rocks south of Essex (Lundgren, 1966), and suggests that their greater uplift post dated activity on the Falls River fault. The other faults between Chester and Centerbrook that strike ~N60W and produce dextral displacement do not contain pegmatites, show very little ductile drag of earlier fabric (Stop 3) and sometimes contain lower amphibolite or upper greenschist facies assemblages. They are thus interpreted as having occurred at a shallower crustal level than the faults they cut.

Absolute dating of this faulting is difficult, but the retrograde metamorphic path combined with the cooling history indicates that much of the deformation occurred in the latest Paleozoic and early Mesozoic (Wintsch and Lefort, 1984). The 270 Ma date on a folded granite dike cutting Zwe (Stop 2a) further demonstrates the significant role of late Paleozoic deformation as well as metamorphism in this area. Of particular importance, is the potentially complete overprinting of pre-Carboniferous fabric by this complex late Paleozoic polydeformational history.

#### Interpretation of Fibrolite

Another important difference between the conclusions of Lundgren (1962; 1963; 1964) and Wintsch (this study) stems from the interpretation of fibrolite or sillimanite bearing assemblages. In this study sillimanite is considered a metamorphic mineral generated either by (1) prograde metamorphism of aluminum-rich, calcium poor pelitic rocks, or (2) by small scale stress and/or strain-induced metasomatic reactions involving biotite, orthoclase or plagioclase as reactants. The usual "prograde" sillimanite occurs in Ztal (Fig. 2). The most convincing examples of the latter "syntectonic" examples occur in coarse grained pegmatites. In pegmatites in the fault between Zwe and Zwc just north of Centerbrook and to a lesser extent at Stop 2, mats of pure fibrolite up to 2 mm thick and 20 cm long define a complex network of shear zones which cut around and through the very coarse grained feldspar crystals. In this case the reactions producing sillimanite from feldspar, e.g.:



are metasomatic, and are high grade analogues of reactions producing muscovite from feldspar (sericitization). In other examples quartz-

sillimanite nodules occur in tandem in foliated pegmatites, in the meta-volcanic Monson Gneiss, and the gneiss of Chester. In these examples the occurrence of sillimanite reflects an unusual metamorphic environment and does not indicate a pelitic protolith.

One of the more interesting candidates for this syntectonic occurrence of sillimanite is the blastomylonitic gneiss along the Falls River fault (Stop 2). Here layers of sillimanite-biotite-garnet schist are interlaced with quartz-plagioclase gneiss of chemical composition very similar to Zwe, an intrusive orthogneiss. The interlayering of these schists with an intrusive rock makes the interpretation of these schists as Ordovician Brimfield Formation difficult to defend. In spite of the larger scale at this outcrop, the syntectonic, metasomatic explanation offered for the fibrolite above is also endorsed in this fault zone. Other smaller fibrolite bearing schist layers are intercollated with biotite and calc-silicate schists in the Hebron Formation along its faulted NE and SW margins and these are also considered candidates for the metasomatic growth of sillimanite.

Lundgren interpreted all occurrences of sillimanite as indicative of a primary pelitic protolith. The locally fibrolite bearing Zwc was mapped as Putnam gneiss by Lundgren (1963, see Straits Road Section, p. 34) and the locally fibrolite bearing Hebron Formation along the Bonemill Brook fault was mapped by Lundgren (1963, see Bochim Road section, p. 32) as Brimfield Formation. Thus the local occurrence of sillimanite on both sides of the Hebron Formation in the Appendix, which Lundgren (1962) interpreted as stratigraphic repetition in the Chester syncline is interpreted here as reflecting syntectonic, metasomatic reactions associated with the complex strike parallel faulting which isolates the Hebron Formation in the Appendix. The unusual example along the Falls River fault is examined at Stop 2, and an example of syntectonic muscovite (the lower grade sericitization analogue of sillimanite in these reactions) in the fault separating Zwc and O?h is examined at Stop 4.

#### Acknowledgements

I have benefitted greatly from many discussions with H.R. Dixon, R. Goldsmith, L. Lundgren, S. Quarrier, and M. Pease about the geology of south central Connecticut and H.R. Dixon, R. Goldsmith, S. Quarrier and M. Pease read earlier drafts of this manuscript. Field assistance was ably provided by K.K. Thomas (1980), A. Owens (1983) and J. Bernitz (1984). Permission to include private property in this field trip was kindly given by Wildwood Medical Specialists (Stop 1), Mr. Richard Moore (Stop 7) and Stephen and Deborah Moore (Stop 9). A. Held patiently typed this manuscript several times, and B. Moran and R. and B. Hill prepared the Figures. Field and analytical work was supported by the Connecticut Geological and National History Survey, the U.S. Geological Survey, and grants from the U.S.G.S. (14-08-0001-G-632), Research Corporation, U.S. Nuclear Regulatory Commission (to P. Barosh) and the National Science Foundation (EAR-8313807 and EAR 85-4166).

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## ROAD LOG

## Mileage

- 0.0 Assemble at 8:00 a.m. at the commuter parking lot at the southeast corner of the intersection of Rts 153, 9 and 9A (Middlesex Turnpike) between Essex and Centerbrook. The lot can be approached from both the north and south via exit 3, Rt 9. Follow both the direction and the intent of the signs to this lot "PARK AND RIDE" by consolidating into as few cars as possible. Breakfast can be obtained locally on Rt 153 in Centerbrook (Ted's Restaurant) or in Ivoryton under the Ivoryton pharmacy (Aggie's). Set odometer at 00.0 miles. On leaving the parking lot, turn left (south) on Middlesex turnpike (Rt 9A).
- 0.2 Turn left (east) into the driveway of Wildwood Medical Specialists.
- 0.3 Park in extreme eastern lot just south of STOP 1.

STOP 1. Orthogneiss of Essex (Zwe). The rocks at Stop 1 raise several issues of fundamental importance to the problems of the area. The first is the interpretation of the protolith of the plagioclase gneiss. At the heart of earlier interpretations of the structure of the Deep River area is the equivalence of all plagioclase gneiss as a single stratigraphic unit: Monson Gneiss Table 1. The most conspicuous features of this outcrop are the abundant amphibolite blocks in the gneiss of granodioritic composition (Fig. 3). These blocks could have formed either as xenoliths during the initial intrusion of a granodioritic magma, or they could represent dismembered mafic dikes boudinaged into these blocks during later ductile deformation. The xenolith hypothesis is supported by the occurrence of diopside bearing blocks of granofels with amphibolite blocks on the hill north of Viney Hill, in outcrops 4 km SW of Essex on Rt 154, as well as farther to the SW, and the gneiss is accordingly considered an orthogneiss. As such, it cannot have stratigraphic equivalents, although it is probably closely related in time to Zwh and Zwc.

The second question raised by this exposure is the origin of metamorphic fabric. The outcrop exposes several ductile shear zones cutting the coarse grained gneiss (Fig. 3). The gneissosity defined by disseminated biotite flakes dips  $50^{\circ}$  NW, and hosts a lineation defined by biotite streaks plunging  $40^{\circ}$  NNW. Little or no compositional banding exists parallel to this foliation. This is in sharp contrast to the conspicuous grain scale compositional layering which is parallel to the foliation in the narrow ductile shear zones. The contacts of the shear zones are gradational across 2 cm, and the modification of gneissosity to produce the younger, better defined foliation is clear. The deformation of xenoliths to aspect ratios  $>100:1$  further demonstrates the very high strain in these zones. That several shear zones of differing orientation cut this outcrop demonstrates that several fabric forming events occurred in this area. Thus map scale structures cannot be interpreted from foliation attitudes until it can be demonstrated that all foliations compared are of the same generation.

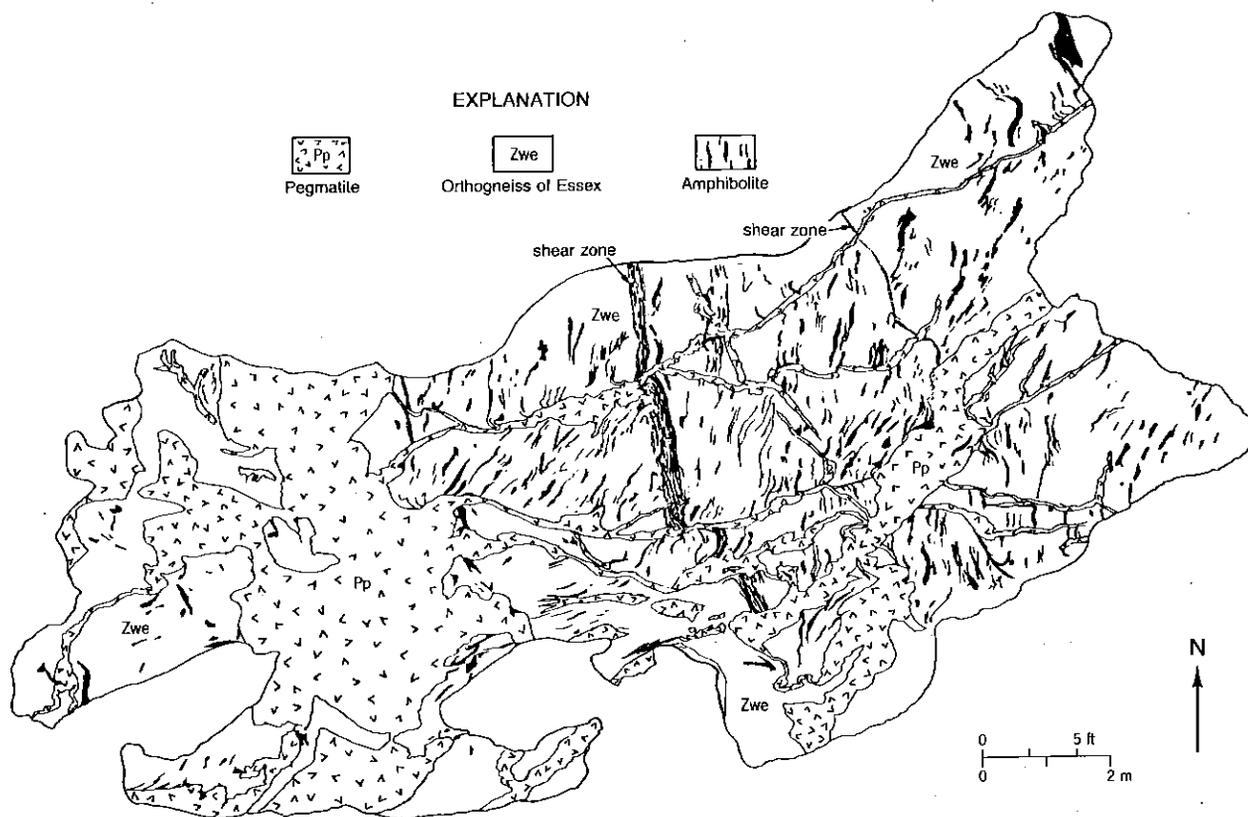


Fig. 3. Detailed geologic map of the northern exposure of the orthogneiss of Essex (Zwe) at Stop 1, based on a photo mosaic (compiled by J. Bernitz). A weak N-S layering defined by amphibolite xenoliths is modified by two narrow ductile shear zones. The outcrop surface is nearly parallel to late cross cutting pegmatite dikes which exaggerates the width of pegmatite exposure.

The small shear zones cutting the gneiss in this outcrop provide a conceptual model for many of the structures in the Deep River area. Most foliation in the Deep River area is interpreted to be variably transposed and modified by dextral and to a lesser extent sinistral shear strain, probably of Variscan age (see Stop 2a), and no specific foliation can be ascribed to the Acadian or Taconic orogenies with confidence.

0.4 Return to Middlesex turnpike and turn right (north).

0.6 Turn right and return to Park and Ride assembly point. Walk along the western and northern edge of the parking lot to the road cuts along the northbound entrance ramp of Route 9.

STOP 2. Blastomylonite Schist (Zwe). Rocks mapped by Lundgren (1964) as pelitic Putnam Gneiss based on the abundance of sillimanite in schistose layers. These rocks are interpreted in the present study as part of the Falls River fault zone. The rocks reveal several structures characteristic of ductile fault zones, including blastomylonitic gneiss that is locally isoclinally folded (especially at the south end of the cut), small and large scale boudinage, penetrative lineations (here sillimanite needles and biotite streaks plunging  $\approx 35^\circ$  NE) that are parallel to small isoclinal fold axes, and highly deformed concordant feldspathic segregations and syntectonic fibrolite mats in some pegmatites. There is also an apparent decrease in strain from south to north which becomes even more apparent when outcrops farther up the hill are included in the strain profile.

I propose that the strain in this Falls River fault zone is not only responsible for the well developed foliation, but might also have driven the reaction consuming biotite and plagioclase and producing sillimanite and K-feldspar. Deformation induced metamorphic differentiation might be responsible for the segregation of quartz and feldspars into augen and veins, leaving the aluminous biotite schists behind. Evidence for this comes from the nearly constant oxide ratios of Fe, Mg, Al and Ti in five samples of plagioclase gneiss collected across this outcrop. Constant element ratios of these less mobile components are similar to the same element ratios of samples of Zwe from nearby outcrops (e.g. Stops 1, 2a). This indicates that addition or removal of alkalis and  $\text{SiO}_2$  could have modified the composition of these plagioclase gneiss, and that Zwe could be the protoliths of this gneiss. This is an alternative to the interpretation that the bulk composition of these schists is inherited from a metasedimentary pelitic protolith. The sedimentary protolith hypothesis is hard to defend given that there is a gradational contact between the schists and the intrusive orthogneiss Zwe.

0.7 Leave the parking lot. Turn right (north) on Middlesex turnpike. Turn right (East) at traffic light.

0.8 Turn left (north) on northbound entrance ramp to Rt 9, passing through the cuts of STOP 2.

1.1 Relatively undeformed Orthogneiss of Essex is exposed in road cuts on the left.

1.4 Begin large road cut on the right at the base of Book Hill. A late vertical shear zone cuts the plagioclase gneiss and folds a late granite dike.

Optional STOP 2a. Orthogneiss of Essex (Zwe). This road cut exposes several key features of this gneiss. The gneiss is a plagioclase gneiss containing a low density of amphibolite xenoliths similar to those present at STOP 1 south of the Falls River fault. Foliation dips gently to the northeast and generally parallels compositional layering defined by amphibolite layers. The gneiss is cut by a sucrosic textured unfoliated granite dike, and this in turn is cut by a vertical dextral shear zone believed to be associated with deformation along the Falls River fault (Fig. 4). Zircons analysed by J. Aleinikoff suggest a 620 Ma age for the gneiss and a 270 Ma age for the dike. This establishes a Permian or older age for most of the deformation in the Late Proterozoic plagioclase gneiss, and a middle Permian or younger age for the steep shear zones that deform the dike and for the associated deformation along the Falls River fault as well as for the late, crosscutting pegmatites. Return to cars.

1.6 Passing large, gently folded generally rare amphibolite layers in the Orthogneiss of Essex.

1.7 Enter valley underlain by schists of Rogers Pond fault zone.

2.3 Exit at exit 4.

2.5 Turn left (north) on Rt 9A toward Deep River. Rt 9A closely follows the strike of the schists in the Appendix.

2.9 Turn left (west) on Kelsey Hill Road. Road passes through Hebron gneiss.

3.2 Park on soft shoulder under Rt 9 overpass.

STOP 3. Hebron Formation (O?h). Start at the east end of the road cut. The Hebron Formation here is typical in both lithology and structure of the rocks immediately east of the Bronson Hill anticlinorium and north of the Honey Hill fault. The gneiss frequently contains alternating 0.5-1.0 cm layers of olive green calc-silicate-rich and biotite-rich schist. Calc-silicate-rich schist is well exposed just below a pegmatite. The schist is especially fissile adjacent to a late cross cutting fault (N70W, 40NE). Such faults are common in Appendix rocks, and from map plan have displacements of 200 m or less. Under the power line the foliation takes on the normal N30W, 70NE attitude. Close to the Rt 9 overpass is a second small shear zone (N80W, 40N) which shows that the north side moves west, consistent with the map plan. Just east of the north bound overpass of Rt 9 the schist is rusty weathering and contains fibrolite. Under the overpass is a mesoscopic fold showing

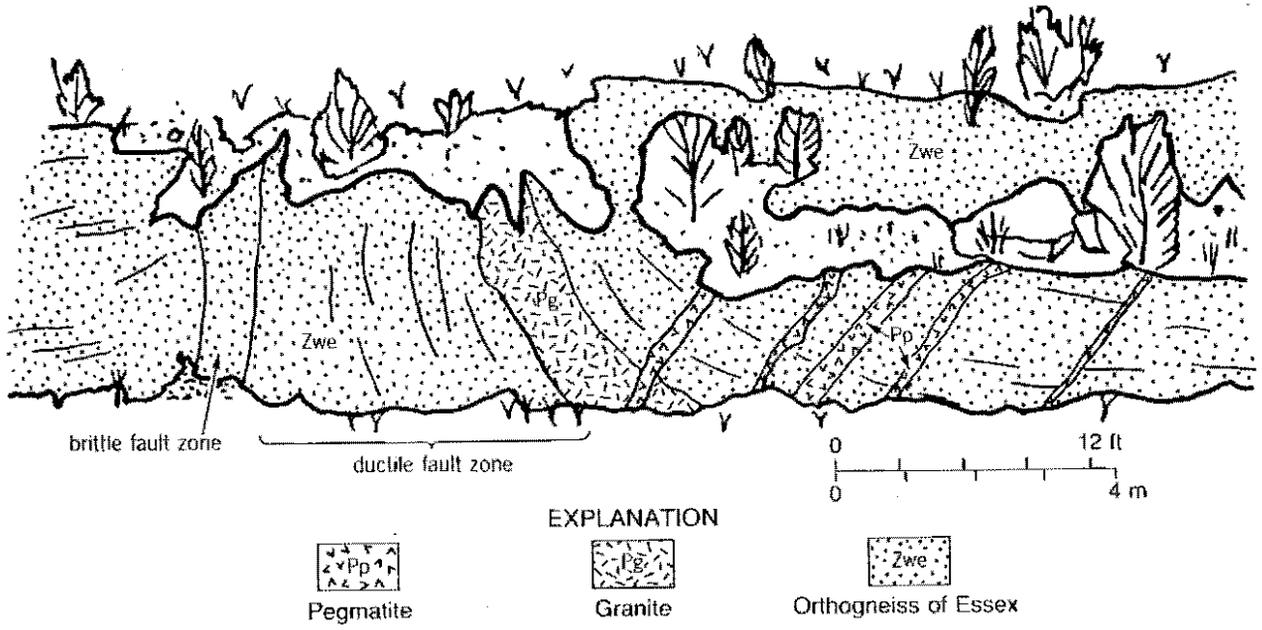


Fig. 4. Tracing of a photo mosaic of the west facing road cut on Rt 9 on the SW edge of Book Hill, 1000 m north of Interchange 3 in Essex. Foliation in Zwe is indicated by dashed lines. The orthogneiss of Essex (Zwe) is cut by late granite (Pg) and very late pegmatite (Pp). A very steep dextral shear zone deforms both Zwe and Pg, but not Pp. A U-Pb age of ~270 Ma (J. Aleinikoff, pers. comm) from zircons from the granite dike demonstrates that the vertical ductile fault zone is early Permian or younger, and thus that the "Permian disturbance" of Zartman et al. (1970) was dynamic as well as thermal.

overturning to the west as does the much larger Deep River anticline. The fine scale layering common to the Hebron Formation is well exposed on the glacier-polished surfaces here. Under the southbound overpass is a plagioclase-rich pegmatite. It is well foliated and boudinaged showing significant post crystallization deformation. Toward the west end of the outcrop is another west facing overturned fold. The steep limb of this fold is cut by a pegmatite dike parallel to the axial plane of the fold. In this case the pegmatite is only weakly deformed and boudinaged.

Return to cars. Turn around, go back to Middlesex turnpike (South Main Street of Deep River).

3.5 Turn left (north) on South Main St.

4.2 Turn right on to South Worth. Park in the lot behind the baseball field.

STOP 4. Deformed Hebron Formation (O?h). The rocks here show the overprinting of greenschist facies fabrics on the higher grade gneisses of the Hebron Formation. Walk west over the small grass covered hill east of North Main St. where coarse-grained plagioclase-rich pegmatites are boudinaged in the Hebron gneiss near its contact with Zwc. In the next 5 m west, a transition to blastomylonitic schist is exposed on the north side of the hill, and fresh samples may be collected in the road cut on South Main St. just south of the intersection.

**THIS IS A DANGEROUS INTERSECTION!! STAY CLOSE TO THE ROAD CUTS!!**

Note particularly the 1-3 mm thick layers of muscovite and chlorite which anastomose through the higher grade assemblages. This is interpreted as an example of reaction softening, where relatively weak phyllosilicates have replaced the stronger feldspar and allowed strain to be localized in these new foliation planes. As the assemblage clearly documents greenschist facies conditions, the outcrop farther provides evidence of later strain at these lower grade conditions.

Return to cars. Leave parking lot, turn right (north) on South Main St.

4.7 Turn a sharp left at the traffic light onto Rt 80 west.

4.9 Cross Union St.

5.2 Cross the rarely exposed fault (potential suture?) separating Hebron Formation (probably of the Nashoba terrane) from the Monson Gneiss of the Bronson Hill terrane. Continue west under Rt 9.

5.7 Turn right (north) on unidentified road (West Bridge St.).

5.8 Park under the first overpass.

STOP 5. Monson Gneiss (Omo). Climb up the grassy bank to the cuts along the exit ramp of Rt 9.

STAY WELL OFF THE HIGHWAY!!

This exposure displays many features characteristic of the Monson Gneiss where it is close to the Bonemill Brook fault zone. Relative to exposures farther west of the fault, this gneiss is generally finer grained, much better foliated and contains conspicuous lineations formed by biotite streaks and less often by quartz-feldspar rods. It also contains conspicuous feldspar augen possibly formed by strain induced feldspathization (Wintsch, 1975). These are best exposed on top of the outcrop on the weathered surface. Pegmatites locally cut the gneiss, and evidence that dikes and sills are forcefully injected is present in the form of mushroom shaped structure on the south face of the outcrop. Large (up to 1 x 4 m) amphibolite boudins are well exposed on the east facing cut. Foliation and lineations conspicuously wrap around these boudins.

Southeast of the major outcrop is a smaller outcrop of pin-striped amphibolite similar to Middletown Formation (Omm). Amphibolite with this alternation of plagioclase-rich and hornblende-rich layers occurs discontinuously along the fault separating the Hebron Formation from the Monson Gneiss. This layering strongly contrasts the massive structure of the amphibolite in the boudins, and is interpreted here as the result of high strain, similar to the shear zones at STOP 1. This outcrop also contains internal boudinage, where boudin necks are filled with quartz alone. The lack of plagioclase in these veins suggests that the boudins formed under lower amphibolite or greenschist facies conditions where quartz dominates vein assemblages.

Return to cars.

- 6.1 Continue east on West Bridge St. passing road cuts exposing blastomylonitic Hebron Formation on the east side of the potentially major fault.
- 6.3 Stop sign at Union St.--Proceed east.
- 6.7 Turn right (south) on Rt 9A.
- 6.8 Turn left (east) on High St.
- 6.9 Turn left on River St.
- 7.6 Cross Valley Railroad tracks and park in the lot by the Connecticut River.

STOP 6. Potter Hill Gneiss and late granite (Zsph, Pg). In the railroad cuts, and on the natural exposure on the west side of the lot can be seen the interlayering of a medium- to fine grained, even textured or sucrosic pink magnetite-bearing granite with a medium

grained, gray biotite-bearing granite gneiss. The pink sucrosic granite dominates at this exposure, and boudinaged amphibolite layers are also present in a low natural exposure just west of the tracks. The sucrosic granite forms sill-like structures on the natural exposure on the west side of the lot, suggesting that it is a relatively late intrusive unit, and post dates much of the fabric forming deformation in these rocks.

Return to cars. Leave parking lot, bearing right on to Kirland.

8.1 Hill on left contains mixed sucrosic granite and biotite-granite gneiss. A single lens of amphibolite 30 m long is exposed in a cliff face.

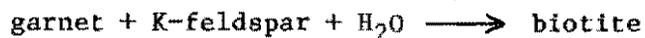
8.4 Turn right (north) on to Rt 9A.

9.2 Turn right (east) at blinking light on to Old Depot Road on the north side of the Chevron Station.

9.3 STOP 7. Honey Hill Fault (Ztal, Zwe). Probably the best place to visit the Honey Hill fault zone in this area is along Old Depot Road. Several exposures of the garnet-biotite-rich lower member of the Tatnic Hill formation can be examined in the gardens of the large stucko house north of Old Depot Road.

THIS IS PRIVATE PROPERTY--OBTAIN PERMISSION FROM OWNERS BEFORE EXAMINING ROCKS!!

Exposures of amphibolite-bearing garnet biotite schists in the front yard of the house and under the patio along the NE side of the house reveal sets of asymmetric folds which coalesce to form ductile shear zones. Garnet bearing pelitic gneiss is better exposed 60 m SE of the house just north of Old Depot Road. Here two sets of ductile shear zones at high angles to the regional NNW dipping foliation outline four sides of a block of gneiss reminiscent of the tectonic blocks present along the Willimantic fault (Wintsch, 1979). The retrograde reaction (approximately):



can readily be seen in these shear zones. The biotite product forms blade-shaped streaks and sheaths around the garnet porphyroblasts, suggesting that the reaction was syntectonic. The biotite being weaker than the reactant minerals provides an example of reaction softening, and as it is a retrograde reaction occurring in the lower amphibolite facies, it documents activity in the Honey Hill fault zone during waning (lower amphibolite-upper greenschist facies) metamorphic conditions. Small road cuts along Old Depot Road provide the best location for collecting fresh samples. Just south of Old Depot Road in a narrow vacant lot, amphibolite of Zwc is exposed. One sample of this rock was analysed by Lundgren (1963) and is the metamorphic equivalent of an olivine tholeiite. The actual contact between the pelitic Tatnic Hill

formation and these amphibolites is not exposed, but drill core from Gillette's Castle shows that the contact contains only ductile (lower amphibolite facies) mylonitic schists. Rocks at this contact, indicated by Lundgren (1963) as the locus of deformation along the Honey Hill fault, do not contain the retrograde metamorphic assemblages present in blastomylonites stratigraphically and structurally overlying these rocks.

Return to cars. Turn around, return to Rt 9A.

9.5 Right (north) on Rt 9A.

10.2 Left (west) at traffic light on Rt 148 toward downtown Chester.

11.1 Straight through stop sign.

11.6 Right (north) on Pleasant St. following signs to "Elementary School."

11.9 Right (east) on Ridge Rd.

12.1 Right (south) to elementary school property. Drive around school to basketball court on south side of school.

STOP 8. Canterbury Gneiss (Dc). Follow the small trail south west for 60 m. Turn right (west) toward 2 m high rubbly outcrop of Canterbury gneiss. Note the gentle northward dip of the coarse gneissosity. Continue west 30 m along the contour to the break in slope at the edge of the cliff. Here the foliation in the rock is abruptly overturned. The fine grain size and the conspicuous compositional banding in this gneiss in the small abandoned quarry are thought to be caused by relatively high strain in these overturned zones. Similar zones of overturned foliation between the Pataconk Brook and Great Brook faults are mapped as reverse fault zones on Fig. 2. The detailed relationships of this area are shown on Fig. 5.

Return to cars. Leave parking lot. Return to Ridge Street.

Left on Ridge St.

12.6 Right (north) on Pleasant St.

12.9 Left at the stop sign on East Wig Hill Road

14.2 Right at stop sign on Bartkiewicz Road.

14.3 Park across from the fifth house on the left of the road.

STOP 9. Sheared Pegmatite. The pegmatite is well exposed in the rock garden behind the house, but it is PRIVATE PROPERTY. OBTAIN PERMISSION BEFORE ENTERING.

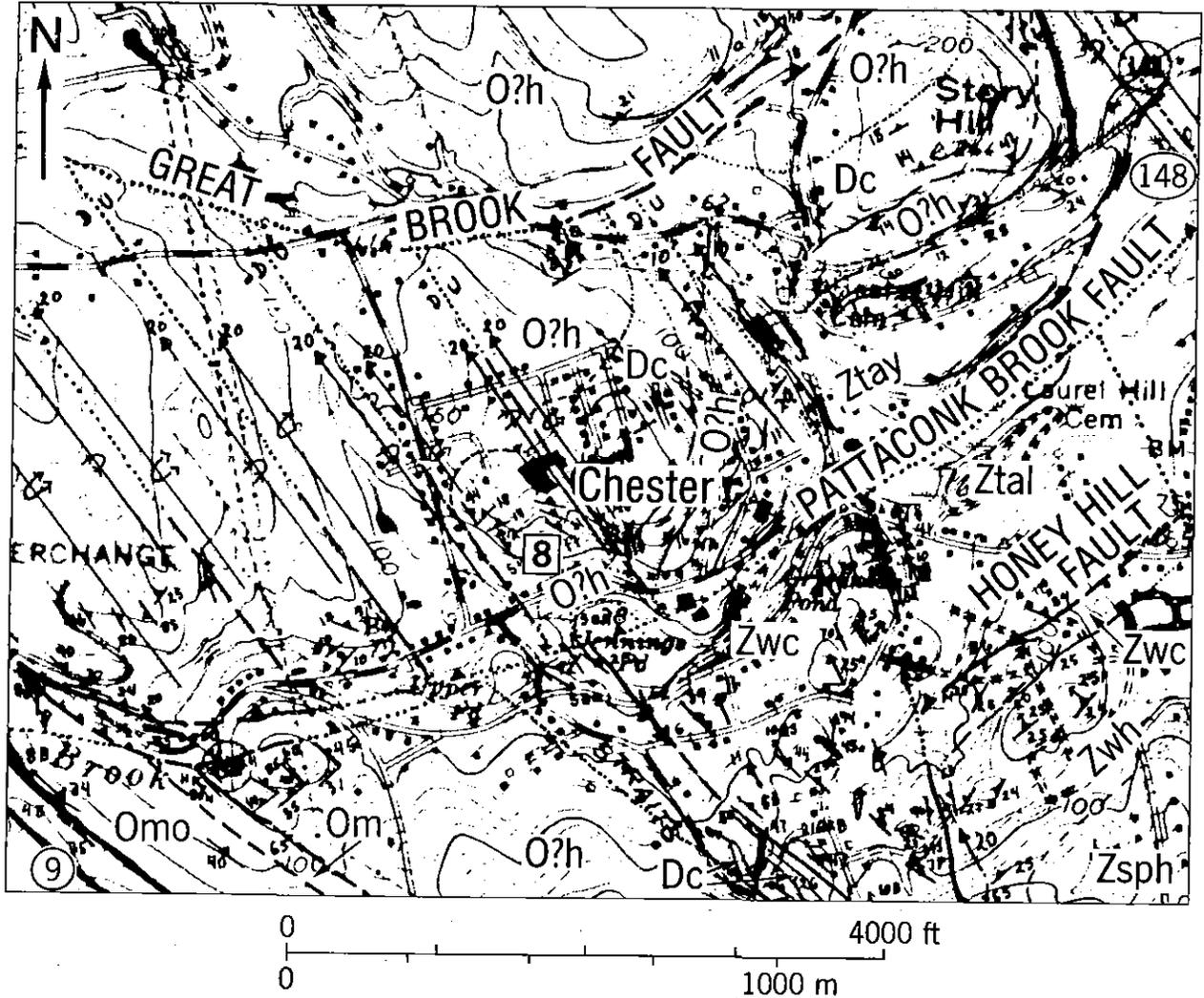


Fig. 5. Detail of the geology of Chester, showing the 'S' shaped overturned folds between the Pattaconk and Great Brook Faults. These, as well as abundant outcrop scale folds all mimic the much larger scale 'S' shaped fold defined by the Late Proterozoic Avalon Terrane next to the Ordovician Bronson Hill terrane (see also Lundgren and Ebblin, 1972, p. 2783).

The features to be seen in this outcrop are the very shallow dip of the pegmatite dike cutting the very steeply dipping plagioclase gneiss of the Turkey Hill belt of Monson Gneiss. The pegmatite is on strike with the Pattaconk Brook fault, and probably intruded the fault zone during movement on it. Consistent with this, a weak layering defined by a smaller grain size cuts the coarse grained quartz-feldspar aggregate and is parallel to the boundaries of the dike. These layers are thought to be caused by deformation of the pegmatite after crystallization, and reflects a second stage of movement on the Pattaconk Brook fault.

Return to cars. Turn around.

14.4 Bear right on Wig Hill Road.

14.8 Left (east) at stop sign, onto Rt 148. Valley of Rt 148 is a major splay of the Pattaconk Brook fault.

15.7 Park in Commuter parking lot and walk to outcrop of entrance ramp to Rt 9.

STOP 10. Monson Gneiss (Omo). This is a typical exposure of anthophyllite bearing Monson gneiss, but displays a more complicated deformation history than many single outcrops. The earliest prominent deformation occurred under upper amphibolite facies metamorphic conditions where ductile deformation folded preexisting foliation into tight isoclinal folds, deformed and boudinaged amphibolite and early pegmatites, and established a subhorizontal lineation defined by anthophyllite blades and biotite streaks. The resulting foliation dips  $\sim 30^{\circ}$ NE and locally contains syntectonic feldspar porphyroblasts. This deformation is interpreted as Late Pennsylvanian in age, and reflects sinistral strike slip motion in the faults bounding all Appendix lithologies.

This foliation is cut by several pegmatites, the walls of which show small offset. They could be related to pegmatites intruded into the Pattaconk Brook Fault (Stop 9), but their WNW dip is not easily reconciled with this fault. Later deformation produced several narrow shear zones dipping steeply NE which contain fine-grained quartz and feldspar. These structures are all cut by at least 6 very narrow shear zones which contain both muscovite and chlorite in a NW dipping foliation. This assemblage apparently developed under greenschist facies conditions, and records the relatively low grade metamorphic conditions of this latest stage of deformation.

The outcrop records an unusually large number and variety of structures, possibly because of its position in the footwall of the Pattaconk Brook fault. Whatever the cause, the multiple stages of deformation from upper amphibolite to greenschist facies metamorphic conditions provides on a small scale an example of a similar protracted deformational history over a similar range of crustal levels experienced by the entire Deep River Area.

Leave parking lot. Turn right (east).

- 15.8 Turn right (south) on Rt 9 passing STOP 10 towards Old Saybrook.
- 16.3 Passing through road cuts of well foliated and well layered Monson gneiss. Amphibolite boudins are separated by as much as 30 m.
- 17.9 Passing Exit 5. (STOP 5)
- 18.1 Passing through cuts of Monson gneiss with conspicuous biotite streaks up to 30 m long.
- 18.6 Passing into schists of the Appendix.
- 18.9 Passing over Hebron formation of STOP 3.
- 20.2 Passing plagioclase gneiss of Book Hill on left (STOP 2a).
- 20.6 Turn right at Exit 3, passing smaller cuts of orthogneiss at Essex (Zwe).
- 20.8 Turn left at stop sign, driving under Rt 9.
- 20.9 Pass STOP 2 blastomylonitic gneiss of the Falls River fault.
- 21.0 Left into Park and Ride Parking lot. End of trip.



## RECESSIONAL MORAINES, SOUTHEASTERN CONNECTICUT

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## SETTING

The recessional moraines of southeastern Connecticut and adjacent Rhode Island occupy a belt about 20 km wide north of the terminal position of the Laurentide ice sheet on Long Island, Block Island, Martha's Vineyard, and Nantucket (fig. 1<sup>1</sup>; Schafer and Hartshorn, 1965; Flint and Gebert, 1976; Goldsmith, 1982). The recessional moraines are characterized by an alignment of morainic segments approximately perpendicular to the direction of ice movement as indicated by striations and grooves on bedrock surfaces (fig. 2). In places, the moraines are double or paired. The segments are mostly about 100 m wide and 2 - 6 m high and consist of sandy, bouldery, generally loose-textured glacial till, with or without beds and lenses of well to poorly sorted stratified drift. Existing exposures of the material are rare. Surface form may be smoothly rounded or undulating and more bouldery than adjacent smooth-surfaced till, or the surface may be irregular, hummocky, and in places contain closed depressions. In some places, as at Glacial Park (Stop 3) and at the Waterford Country School (Stop 2), the surface may consist entirely of boulders without interstitial fine material. The moraines are little affected by topography except for some inflection in large valleys such as that of the Thames River. This lack of topographic influence is attributed to the thickness of ice which must have exceeded the local topographic relief when the moraines were deposited over the tops of the hills at the active ice front. Deposits of glacial streams derived from melting ice overlap partly dissected moraines (fig. 3). Ice-contact landforms in meltwater deposits located near the moraines mark the heads of streams that were derived from belts of stagnant ice lying at and north of the moraines after the active ice front had retreated northward to new stillstands or ephemeral positions.

In eastern Connecticut, north of the morainal belt, recessional moraines are rare, probably because of an accelerating rate of ice retreat. Ice retreat in eastern Connecticut is recorded by a succession of ice-marginal deltaic deposits in valleys in which the drainage to the south was temporarily blocked by earlier deposits and isolated remnants of ice. These deposits formed near and in the temporary position of the fringe of stagnant ice marginal to the ice sheet as it retreated northward.

This trip will be primarily a sight-seeing trip to look at landforms and characteristics of the surface of the moraines. As mentioned above, exposures of the material in the moraines are rare and ephemeral, but I have located a couple of places where we might examine the material (Stops 5 and 7). We will look at places where the moraines are covered or partly covered by meltwater deposits (Stops 1, 3, 8), at boulder concentrations in two different topographic settings (Stops 2 and 3), at several kinds of typical forms of the

<sup>1</sup>All illustrations except Figure 5 are taken from Grahame J. Larson and Byron D. Stone, eds., Late Wisconsinian Glaciation of New England, Copyright 1982 by Kendall/Hunt Publishing Company, and reprinted by permission of Kendall/Hunt Publishing Company.

Figure 1. Distribution of moraines and related features in southeastern New England.

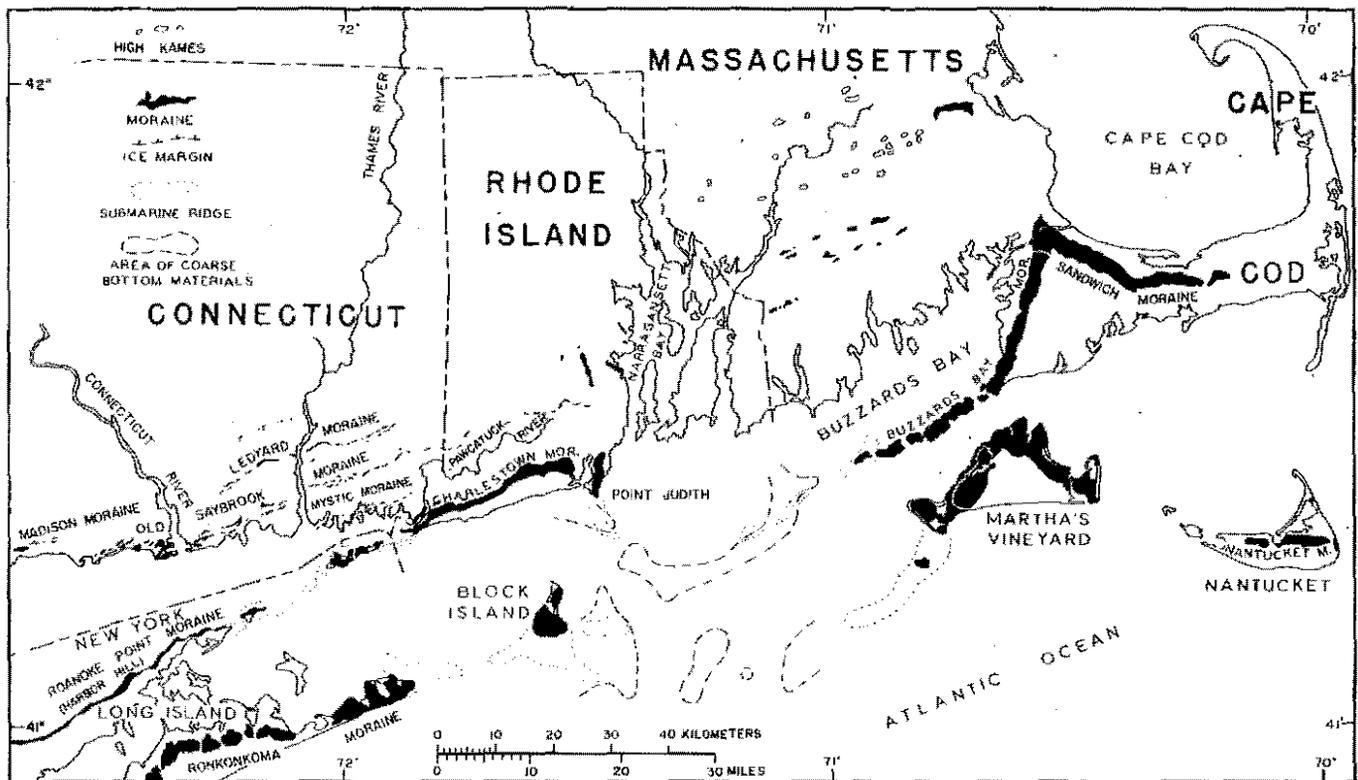


Figure 2. Distribution of morainal segments (black) in the New London area, Connecticut-New York, and direction of ice movement (arrows) as indicated by striations and grooves on bedrock surfaces. Numbers indicate field-trip stops.

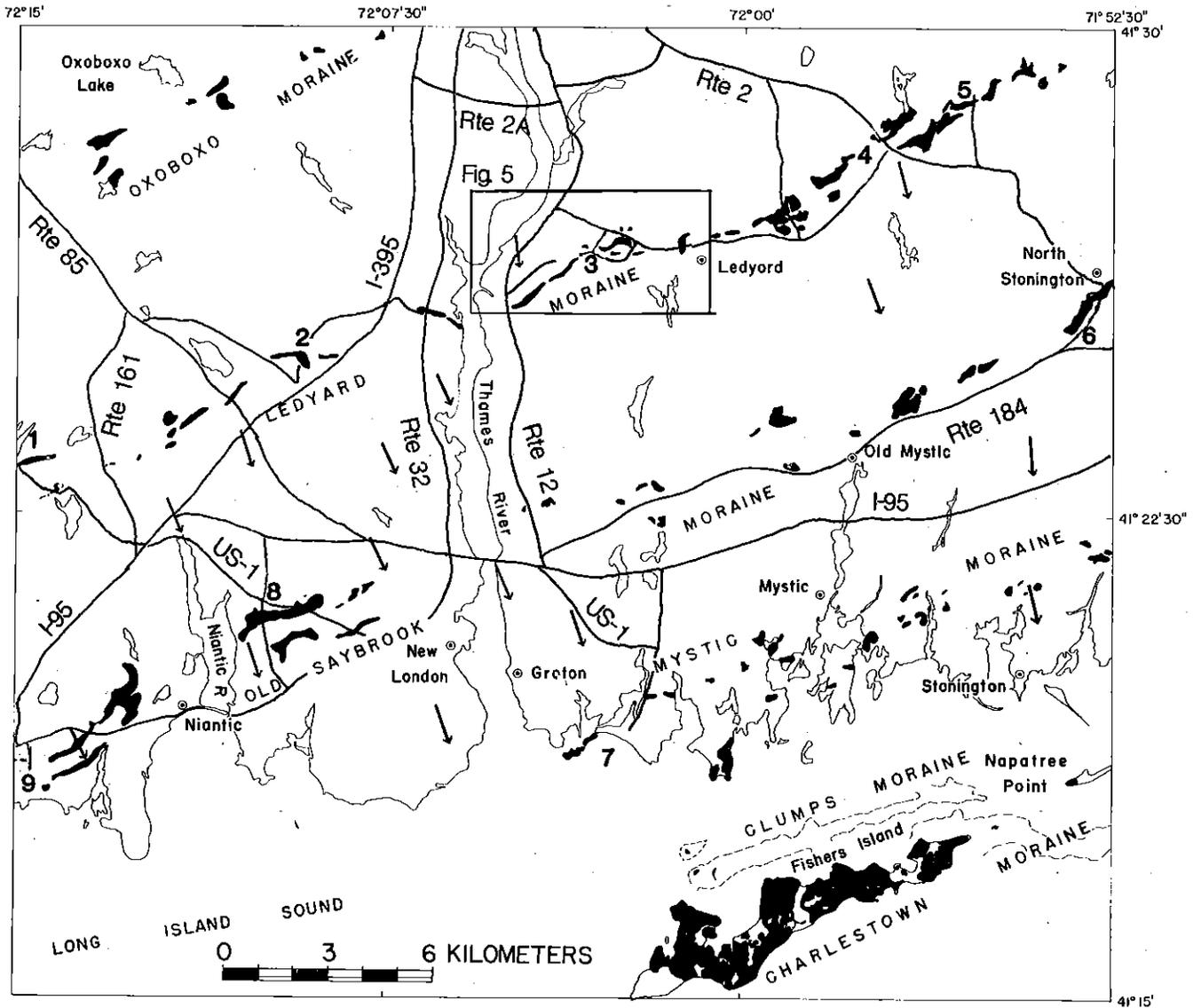
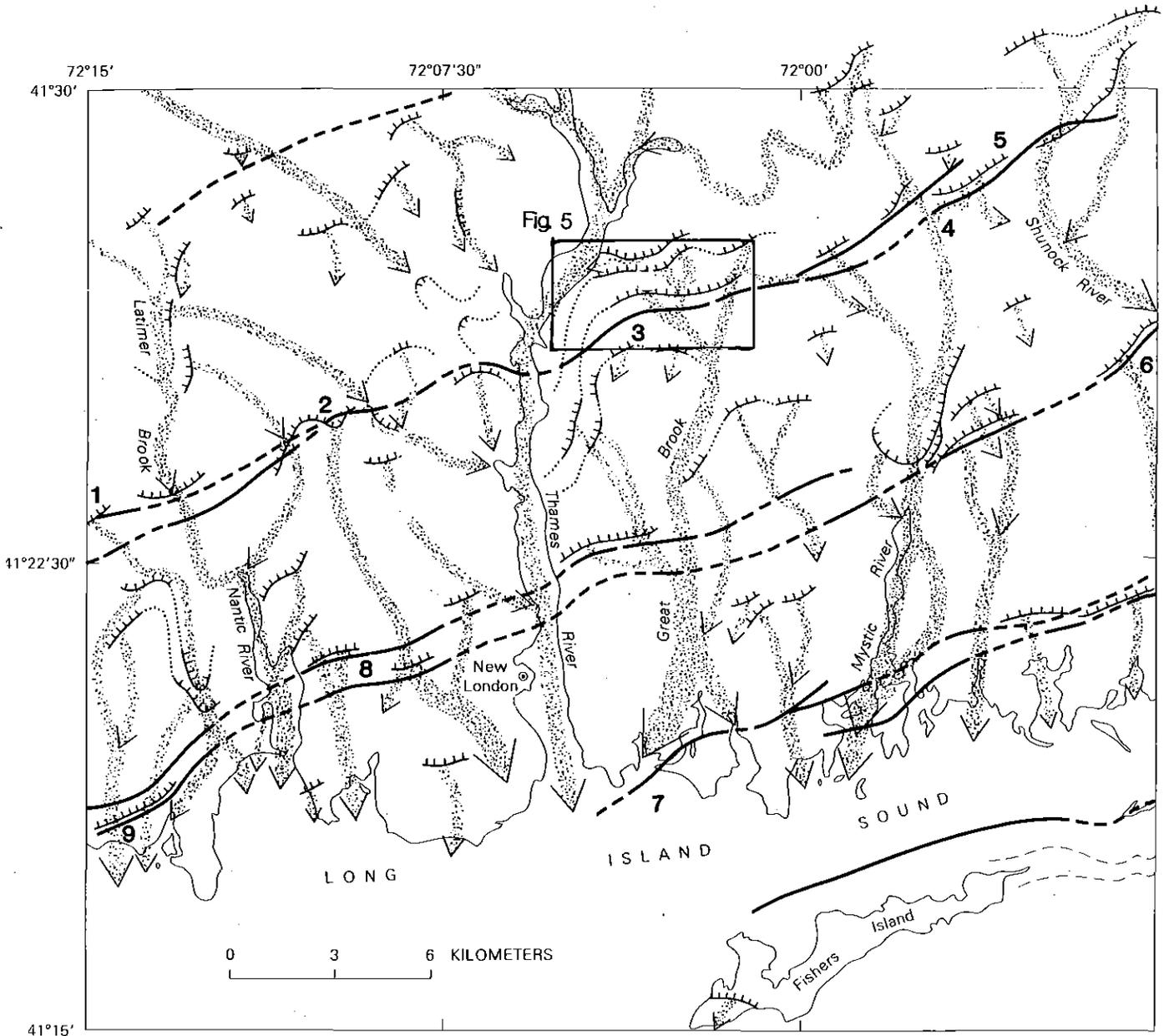


Figure 3. Distribution of heads of glacial stream deposits and their relation to recessional moraines in the New London area, Connecticut-New York. Numbers indicate field-trip stops.



EXPLANATION

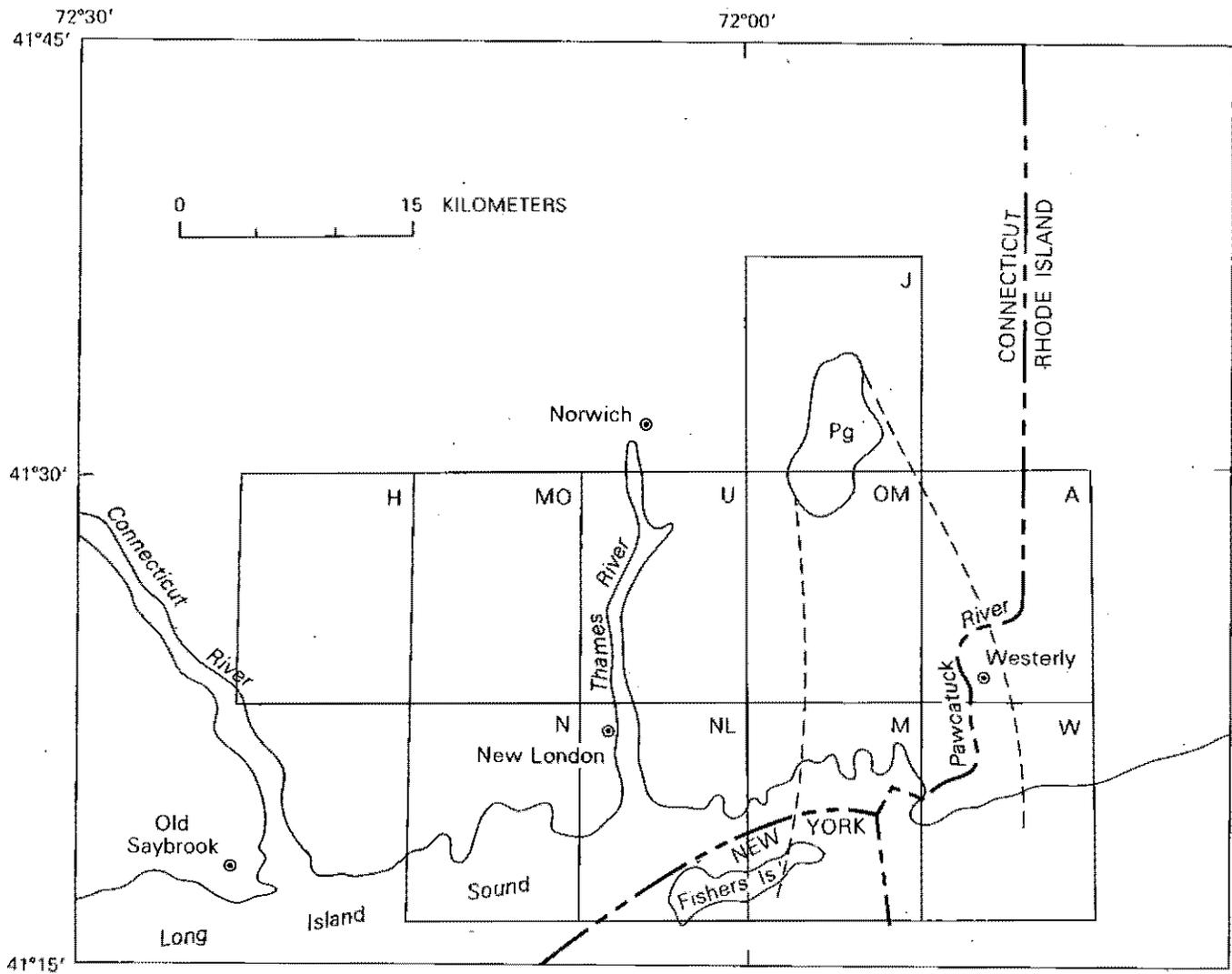
- Moraine; dashed where inferred
- Head of glacial stream deposit and margin of stagnant ice; hachures on up-valley side. Possible continuations of stagnant ice margin dotted; directions of flow of glacial streams shown by stippled arrows.

moraines (Stops 1, 3, 4, 5, 6, 7), and at the relation of the moraines to meltwater deposits (Stops 1, 3, 5, 8). Sources for the mapping on which this trip is based are shown in figure 4. The summary given above of the nature of retreat in southeastern Connecticut of the late Wisconsinan Laurentide ice sheet is derived from Goldsmith (1982).

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Figure 4. Index map showing locations of quadrangles providing data. J-Jewett City, H-Hamburg, MO-Montville, U-Uncasville, OM-Old Mystic, A-Ashaway, N-Niantic, NL-New London, M-Mystic, W-Westerly. PG-Preston Gabbro. Dashed line indicates borders of boulder fan of Preston Gabbro.



Upson, J. E., Surficial geologic map of the Mystic quadrangle, Connecticut, New York and Rhode Island: U.S. Geological Survey Geologic Quadrangle Map GQ-940.

Wells, D. A., 1890, Evidences of glacial action in southeastern Connecticut: Popular Science Monthly, v. 37, p. 691-703.

#### ROAD LOG AND STOP DESCRIPTIONS

Miles  
Cum. Int.

- 0 0 From assembly point, the parking lot of the Connecticut Yankee Travelodge on Rte 161 just south of I-95 at Exit 74 (Niantic-Flanders) East Lyme, turn left (north) on Rte. 161.
- 0.5 0.5 Turn left on U.S. 1, Boston Post Road at Flanders.
- 2.8 2.3 Turn right on Scott Road before reaching Scott's Orchard outlet.
- 4.2 1.4 Ledyard moraine forms rise in road and top of hill.
- 4.8 0.6 Turn right (east) on E. Pataguanset Road. Yale Outdoor Education Facility on left.
- 5.9 1.1 Ledyard moraine forms the ridge under farm. Turn right on Hickory Road along moraine.
- 6.1 0.2 Turn right on dirt road. Park.

STOP 1. Ledyard moraine and glacial stream deposits -- The Ledyard moraine forms the low ridge we just traversed. This is a northern belt of a double segment of the moraine; a southern belt is less well expressed to the southeast, near the head of Pataguanset Lake. At this point, we should be in deltaic ice-contact deposits but these have been removed here for use as sand and gravel aggregate. These stream deposits head north of the Ledyard moraine and have partly buried it (fig. 3). Many examples of this sort of relationship exist in the coastal area. The moraine segment we crossed on Scott Road is a continuation of this belt of moraine. On the hill to the northeast the moraine does not have distinctive form and consists of dispersed boulders or has no expression at all. The only place in the coastal moraines where I have seen deformed beds was in the excavation for the basement of the house at the southwest end of the ridge we just traversed.

- 6.3 0.2 Turn right onto E. Pataguanset Road.
- 6.4 0.1 Cross moraine. Travel along edge of meltwater deposit derived from area of the Ledyard moraine (fig. 3).
- 7.1 0.7 Intersection with U.S. 1. Turn left.
- 7.7 0.6 Intersection with Route 161. Turn left. Road lies along edge of a series of glaciofluvial - glaciolacustrine deposits of morphosequences in Latimer Brook valley (fig. 3).

9.3 1.6 Cross line of Ledyard moraine. To right, the moraine is partly buried by glacial stream deposits along Latimer Brook. Boulders can be seen in excavations and locally projecting through the surface. Continue to travel along edge of slightly younger stream deposits.

12.6 3.3 Intersection with Rte 85. Turn sharp right.

13.1 0.5 Turn left on Turner Road.

14.6 1.5 Bear right on E. Lake Road.

15.2 0.6 Turn sharp left onto Hunts Brook Road.

15.9 0.7 Waterford Country School.

STOP 2. Ledyard moraine on north slope of hill. -- Walk to west of Waterford Country School about 50 yards to boulder concentration forming a bench on the north slope of the hill at about the same level as, or slightly higher than, the school. This sort of boulder concentration without fine material is an unusual form of the moraine but is present in several moraine segments in the coastal area. The mechanism for accumulating this sort of material is not wholly clear, but sorting by glacial meltwater in an ice-marginal crevasse environment is a proposed explanation. A concentration similar to this lies east of Waterford Country School along the same northeast trend but is difficult to reach. The moraine rises from the valley to cross over several bedrock controlled ridges and is unaffected by this topography. We shall see another boulder concentration at Stop 3 in a slightly different topographic setting.

Continue north on Hunts Brook Road.

16.2 0.3 Turn right on Unger Road.

17.5 1.3 Cross Old Colchester Road.

18.6 1.1 Overpass I-395. Ledyard moraine crosses road down slope beyond overpass.

18.8 0.2 Turn right at foot of hill.

19.2 0.4 Cross Ledyard moraine. Note bouldery, irregular surface.

19.3 0.1 Turn left on Star Road.

19.5 0.2 Turn left on Rte 32, proceed north.

19.9 0.4 Uncasville. For a short distance beyond Uncasville we are traveling along the edge of ice-contact deltaic deposits forming terraces along the Thames River. Former active gravel pits here are now overgrown or are covered by construction.

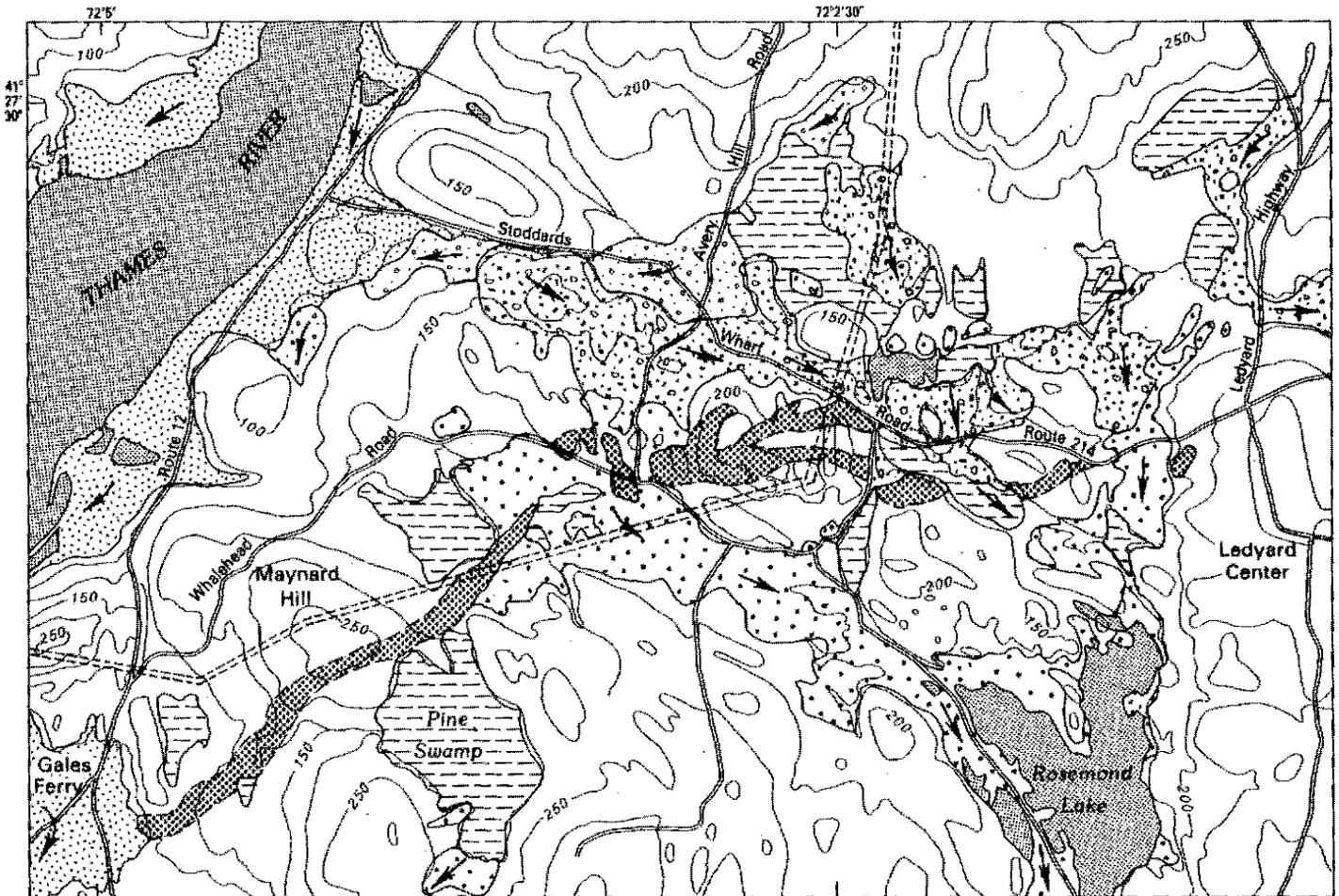
23.4 3.5 Turn right (east) on Rte 2A.

- 25.2 1.8 Turn right (south) on Rte 12. Travel along Thames stratified deposits.
- 27.4 2.2 Turn left (east) onto Rte 214, Stoddard Wharf Road.
- 27.5 0.1 Leave Thames terrace level (50 feet). Higher and older ice-marginal deltaic deposits on right (fig. 5). Outcrops of quartzite and schist of the Plainfield Formation on left.
- 28.0 0.5 Rise to higher terrace level at 170-180 feet.
- 28.2 0.2 Turn right on Avery Hill Road Extension. Kettles in 180 foot surface on right and left are in ice-contact deltaic deposits formed behind the moraine at the head of the glaciofluvial deposits in Great Brook valley.
- 28.6 0.4 Rise to highest and oldest terrace at 190 feet.
- 28.8 0.2 Turn left on Whalehead Road.
- 30.0 0.2 STOP 3. Glacial Park (Maire, 1976) -- Take the trail north from the power line and follow the main trail (fig. 5) clockwise around and over the boulder accumulation first noticed by Wells (1890) marking this segment of the Ledyard moraine. Beyond the top of the hill, the trail crosses back over the moraine which is here approaching a more normal aspect. Boulder accumulations, but less spectacular, are present to the northeast. The moraine continues diagonally across this bedrock-controlled ridge towards the continuation of Whalehead Road near its junction with Stoddards Wharf Road where the moraine consists of bouldery till. We are on what seems to be the southern belt of a double moraine. The northern belt passes from the area of the ravine on the west side of the boulder accumulation up and over the north side of the hill. To the west of Glacial Park, the moraine appears to be dispersed and is marked only by scattered large boulders on an irregular and largely bedrock-controlled surface. The moraine emerges, however, as a discrete feature from beneath glacial stratified deposits northeast of Maynard Hill (fig. 5).

The unusual concentration of boulders without interstitial fine material is attributed to an abundance of granitic source material accumulated from beneath the glacier (the dirt machine of Koteff, 1974) and to sorting of this material by slumping contemporaneous with winnowing by meltwater, probably in a crevasse. The presence of ledges of bedrock beneath the moraine at the side of the ravine and to the east near the power line indicate that till is relatively thin in the area. The boulders at Glacial Park and elsewhere in the Ledyard moraine are derived primarily from rock types that have widely spaced joints and poor to no fissility. Most boulders at Glacial Park are derived from ledges within 1 or 2 km to the north and generally within 5 km (fig. 6; Goldsmith, 1967), but a few have been transported greater distances.

Return to the power line, cross Whalehead Road, and follow the power line southwest along the 180-190 foot surface of the oldest and highest glacial stream deposit heading in the area of the Ledyard moraine (fig. 5). This deposit heads just north of the moraine and was laid down soon after the ice front retreated north of the line of the moraine. Meltwater drainage was to the southeast and south through valleys in the upland because the Thames valley at this time was filled with ice (Goldsmith, 1960). Beyond

Figure 5. Ledyard moraine and stratified drift in the Gales Ferry - Ledyard Center area.



Base by U.S. Geological Survey  
Scale 1:24,000

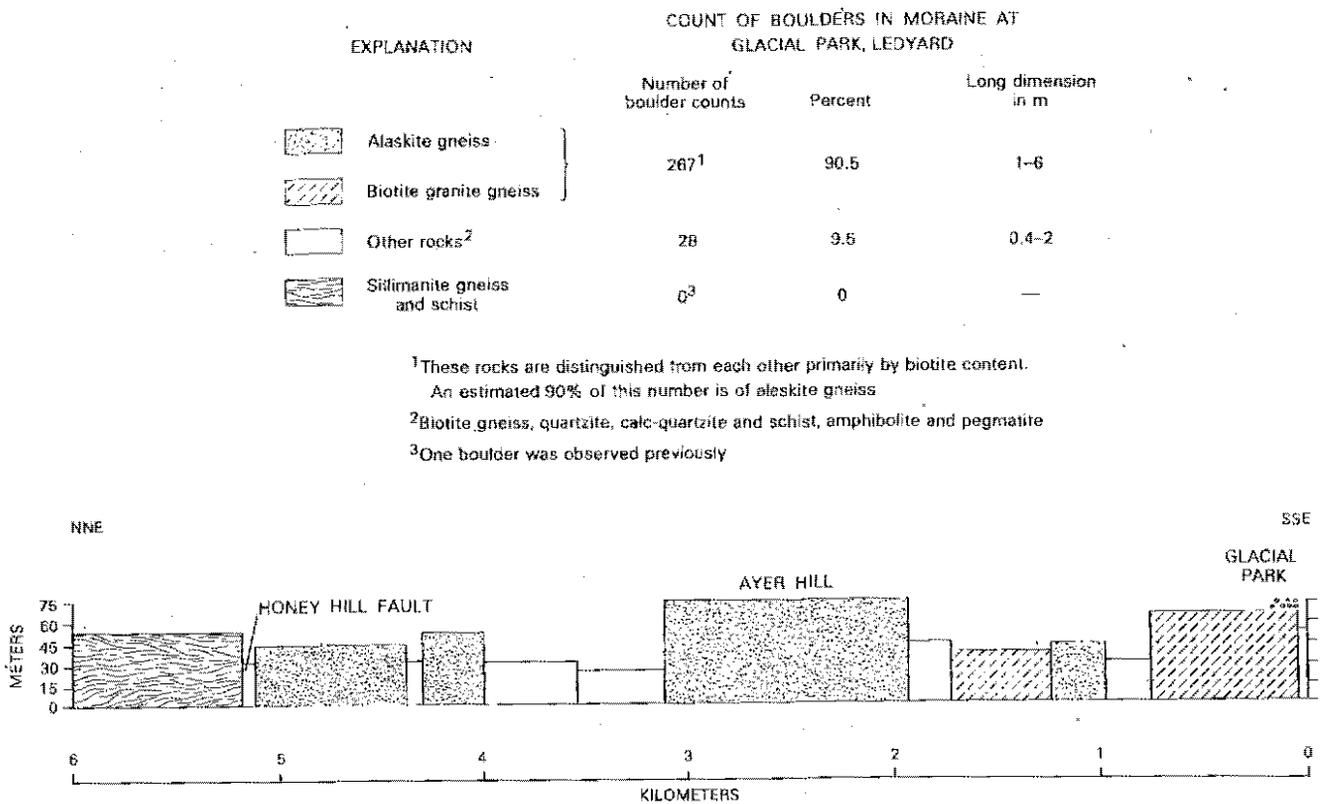


Contour interval 50 feet

EXPLANATION

- |  |                                       |  |  |
|--|---------------------------------------|--|--|
|  | Swamp deposits                        |  | End moraine deposits (Ledyard moraine) |
|  | Meltwater deposits along Thames River |  | Till and bedrock                       |
|  | Meltwater deposits Stage 3            |  | Power line                             |
|  | Meltwater deposits Stage 2            |  | Direction of flow of glacial stream    |
|  | Meltwater deposits Stage 1            |  |  |

Figure 6. Source of boulders in boulder concentrations in Ledyard moraine at Glacial Park, Ledyard Connecticut. Section shows widths of outcrop of rock units along a line extending N.10°W. from Glacial Park, and generalized altitudes along and near the line. Rock units dip about 50°N.



artificially ponded areas where sand and gravel have been removed, the power line crosses a ridge of bouldery till (about 0.8 km from Glacial Park) that is the Ledyard moraine. The poorly sorted material in the moraine is exposed in a ditch dug across the ridge slightly north of the power line.

The moraine continues southwest as a characteristic low ridge of bouldery till up the northeast side of Maynard Hill. This low ridge can be traced over the top of the hill just south of its crest down to and beneath one of the terraces along the Thames River. The moraine reappears west of the Thames River south of Uncasville as a belt of hummocky bouldery till which we crossed earlier in the morning. Here it trends west-southwest. The change in trend indicates a slight lobation of the active ice front down the Thames River valley (figs. 2 and 3).

The kettled topography we passed on Avery Hill Road before arriving at Glacial Park is on the next younger surface of the post-morainial glacial stream system, at 170-180 feet. The glacial streams forming this surface were blocked from flowing directly southeast and at first ponded behind the earlier deposits, the hills and the moraine from flowing directly southeast. They eventually flowed east before passing through gaps in the bedrock ridges to entrench themselves slightly downvalley into the deposits of the earlier stage. The kettled surface indicates that masses of stagnant ice remained just north of the moraine as the front of active ice receded to the north. Patches of still younger stream deposits that form a surface at 130 feet are present near the intersection of Avery Hill Road and Stoddard Wharf Road. The path for this meltwater drainage, however, was west to the Thames valley because the level of ice in the valley had lowered sufficiently to provide an outlet in this direction. The low terraces along the Thames River were deposited by a succession of still later glacial streams when ice remnants were at a low level in the Thames valley. These deposits, where exposed, are seen to be largely deltaic in character.

Continue east on Whalehead Road.

30.4 0.4 Bear left at road junction.

30.7 0.3 Cross moraine.

30.9 0.2 Turn right on Rte 214, Stoddard Wharf Road.

32.0 1.1 Cross Rte 117.

33.1 1.1 Spicer Hill Road. Continue Rte 214.

33.7 0.6 Sawmill Park. LUNCH.

34.3 0.6 Turn left on Shewville Road.

34.9 0.6 Moraine on right.

37.6 2.7 Turn right (east) on Rte 2. Pass onto surface of Shewville-stage stratified drift which is graded to threshold to west and drained to the Thames River because a normal course to the south was blocked by ice-contact deposits along the line of the Ledyard moraine.

- 38.4 0.8 Ledyard town line. Pits near here are in deltaic kame terraces formed at the head of older meltwater drainage flowing south down Lantern Hill Brook valley from the area of the Ledyard moraine.
- 39.7 1.3 Preston Gabbro on left, Lantern Hill ahead. Preston Gabbro forms a boulder-fan recognizable as far as Fishers Island. Lantern Hill is a resistant mass of vein quartz from the top of which King Philip of the Pequots watched in mounting concern as ships arrived carrying settlers from England.
- 40.1 0.4 Ledyard Rest Stop.

STOP 4. We are just south of the southern belt of the double Ledyard moraine in this area. To the southwest, the moraine has crossed diagonally over the large bedrock ridge where it is expressed as a discontinuous low ridge without appreciable boulder concentration until the lower slopes are reached. It passes beneath the ice-contact head of deposits of a glacial stream originating just north of the moraine. These deposits have blocked the drainage in the Cedar Swamp area to the north so that the drainage is forced to flow west to the Thames River instead of south down Lantern Hill Brook (fig. 3). We will pass to the south of the moraine as we proceed east. Here the moraine appears on the southeast side of the hills of gabbro to the north as very bouldery till. A linear boulder accumulation to the north at the southern end of Lake of Isles marks a parallel belt of moraine (figs. 2,3).

- 40.3 0.2 Turn left on Milltown Road. Southern branch of moraine to left behind cemetery.
- 40.8 0.5 Turn left on Rte 2. Outcrops of alaskite gneiss.
- 42.0 1.2 Turn left on Swantown Road. Outcrops of quartzite in the upper part of the Plainfield Formation along Rte 2 before the intersection.
- 43.0 1.0 Cross Ledyard moraine, southern branch. Moraine has very little expression here.
- 43.1 0.1 Moraine forms slope and low rise in road. Closed depression containing swamp is to right of road ahead. Rise in distance is bedrock controlled. Turn around.
- 43.4 0.3 Stop along stone wall beyond red house.

STOP 5. Ledyard moraine in Swantown Brook valley. -- Follow woods road to west and then south to hummocky, bouldery area on east side of head of Lantern Hill Brook. This is another expression of the moraine. The hummocky topography may indicate deposition of end-glacial material on stagnant ice. The topography of the slope to the east is suggestive of bedrock control but no bedrock is present. I interpret this area to be also underlain by moraine but covered by wind-blown silt. A bouldery lag is present along the trend of the moraine toward Swantown Road. Return over hill to east to road and cars.

Return to Rte 2.

- 45.2 1.8 Turn left on Rte 2.
- 45.5 0.3 Shunock Brook. Travel along surface of early meltwater deposit.
- 46.2 0.7 Junction Rte 201 south.
- 47.3 1.1 Intersection, traffic light. Continue on Rte 2.
- 48.0 0.7 Turn right (south) on Rocky Hollow Road at Fire Station.
- 48.5 0.5 Rocky Hollow moraine (Gaffney, 1966) to left. Schafer (1968) mapped a segment of the moraine in the Shunock Brook drainage northeast of this hill. Between there and here, across the hill, is a belt of irregular topography which at one place has a closed depression.
- 48.7 0.2 Cross Rocky Hollow moraine. Moraine forms the low ridge to the right. Outwash to left is derived from the area of the moraine. To the north, behind the moraine, are extensive swamp deposits occupying a former ponded area.
- 49.3 0.6 Park near intersection of Rocky Hollow Road and Rte 184.
- STOP 6. Small pit in Rocky Hollow moraine. -- The exposure in this pit consists of sandy, bouldery till containing a few rotten-stone boulders and pockets of poorly sorted sand and gravel.
- 49.4 0.1 Turn right (west) on Rte 184.
- 52.1 2.7 Old high level, ice-contact deposit of streams draining south over upland from area of Rocky Hollow moraine (fig. 3).
- 52.6 0.5 Junction Rte 201.
- 53.9 1.3 Junction Shewville Road. One-half mile to the north and to the right of Shewville Road are abandoned gravel pits exposing the Rocky Hollow moraine beneath a blanket of meltwater deposits that occupy the valley of Whitford and Lantern Hill Brooks. Permission should be obtained from the owner of Riverhead Farm to look at these exposures.
- 57.9 4.0 Center Groton. We are in the vicinity of the Rocky Hollow moraine concealed here by stratified drift derived from the Ledyard moraine area. Rocky Hollow moraine is exposed to the west, mainly south of Rte 184. Turn left (south) onto North Road.
- 60.5 2.6 Cross U.S. 1.
- 60.8 0.3 Midway and railroad. Proceed south on dirt road.
- 61.8 1.0 Cross Mystic moraine.
- 62.5 0.7 Park for walk to Bushy Point.

STOP 7. Mystic moraine at Bushy Point -- Proceed west along the tombolo to Bushy Point, which was once an island and may be again today. The Mystic moraine is less continuous than the Ledyard or Rocky Hollow moraine and consists of dispersed, but more or less aligned, segments (fig. 2). Boulders of the moraine project through the marsh about 1/2 mile north of the beach. This cluster is aligned with the low ridge marking the moraine on the side of the large hill to the east. Offshore, and exposed in a couple of small islands, is the Clumps moraine, marked primarily by a linear shoal aligned parallel to the Charlestown moraine on Fishers Island and trending toward Napatree Point at the west end of the mainland part of the Charlestown moraine.

Return via road to U.S. 1.

- 64.5 2.0 Turn left (west) on U.S. 1.
- 67.0 2.5 Intersection I-95. Proceed west on I-95. Cross Thames River.
- 71.8 4.8 Take Exit 81, Cross Road exit.
- 72.6 0.8 Turn left (south) on Cross Road.
- 73.6 1.0 Turn left (east) on U.S. 1.
- 74.4 0.8 Turn right into restaurant parking area.

STOP 8. Rocky Hollow moraine. -- The exposure here consists of loose bouldery till containing lenses of sand and fine gravel. Boulders are scattered on the surface of the till on the slope to the south. This belt trends southwest. Further up the slope our route will take us across the belt, and at one point near a cemetery is a closed depression. To the northeast, the morainal belt passes beneath sand and gravel of an early meltwater deposit. At one place where sand and gravel have been removed, a train of large boulders aligned northeast across the pit marks the position of the moraine. The extensive north-trending area shown as end moraine south of Oswegatchie on the surficial geologic map of the Niantic quadrangle (Goldsmith, 1964) is now interpreted as an early ice-contact meltwater deposit graded to the meltwater deposits south of the moraine. The segment of the Rocky Hollow moraine that we are on is the northern belt of a double moraine the southern belt of which rises over Mullen Hill to the south of us (fig. 3).

Continue east on U.S. 1.

- 74.7 0.3 Turn right on Ellen Ward Road.
- 75.9 1.2 Turn right on Mullen Hill Road. Moraine to right.
- 76.6 0.8 Road is on the moraine. Note kettle at cemetery. A meltwater deposit spills over the moraine in this area and is graded south over the upland (fig. 3).
- 76.8 0.2 Turn left on Spithead Road.

- 76.9 0.1 Cross the moraine. The moraine is well expressed down the hill to the southwest.
- 78.2 1.3 Turn right on Ropes Ferry Road, Rte 184.
- 79.7 1.5 Bridge over Niantic River.
- 80.6 0.9 Intersection Rte 161, Niantic.
- 81.2 0.6 Intersection Pataguanset Road.
- 81.9 0.7 Rocky Hollow moraine.
- 84.1 2.2 Rocky Neck State Park. Turn left into Park. Proceed south toward beach.
- 84.6 0.5 Park along road.

STOP 9. Rocky Hollow moraine is here partly buried by meltwater deposits in the Bride Brook valley that grade into deltaic deposits of Glacial Lake Connecticut (now Long Island Sound). Again, we have an example of the end-moraine deposit partly covered by succeeding meltwater deposits, this time of streams heading well north of the line of the moraine in the zone of stagnant ice between the Ledyard moraine and the Rocky Hollow moraine (fig. 3).

Return to Rte 184.

- 85.2 0.6 Turn left (west) onto Rte 184 at Park entrance.
- 85.4 0.2 Rocky Neck State Park Connector. Turn right toward I-95.
- 86.0 0.6 Entrance ramps to I-95 north or south. End of trip.

**Mesoscopic and Microscopic Structure of the Lake Char - Honey Hill Mylonite****Zone, Eastern Connecticut**

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**INTRODUCTION**

The Lake Char and Honey Hill faults in eastern and southeastern Connecticut (figure 1) are wide zones of mylonitization localized at the contact between Avalonian-aged metaplutonic and metasedimentary rocks and metavolcanic and metapelitic rocks of the Merrimack synclinorium. Locally, there is truncation of strata against the contact and units in the upper plate are drastically thinned against the fault (figure 2). In general, however, the faults are recognized as wide zones of mylonitization. The faults have been the subject of much speculation, as have their northward continuations (?) the Clinton-Newbury and Bloody Bluff faults. It is common for those who have not worked on these faults to consider them as sutures or major dislocations (e.g. Wilson, 1966) and for those who have worked on them to consider them as having displacements not exceeding several to possibly 10 kilometers (e.g. Castle et al., 1976). In addition to the tectonic significance of the faults, their sense of motion and timing has become a matter of considerable debate. Early workers (Lundgren et al., 1958; Dixon and Lundgren, 1968) considered the faults to be west-over-east (top up) thrust faults probably developed during the Acadian orogeny but with last motions no later than latest Paleozoic. This view (at least as regards to motion sense) is still widely held (e.g. Wintsch, 1979). Lundgren and Ebblin (1972) concluded that the Honey Hill fault represented a zone of very high strain with minimal displacement formed at the contact between basement and cover during basement diapirism. Mesoscopic and microscopic kinematic indicators (Goldstein, 1982a,b, 1984; Goldstein and Hutton, 1984) define top down (low-angle normal) motions. These data have led to the proposal of two new tectonic models for the formation of the faults: basement-cover decollement developed during basement diapirism (Goldstein, 1982a, b; similar to Lundgren and Ebblin, 1972) and low-angle normal faulting developed during extension (Goldstein, 1984). The timing of fault motion is (if that is possible) even more controversial than the motion direction, although there is a growing consensus that the latest episode of high-temperature mylonitization is of late Paleozoic age (O'Hara and Gromet, 1983; Hermes and Zartman, 1985; Wintsch, 1984; Goldstein, 1984). Goldstein, Rodman and Hutton (in review) suggest at least two episodes of high-temperature displacement based on studies along the Bloody Bluff fault. There, xenoliths of mylonite in a gabbro demand that mylonitization be no younger than Silurian and kinematic indicators describe motion as thrust with a left-lateral component. It is proposed (Goldstein, Rodman and Hutton, in review) that the Lake Char, Honey Hill and Bloody Bluff faults all experienced this pre-Silurian displacement and that the Honey Hill, Lake Char and possibly the Clinton-Newbury but not the Bloody Bluff faults were reactivated as low-angle normal faults in late Paleozoic time as a consequence of localized Alleghanian extension.

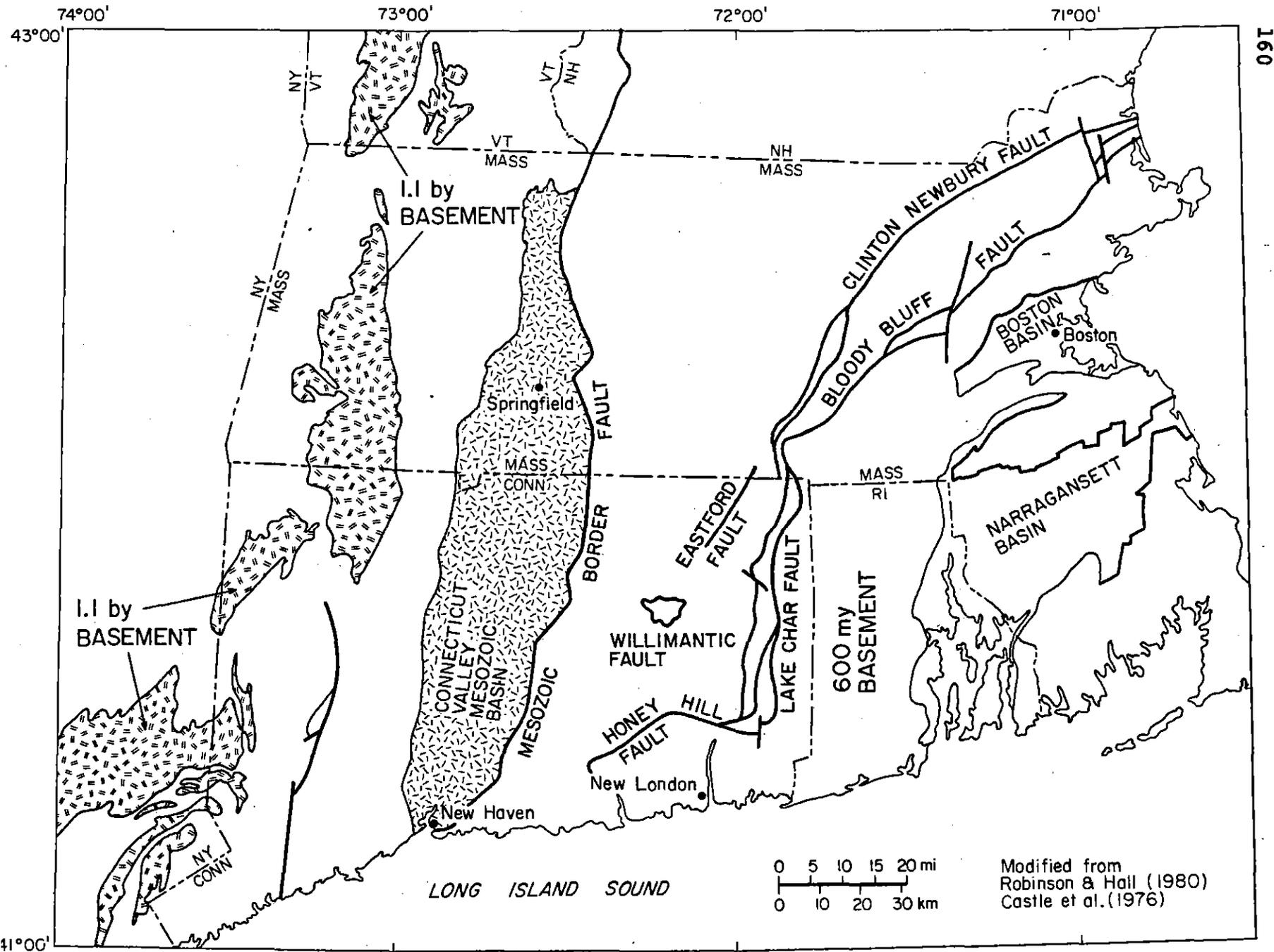


Figure 1. Schematic tectonic map of southern New England showing the location of faults of the southeastern New England fault system.

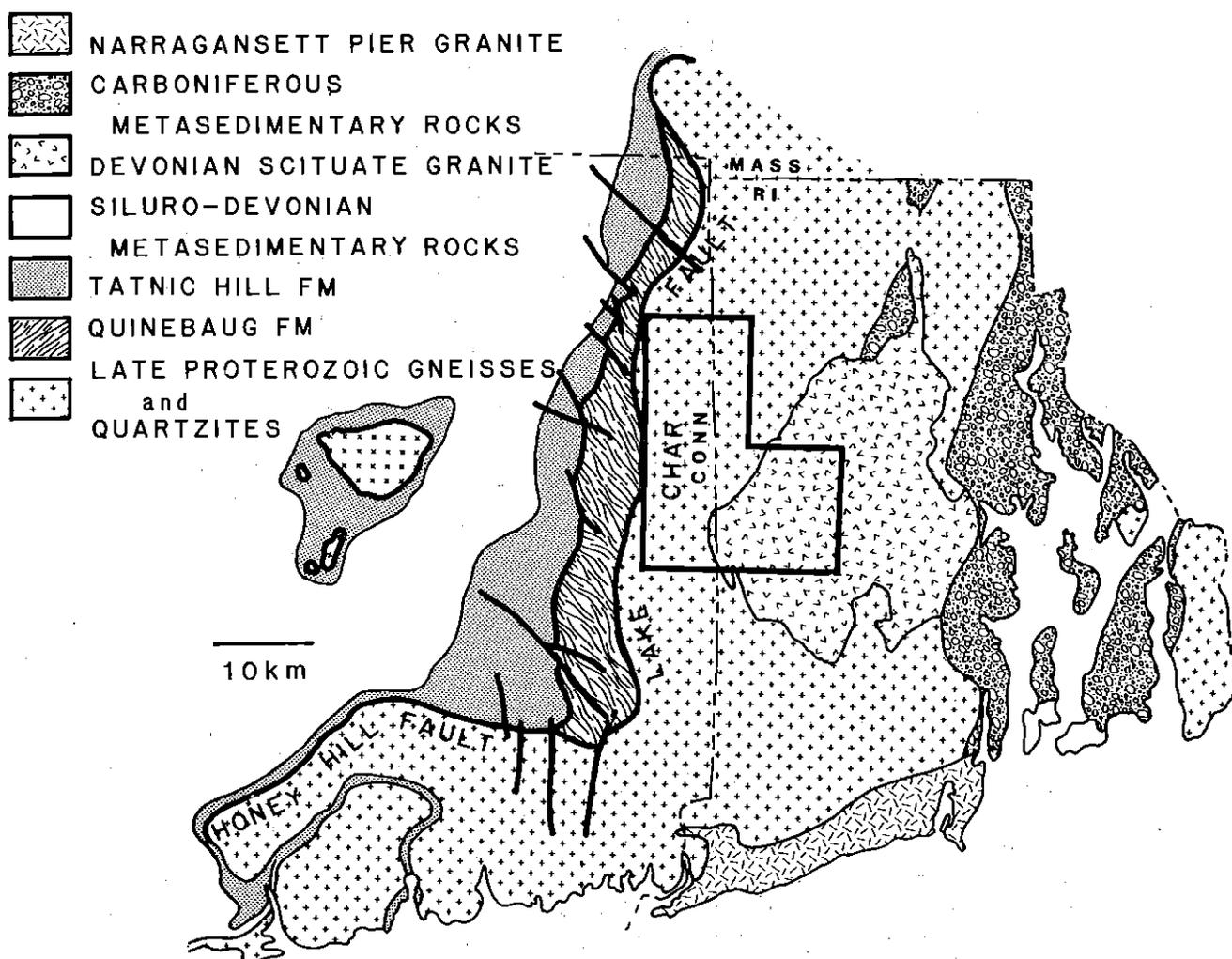


Figure 2. Oversimplified geologic map of Rhode Island and eastern Connecticut modified after Rodgers (1982) and Hermes and Zartman (1985). Unpatterned regions in southern Narragansett Bay are the Cambrian slates of Jamestown Island and Precambrian metamorphic and granitic rocks of Newport. Box is the area covered in Figure 4.

The rock units in and adjacent to the Lake Char-Honey Hill mylonite zone are conveniently divided into upper plate and lower plate although both Dixon (1968) and Snyder (1964) have mapped an upper plate unit in the lower plate. The distinction between upper plate and lower plate becomes confused at the western end of the Honey Hill fault where upper plate stratigraphy is infolded with that of the lower plate (Lundgren, 1962; Dixon and Lundgren, 1968; Goldsmith, 1976). An alternate view of the structure of the western end of the Honey Hill fault is presented by Wintsch (this volume).

#### Upper Plate Stratigraphy

The stratigraphy of the upper plate is dominated by the Putnam Group comprising the Quinebaug and Tatnic Hill formations. Despite the fact that a reasonable stratigraphy has been developed for these units (Dixon, 1964) and used successfully to map in the upper plate, most of the units are lithologically diverse. For example, the lower member of the Tatnic Hill formation contains a rusty-weathering gneiss, perhaps the most distinctive unit in the stratigraphy. However, within that unit are amphibolites which are easily confused with another Tatnic Hill unit, the biotite gneiss, and within those amphibolites are blocks of talc and actinolite (large enough to have been mined for talc) and layers of marble. With these kinds of variables in mind, the stratigraphy of Dixon (1964, 1976) is presented in figure 3. Only the most dominant lithology is noted in this figure and the reader is referred to Dixon (1964) for a more detailed discussion of lithologic variations.

#### Lower Plate

The stratigraphy of the lower plate is dominated by meta-granitic rocks with subsidiary metavolcanic and metasedimentary rocks. Many of the relationships between lower plate units are uncertain and correlations are difficult. Goldsmith (1976) describes the stratigraphy of the lower plate in the New London area (below the eastern Honey Hill fault) as having the Plainfield formation as its lowest member. The Plainfield formation is composed of massive, pure quartzites with a middle unit of (garnet) muscovite-biotite-plagioclase-quartz schist. The Plainfield formation is recognized in easternmost Connecticut (below the Lake Char fault) but much of the stratigraphy above it in the New London area is not recognized further north. The Mamacoke formation, the Monson gneiss and the New London gneiss comprises the upper sequence which consists generally of mafic to intermediate and felsic metavolcanic rocks (Goldsmith, 1976). In eastern Connecticut and Rhode Island the lower plate stratigraphy does not contain Goldsmith's (1976) upper sequence. The oldest rocks of that area are inferred to be the metasediments of the Plainfield Formation and those formerly assigned to the Plainfield Formation (Harwood and Goldsmith, 1971; Moore, 1983) or the "Metavolcanic Rocks" (Quinn, 1971), referred to here as the Unassigned Metasediments, which are differentiated from the Plainfield Formation only on the basis of their association with the Ponaganset Gneiss (figure 4). Both are loosely constrained to pre-date the latest Precambrian. An intrusive contact (??) between the Plainfield Formation and the Hope Valley Alaskite (601±5 ma, zircon age, Hermes and Zartman, 1985) is observed at Stop 9. Similar tenuous intrusive relationships between the Unassigned Metasediments and the Ponaganset Gneiss (also late Precambrian,

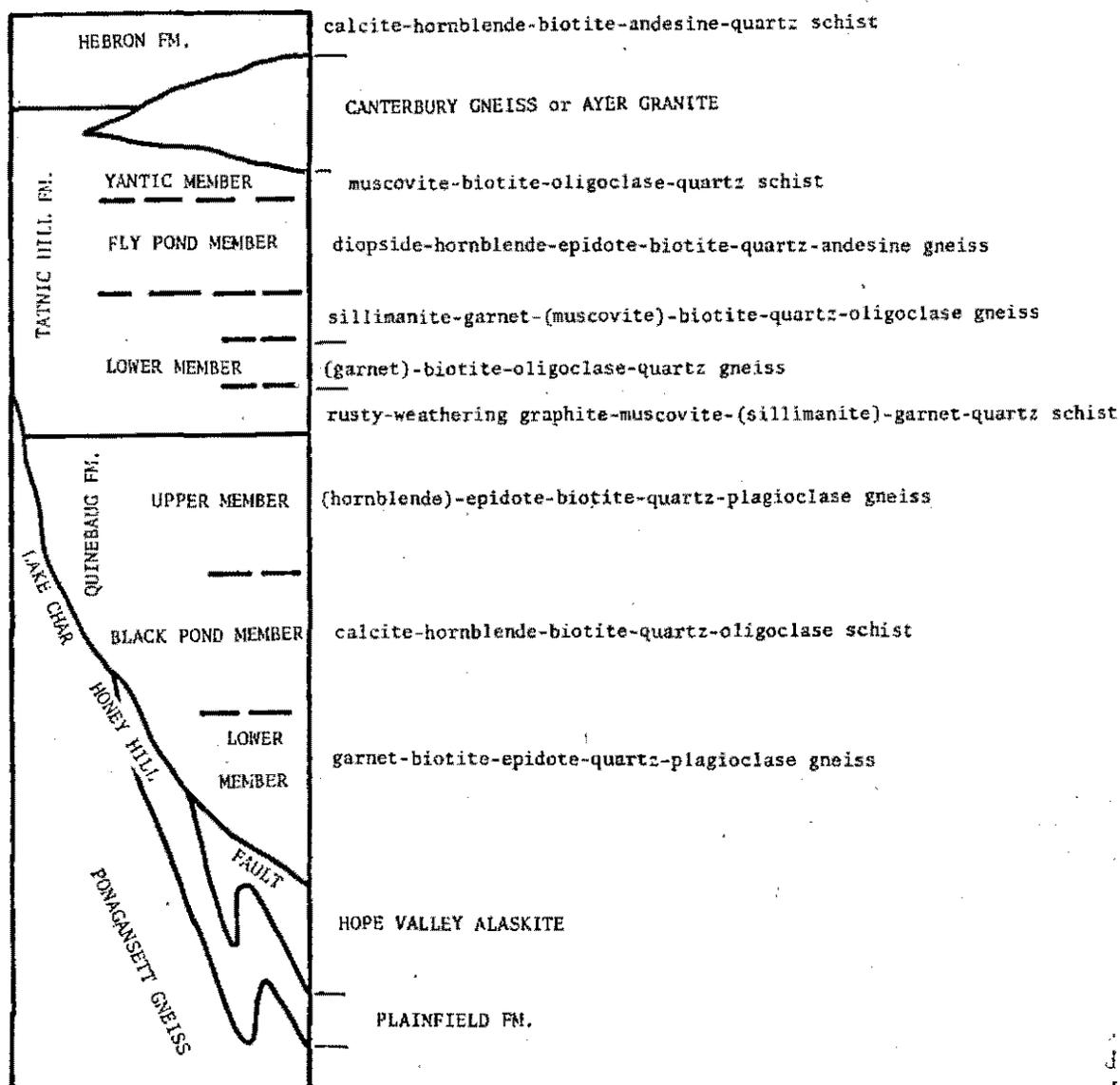


Figure 3. Stratigraphic column and rock descriptions for the eastern Connecticut region modified after Dixon (1964). Only the most dominant lithology is described, see text for further discussion especially of the lower plate.

L.P. Gromet, pers. com., 1985) at Ledge Road and Carbuncle Pond (figure 4) support a Precambrian age for the Unassigned Metasediments. Staurolite and kyanite have been found in the undivided metasediments at Carbuncle Pond (G. Moore, pers. com., 1984; Pope, 1976).

The Plainfield Formation is divided into eastern and western massive, metaquartzarenite members (Stop 7) and an intervening central pelitic member (Stop 8). The metaquartzarenites are 90 - 97% equigranular, recrystallized, fine to medium grained quartz and 3-10% muscovite, plagioclase, biotite, chlorite, and actinolite. The pelitic member is composed of thinly layered schists variably rich in biotite, muscovite, chlorite, epidote, and actinolite, and quartz rich psammities. Other minerals observed in the schists include garnet, sodic plagioclase, calcite, pyrite, and hornblende. Isolated in the schists are boudinaged quartzite layers and quartz veins and bodies of up to 90% actinolite with subsidiary plagioclase and quartz. Transitions between members are gradational, one of which is well exposed at Stop 8 where over a 100 m interval a muscovite rich schist grades, through a psammitic layer, into a sequence of interbedded quartzites and biotite schists ranging in thickness from 2 cm to 15 m where quartzite layer thickness increases generally away from the pelitic member.

Two similar granite gneisses are Late Precambrian in age: The Hope Valley Alaskite and the Ponaganset Gneiss. Regionally, the Ponaganset Gneiss (figure 4) occurs as 1) an andesine phenocryst-bearing porphyritic gray gneiss (with subordinate larger microcline phenocrysts) which ranges in composition from tonalite to granite (Acker, 1950; Frost, 1950; Moore, 1963; Quinn, 1967; 1971), 2) a microcline phenocryst-bearing gray porphyritic granite (Feininger, 1965; Harwood and Goldsmith, 1971; Dixon, 1974; Moore, 1983), and 3) a leucocratic, locally porphyritic pink granite (Frost, 1950; Quinn, 1967; Moore, 1983). At Stop 5 the dominant variety is a medium gray, porphyritic granite gneiss. The 2-4 cm pink phenocrysts (porphyroclasts) are microcline and orthoclase patch perthites (often cored with subhedral, 2 mm plagioclase inclusions) recrystallized in varying degrees to plagioclase free aggregates of 0.5 - 5 mm microcline grains with sutured grain boundaries and broad deformation bands. The matrix is a polygonal aggregate of fine, equant, low strain, quartz, plagioclase, and microcline grains with abundant clots of biotite and accessory phases including magnetite/ilmenite, sphene, allanite, apatite, and zircon. Dark green hornblende can be an important phase locally. Secondary minerals include very abundant epidote and clinozoisite, chlorite, and rare muscovite and garnet. This variety includes 0.1-1 m xenoliths (autoliths?) of medium grained, dark gray tonalite-granodiorite with rare 2 cm plagioclase phenocrysts. Both are cut by a porphyritic leucocratic granite which has, (although depleted in biotite, epidote, and clinozoisite) similar petrography to the porphyritic gray granite. This is interpreted as a leucocratic facies of the Ponaganset Gneiss. Arguments that support a common path of evolution for the tonalite, the porphyritic gray granite, and the porphyritic leucocratic granite include 1) a mutual linear covariance of modal and geochemical data, 2) similar accessory mineral assemblages, 3) similar phenocrysts in the porphyritic gray and the leucocratic granites, and 4) the cross cutting relationships seen in outcrop. These are tentatively termed the andesine, microcline, and leucocratic varieties of the Ponaganset Gneiss.

The Hope Valley Alaskite is also late Precambrian and is a very leucocratic, medium grained, rarely porphyritic, pink granite (Moore, 1963;

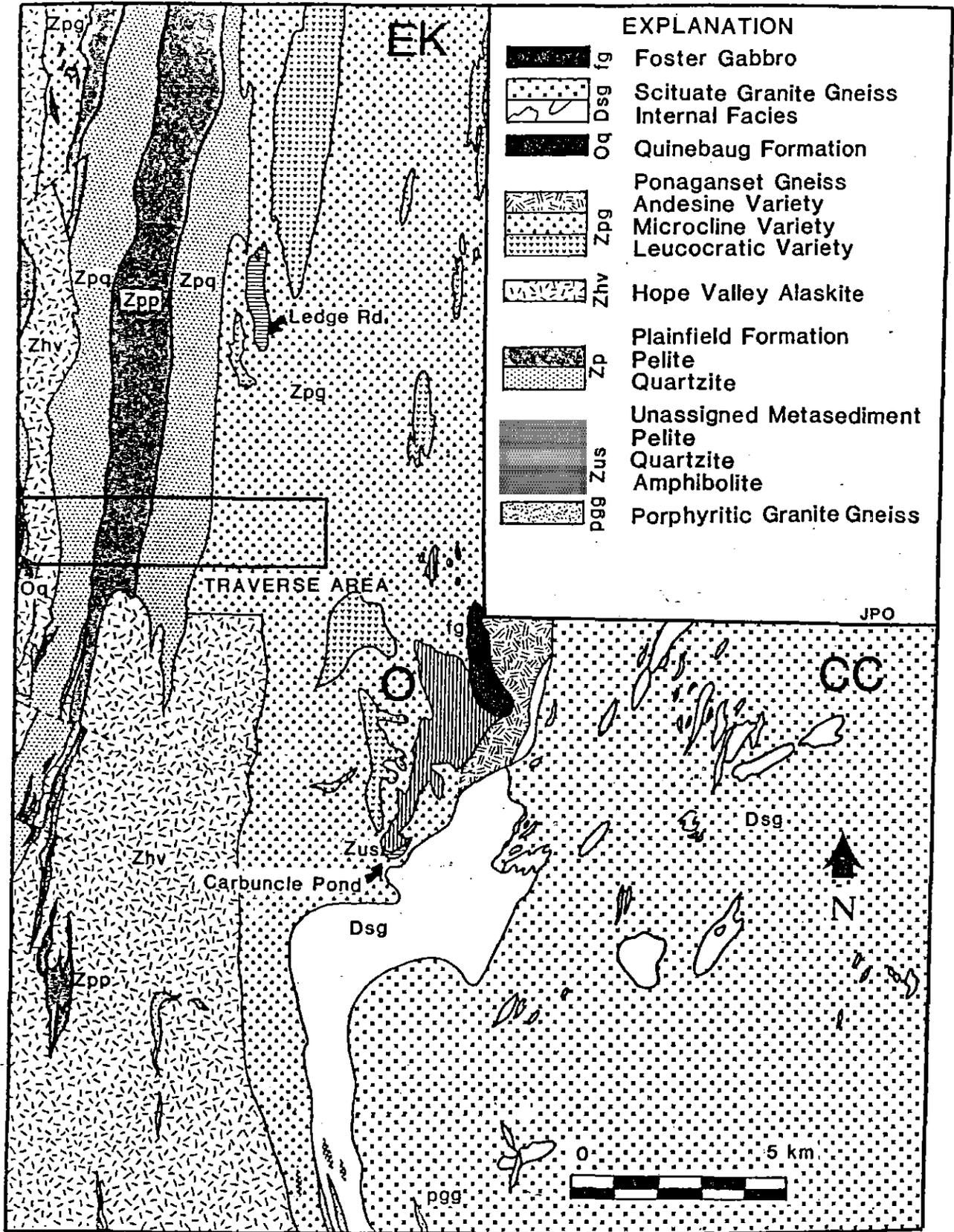


Figure 4. Geologic map of the East Killingly, Oneco and Coventry Center quadrangles modified after Moore (1963, 1983) and Harwood and Goldsmith (1971). Area covered by this map is shown in figure 1. Area shown in figure 11 is noted by the rectangle labeled traverse area.

1983; Feininger, 1965; Harwood and Goldsmith, 1971; Day et al., 1980). It commonly contains 5-10 mm microcline and orthoclase vein perthites and slightly subordinate, well twinned subhedral plagioclase grains. Commonly these are partially recrystallized and often show grain boundary sub-grain development. Many are also fractured where complex networks of very thin fractures are developed in the plagioclase and the perthites are displaced along isolated, wider, matrix filled fractures. Quartz occurs as very elongate ribbons of rectangular undulatory grains, and in a fine, polygonal matrix of quartz, plagioclase, and microcline. Accessory minerals include biotite, muscovite, magnetite/ilmenite, zircon, allanite and very rare garnet. Locally layers up to 10 cm thick are devoid of k-feldspar and are composed principally of very fine-grained quartz. This may be a result of deformation-induced diffusion.

### Structural Geology

The primary structural problem concerns the sense of motion(s) on the Lake Char-Honey Hill fault and its (their) timing. A related problem concerns the structural sequence of rocks above and below the mylonite zone. Goldstein (1982 a,b), based on detailed structural analyses in a small area, concluded that prior to mylonitization the Tatnic Hill and Quinebaug formations were isoclinally folded twice. Peak metamorphic conditions were in the upper amphibolite (sill-ksp) facies during the first folding and were not seriously retrograded during the second isoclinal folding. The younger Hebron formation was only isoclinally folded once. Both rock sequences experienced local post mylonitization folding. Rocks of the lower plate appear to have experienced only one phase of pre-mylonitization isoclinal folding. One of us (J.O.) speculates that this isoclinal folding could be due to early thrust motions on the Lake Char fault which may also have resulted in high-over-low metamorphic zonations. It appears that this isoclinal folding is a regional event as it is commonly observed in the Rhode Island basement (lower plate) (Harwood and Goldsmith, 1971; Skehan and Murray, 1980; Barosh and Hermes, 1981; Skehan, 1983; Barosh, 1984). Correlation of deformations across even moderate distances is a matter dominantly of speculation and more work is required before the isoclinal folding observed in various areas can be correlated in time. The foliation which is axial-planar to isoclinal folds below the Lake Char-Honey Hill fault zone was reactivated as a slip surface during later mylonitization.

As noted above, one of the primary features of the faults is the presence of a wide zone of mylonitization. Despite considerable progress in our understanding of mylonites and the processes by which they form, there is still confusion regarding terminology (for example see Wise et al., 1984, 1985 and Mawer, 1985). Mylonites are fine-grained, well-foliated metamorphic rocks which had as their precursor a coarser-grained equivalent. A clear distinction must be made between mylonites, involving grain size reduction dominated by crystal-plastic mechanisms, and cataclasites, dominated by brittle fracturing. A more thorough discussion of deformation mechanisms and their significance is given below.

Within the mylonites of the Lake Char-Honey Hill fault zone a strong mineral elongation lineation marks the direction of motion. Figure 5 shows mean vector orientations of this lineation from subareas along the fault zone. The lineation is composed, most commonly, of quartz rods and ribbons, elongate biotite flakes and streaks and (less commonly) feldspar rods.

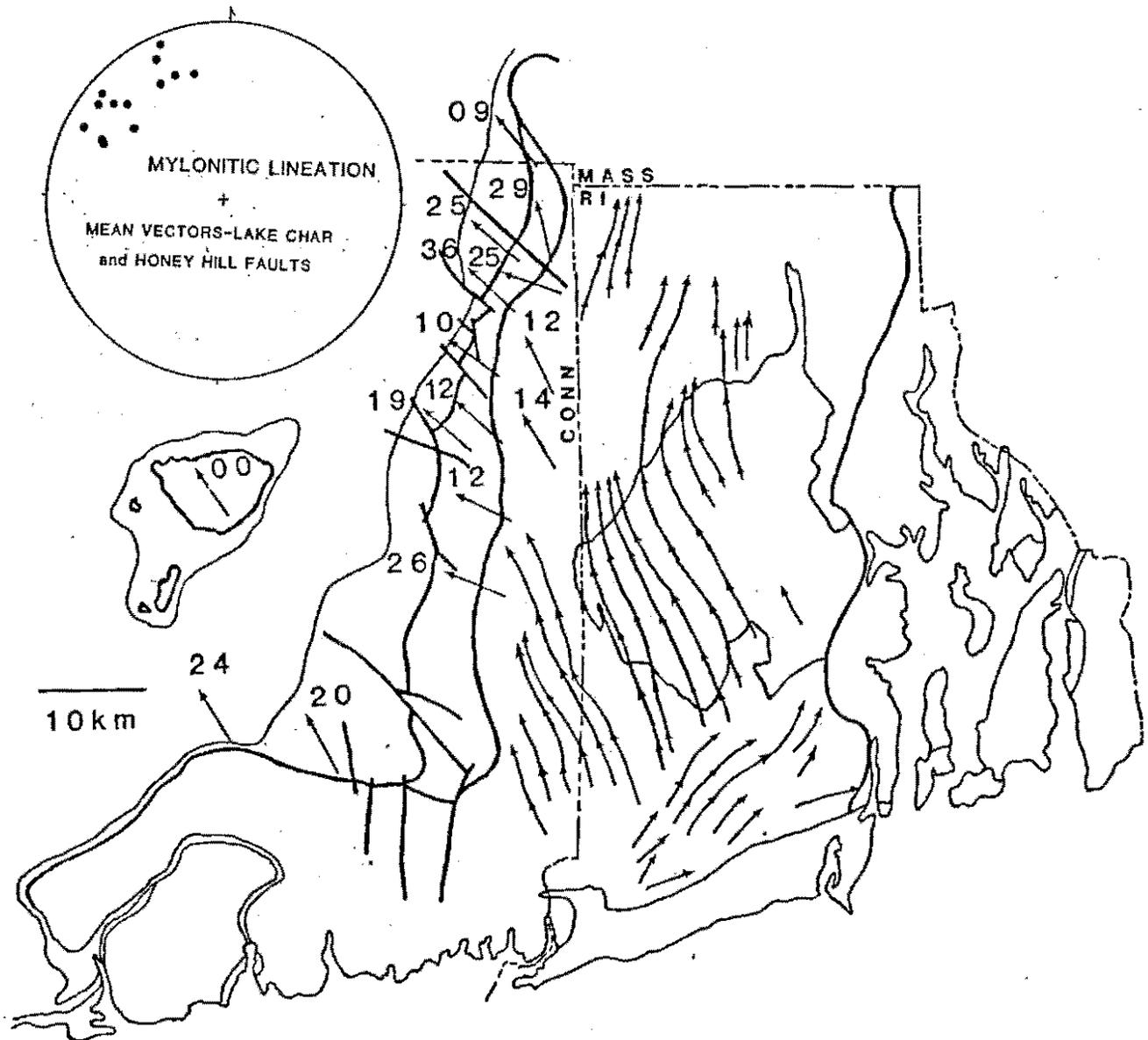


Figure 5. Lination summary of the Lake Char-Honey Hill fault zones. Arrows with labeled numbers represent vector means of field data from one quarter areas of  $7\frac{1}{2}'$  quadrangles. The lower hemisphere equal-area diagram shows those vector mean orientations. Data for the Willimantic Dome (labeled 00) are taken from Wintsch (1982). Shown in the long, unlabeled arrows are the lination trends of Day et al. (1980). See text for further discussion.

Despite small angular differences in the orientation of lineations, the pattern is one of a consistently northwestward trending motion direction. Also shown on Figure 5 is a summary of lineation trends from Day et al. (1980) for the area east of the Lake Char fault. O'Hara and Gromet (1984) speculated that the northwest trending lineations of the Lake Char fault merge smoothly with north-south lineations. They interpret this as relating to a major, north-south striking shear zone. Within this shear zone (Stop 5) the lineation plunges shallowly northward and foliation is almost totally absent. Right-lateral shearing along the lineation has been documented (O'Hara and Gromet, 1984, 1985). Toward the Lake Char fault, foliation becomes more dominant. The transition from north-south lineation orientations to northwest orientations is abrupt (Stop 8) and the relative timing of the two lineations is unclear. There is some reason to believe that Lake Char movements may have outlasted movements to the east in O'Hara and Gromet's (1984, 1985) Hope Valley shear zone. Microstructures associated with the north-south lineations are exclusively ductile for both quartz and feldspar whereas microstructures within the Lake Char mylonites (northwest lineations) span the range from ductile to brittle for quartz and feldspar and ultramylonite layers and dikelets occur locally (Stop 10, figure 7F). Currently, we do not believe that the Lake Char fault forms the less sheared western boundary of a more fundamental shear zone located to the east. Post-mylonitization folding is locally present. This is seen as mesoscopic folds of mylonites (Stop 3) and as a north-south spread in poles to mylonitic layering and foliation.

#### Kinematic Indicators

Much of the current controversy surrounding the Lake Char-Honey Hill fault regards the sense of motion. Goldstein (1982a,b; 1984) and Goldstein and Hutton (1984) have proposed top-down (low-angle normal) motion which stands in marked contrast to the traditional view of thrust motion. One of the main purposes of this trip is to examine mesoscopic structures which bear on fault motion and to present microstructural data of similar significance at the outcrop. Participants and followers of this field guide are urged to collect oriented samples and cut oriented thin sections so that they can verify for themselves the microstructural asymmetry which defines fault motion. Recently, a number of papers have been published which discuss mesoscopic and microscopic structures in shear zones and their interpretation (Simpson and Schmid, 1984; Lister and Snoke, 1985). These papers should help those interested in confirming (or denying) the results presented here. A summary diagram of mesoscopic and microscopic structural asymmetries is shown in figure 6 and a brief description of the more common asymmetric microstructures follows.

Composite Planar Fabrics (C and S) - First described by Berthe et al. (1979), these non-parallel planar fabrics are considered to be one of the most reliable shear sense indicators. In the "traditional" sense, C (cisaillement) planes are parallel to shear zone boundaries and represent zones of high shear strain; S (schistosite) is a grain shape foliation related (perhaps) to the accumulation of finite strain. Thus, in the initial stages of shear, the angle between C and S should be approximately 45° and should decrease as shear strain accumulates. There is some debate over the timing of formation of C planes relative to S. This, however, is not significant to shear sense determination. It need only be established that S curves into C, the sense of obliquity will define the shear sense.

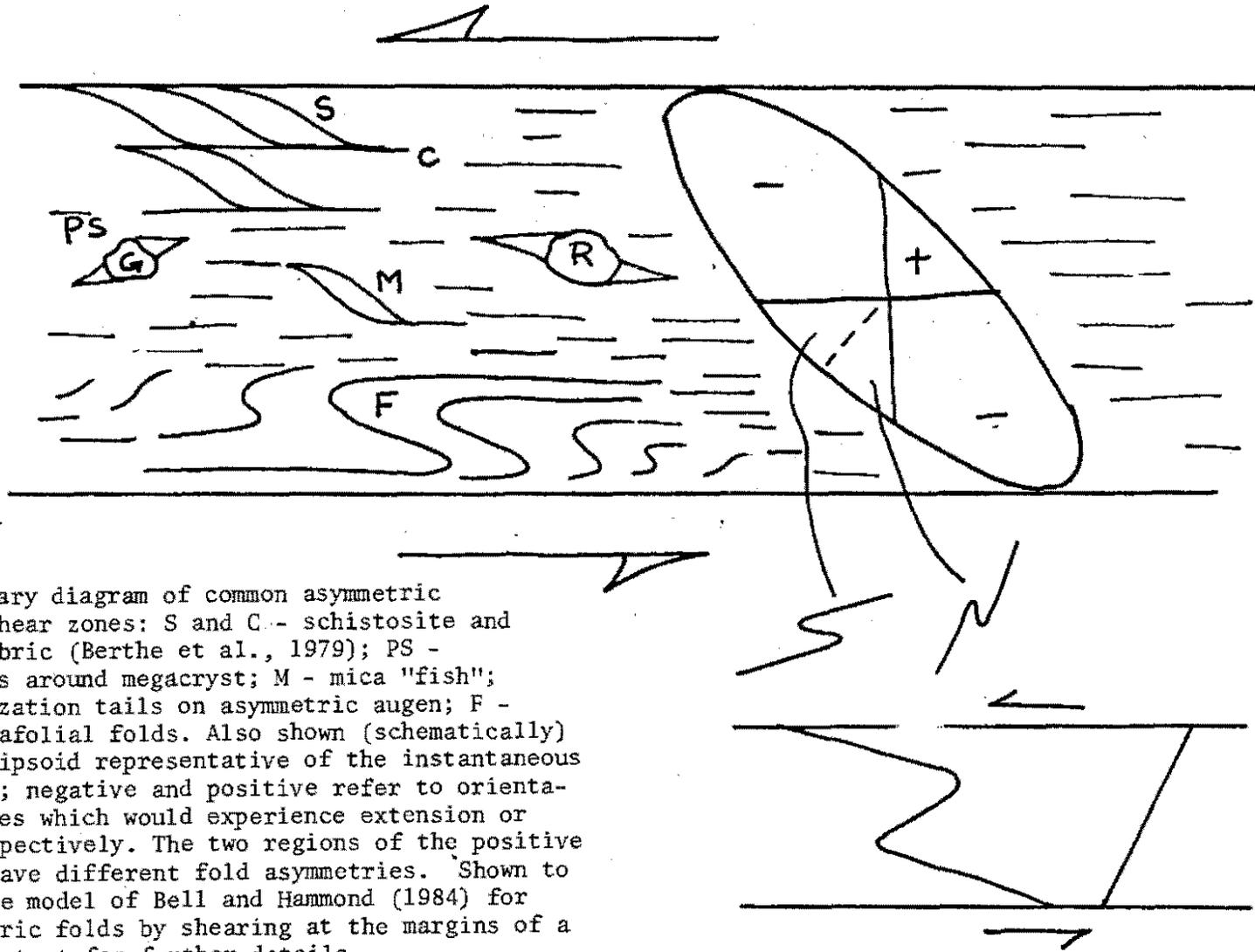
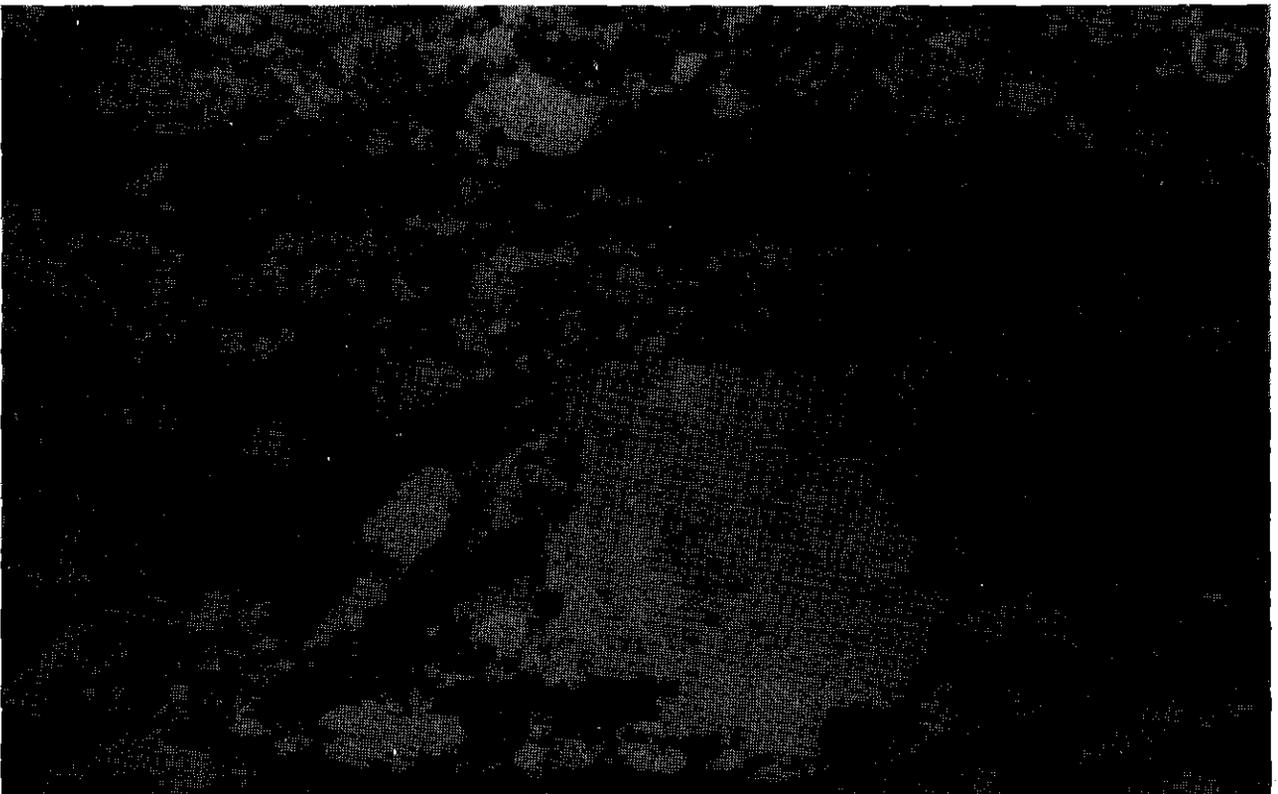
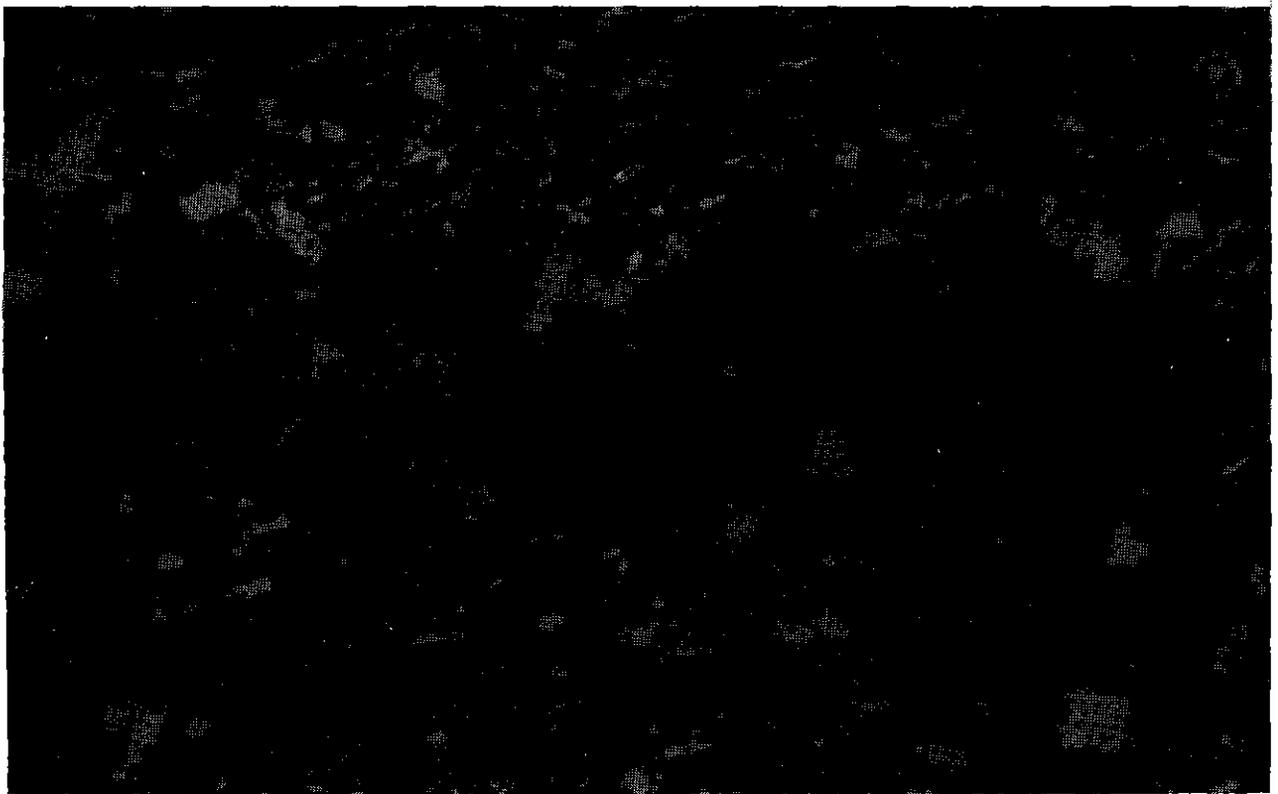


Figure 6. Summary diagram of common asymmetric structures in shear zones: S and C - schistosity and cisaillement fabric (Berthe et al., 1979); PS - pressure shadows around megacryst; M - mica "fish"; R - recrystallization tails on asymmetric augen; F - asymmetric intrafolial folds. Also shown (schematically) is a strain ellipsoid representative of the instantaneous state of strain; negative and positive refer to orientations of features which would experience extension or compression respectively. The two regions of the positive segment would have different fold asymmetries. Shown to the right is the model of Bell and Hammond (1984) for forming asymmetric folds by shearing at the margins of a shear zone. See text for further details.

Figure 7. Typical microstructures in mylonites from the Lake Char-Honey Hill fault. All photomicrographs are from thin-sections cut parallel to lineation and perpendicular to foliation with the down-plunge direction of lineation on the right. A - S and C defined by biotite flakes in the Plainfield Formation; Thompson quadrangle, intersection of Tucker Hill Road and Route 44 (Stop 11, this trip), photograph is approximately 1 mm across; B - Large muscovite "fish" in mylonitized Biotite Gneiss of the Lower Member of the Tatnic Hill Formation, Fitchville quadrangle, 150 m south of Stop 2 (this trip), photograph is approximately 2 mm across; C - Oblique quartz grain shape from same locality as A, photograph is approximately 2 mm across; D - Oblique quartz grain shape, subgrain development and asymmetric recrystallization tails in plagioclase augen in mylonitized Biotite Gneiss of the Lower Member of the Tatnic Hill Formation, intersection of Bishop Road and South Road, Fitchville quadrangle (Stop 2, this trip), photograph is approximately 2 mm across; E - Microfaults in plagioclase megacryst from mylonitized Quinebaug Formation, north of Squaw Rock Road, southeastern Danielson quadrangle, photograph is approximately 2 mm across; F - Ultramylonite layer and dikelets in mylonitized Fly Pond Member of the Tatnic Hill Formation, Fitchville quadrangle, 500 meters west of Stop 2 (this trip), photograph is approximately 2 mm across.







This fabric is locally well-developed in the Lake Char-Honey Hill mylonites. It is best seen in pegmatites within the Fly Pond Member of the Tatnic Hill Formation in the Fitchville quadrangle (Stop 1), in the Hope Valley Alaskite in the East Killingly quadrangle (Stop 9) (only obvious on cut and stained slabs) and in thin sections of the Plainfield Fm. from the Thompson quadrangle (Stop 11, figure 7A).

Lister and Snoke (1985) have suggested that mylonites with oblique, composite fabrics should be considered as a separate broad class of mylonites, S-C mylonites, of which there are two types. Type I is characterized by C and S in the sense of Berthe et al. (1979) (described above) and are common in deformed plutonic rocks, as in the Hope Valley Alaskite (Stop 9). Type II S-C mylonites are defined by other composite planar fabrics, mica "fish" and oblique quartz grain shapes, and are common in quartz- and mica-rich rocks (Lister and Snoke, 1985). These fabrics are exceedingly common in mylonites of the Lake Char-Honey Hill faults.

Mica "Fish" - Muscovite in mylonites is frequently quite large despite the small grain size of most phases, including biotite. These large muscovite grains will commonly take on a sigmoidal shape or form "stairs" connecting C-planes. They have been referred to as mica "fish" and are another reliable, easily interpreted kinematic indicator. Lister and Snoke (1985) suggest that very large muscovite grains are boudinaged or cut by listric normal microfaults when they are inclined to the shear zone boundary such that they will experience instantaneous extension (figure 6). The boudin or fault "blocks" are eventually separated along C-surfaces, their tails experience intense dynamic recrystallization and merge sigmoidally into the C-planes. The sense of obliquity and sigmoidal rotation are excellent shear sense indicators. These "fish" are exceedingly common in the metasedimentary rocks of the upper plate (figure 7B, Stops 1 and 2) and, locally, in the Hope Valley Alaskite (Stop 4).

Oblique Quartz Grain Shapes - Because quartz so easily experiences dynamic recovery and recrystallization, it is able to develop a grain shape which is directly related to the sense of shear. Syntectonic recrystallization results in equant quartz grains; these grains will then act as small strain indicators, elongating initially at about  $45^\circ$  to the shear zone boundaries (C-planes) and then at progressively lower and lower angles until further recrystallization produces a new equant fabric. Further shear strain then acts in a similar fashion on this equant fabric. Because micro-scale domains may undergo dynamic recrystallization at different stages of deformation, one thin section might have quartz grain shapes at angles of  $40^\circ$  to  $10^\circ$  to C-planes (Lister and Snoke, 1985). Oblique quartz grain shapes are common in all quartz-rich rocks of the Lake Char-Honey Hill mylonite zone. They are especially common in the Plainfield quartzite (Stop 11, figure 7d), in the mylonitized Hope Valley Alaskite (Stop 4 and 9, figure 7C) and in the Tatnic Hill formation (figure 7D, Stop 1 and 2).

Quartz C-Axis Orientation - Crystallographic preferred orientation of quartz c-axes developed by progressive simple shear is one of the most widely discussed topics in the structural literature. It is not easily interpreted, although it is a commonly used technique. Among the difficulties relating to the development of crystallographic preferred orientation are the relative dominance of slip on basal versus prismatic

slip planes (Simpson and Schmid, 1984), late stage reorientation of the bulk slip plane due to folding (Carreras et al., 1977), and the component of pure shear involved in the deformation (Wenk et al., 1984). Despite these uncertainties, the C-axis fabric is commonly interpreted with respect to shear sense when either a "skeletal" fabric defining cross-girdles (Lister and Hobbs, 1980) or a point maximum misoriented with respect to the foliation and lineation is observed. The reader is referred to Simpson and Schmid (1984) for a more thorough discussion.

Mesoscopic Asymmetric Intrafolial Folds - Folds in mylonites which have axial planes at low-angles to the mylonitic foliation, deform the mylonitic layering and foliation, have an axial plane foliation which is essentially mylonitic and decrease in amplitude along their axial planes (figure 6) are common in many mylonite zones, including the Lake Char-Honey Hill fault zone. Axes of such folds commonly conform to a separation-angle geometry and have been used to determine the shear sense of the Lake Char-Honey Hill fault (Goldstein, 1982 a,b). The origin of such folds and thus, their interpretation with respect to shear sense is uncertain and is a topic of debate in the literature (e.g. Huddleston, 1983; Platt, 1983 ; Bell and Hammond, 1984). Folds developed in dikes and veins are used with great caution because the rotation sense of the fold as well as the orientation of its axis depends on the initial orientation with respect to the shear zone (Figure 6). Folds developed in mylonitic layering and foliation should be more easily interpreted but their origin is unclear. Layering and foliation in shear zones should roughly parallel the X-Y plane of the bulk strain ellipsoid and because mylonites are the expression of high shear strains, the layering and foliation are sub-parallel to the shear zone boundaries. Planar features, such as mylonitic layering and foliation will deform in accord with their orientation with respect to the instantaneous state of strain in the shear zone and will rotate with respect to the shear zone. Thus, they should always lie within the field of instantaneous finite elongation, cannot rotate through the plane of the shear zone and thus, should not fold. This, of course, assumes homogeneous simple shear which probably never represents reality. There are currently two viable mechanisms to explain the origin of folds in shear zones. Huddleston (1983) argues that folds will only develop when layering is inclined to the shear plane. He explains the origin of folds in glacial ice as the result of such a discordance when a glacier flows over a discontinuity (a hill) as its base. On the down-flow side of the hill, the layering is inclined steeper than the flow planes and folds result which have asymmetries in harmony with the direction of glacial flow (sense-of-shear). Talbot (1979) uses a similar mechanism to explain folds in flowing salt glaciers. Bell and Hammond (1984) also rely on a similar mechanism to explain folds in mylonites. In their model, folds develop on the down-flow side of ellipsoidal pods of less deformed rock. Although they urge great caution in the use of folds as shear-sense indicators, they note that they are potentially useful.

A very different mechanism has been proposed by Platt (1983). Based on a mathematical model involving variations in the rate of shear along layering, Platt (1983) shows that folds with vorticities of both equal and opposite signs to that of the shear (equal and opposite rotation sense) will initiate depending on whether the rate of change of shear on layering is faster or slower than that surrounding it. Platt (1983) further shows that folds which initiate with rotation senses the same as the shear zone will

continue to develop whereas those with opposite senses of rotation will devolve with progressive deformation. We prefer the model of Platt (1983) to explain most of the folds observed in the Lake Char-Honey Hill mylonite zone because the presence of less deformed pods with anastomosing mylonites is only rarely observed and is not present where folds are best developed. Thus, although fold asymmetry is controversial and must be used with caution, we believe that it can be a reliable indicator of shear sense in the Lake Char-Honey Hill mylonites. This has been corroborated at several localities through the observation of more easily interpreted and generally accepted microstructures. In those instances, fold asymmetry agrees with the sense of shear deduced from microstructural asymmetry (Stops 2 and 11).

Other Shear Sense Indicators - A large number of other mesoscopic and microscopic structural features can be used to interpret shear sense. These include asymmetric pressure shadows, asymmetric recrystallization tails on feldspar augen (figure 6), microfaults in feldspars (figure 7E) and almost any other asymmetric fabric. Simpson and Schmid (1984) review the more common of these. One less common asymmetric fabric which is present in some localities visited by this trip is an oblique transposed layering or foliation. It is proposed that this develops in zones of high shear strain by extreme attenuation of short limbs of asymmetric intrafolial folds. The result is a true transposed layering.

#### **Results of Shear Sense and Microstructural Studies**

A wide variety of shear sense indicators have been observed in mylonites of the Lake Char-Honey Hill mylonite zone (figure 7). These include asymmetric intrafolial folds, asymmetric augen, extensional crenulations, sense of rotation at the margins of a small (1 m wide) shear zone, oblique transposed layering, oblique quartz c-axis asymmetry (figure 8), C and S structure, asymmetric mica "fish", oblique quartz grain shape, and a variety of other microstructures including microfaulted feldspar and displaced grains. The vast majority (approximately 95%) of these show that mylonites were formed during motion of the upper plate towards the northwest with respect to the lower plate (top-down or low-angle normal). The body of data is so large and the microstructures are so unambiguous (figure 7) that this conclusion is believed to be an absolute certainty. The temperatures ambient during mylonitization can be approximated by observations of microstructures. Plastic deformation of feldspar is suggestive of amphibolite facies temperatures (approximately greater than 525°C) and the transition from brittle to ductile deformation of quartz is reflective of temperatures in the range of 300-350°C. In most mylonites from the Lake Char-Honey Hill faults, both quartz and feldspar show the effects of dynamic recovery and recrystallization (figure 7D). This suggestion of amphibolite facies temperatures has been locally corroborated (Goldstein, 1982c) with geothermometry and is in agreement with the local occurrence of sillimanite in mylonite, although most mylonites do not contain an aluminosilicate phase. Locally, quartz can be observed deforming by crystal plastic mechanisms and feldspar by brittle mechanisms (figure 7E) suggesting a temperature of formation between 350 and 525°C. In some very narrow zones, even quartz deforms by brittle mechanisms (figure 7F). This leads to the conclusion that motion on the Lake Char-Honey Hill fault began its motion at high temperature and at depths of between 12-17 km. Most mylonites were active during this early phase because all but the most mafic phases were deforming ductily and were, therefore, weak. As temperatures

and depths decreased and feldspars became stronger, deformation was concentrated in more narrow zones and at reasonably cold, near-surface conditions the deformation was localized in very narrow zones. It is difficult to place an absolute time span on this sequence and to know the sense of motion during the most brittle, latest motion. One of us (AG) speculates that the sequence spanned a geologically brief period during the Alleghanian orogeny and represents top-down motion during the entire period.

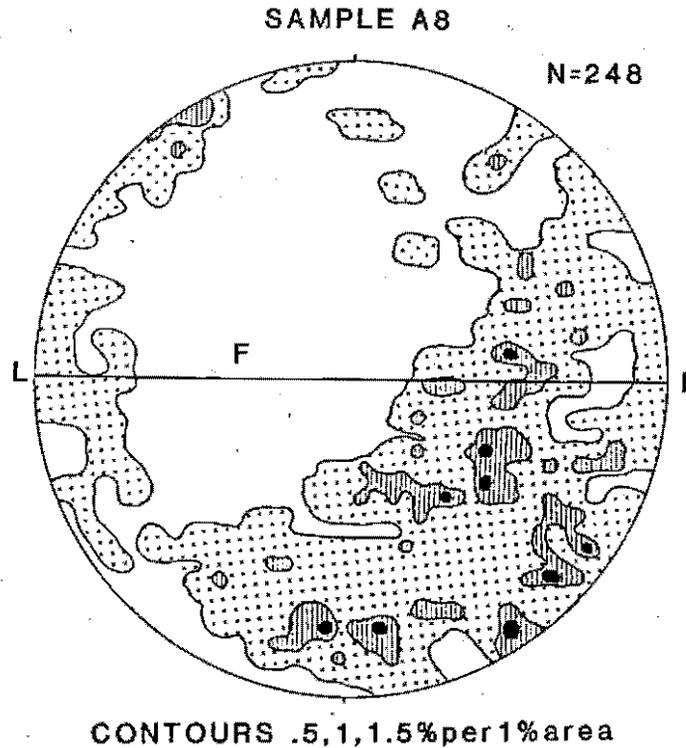


Figure 8. Lower hemisphere, equal-area diagram of quartz c-axes from mylonitized Hope Valley Alaskite in southern Oxford quadrangle (Goldstein's (1982) Stop 1). C-axes were measured from three mutually perpendicular thin-sections and are plotted such that lineation plots as east-west and horizontal with the northwest end of the lineation on the right (viewed looking southwest).

Results of microstructural studies of samples from the Bloody Bluff fault (Goldstein, Rodman and Hutton, in review) contrast with those noted above. All samples of the Bloody Bluff mylonites show thrust motion with a left-lateral component and ductile deformation of feldspars indicates that thrusting was a high-temperature event. Further, xenoliths of mylonite in unmylonitized gabbro require that mylonitization be no younger than Silurian. These conclusions provide a potential explanation for the local occurrence of thrust-motion indicators on the Lake Char-Honey Hill fault. It is likely that early thrusting occurred on the Honey Hill-Lake Char faults as well as the Bloody Bluff. Thrust indicators, then, could be localized in blocks which did not enjoy the later, top-down motion. Alternatively, top-down motion could have been preceded immediately by a thrusting event (S. Mosher, pers. comm., 1984). The later phases of movement, however, were sufficiently intense to obliterate nearly all traces of thrusting. A final alternative is that thrust motions represent nothing more than "antithetic" shear developed around blocks or pods resistant to the mylonitization.

#### ACKNOWLEDGEMENTS

Owens would like to thank his principle advisors O. Don Hermes and Daniel P. Murray for constructive criticism and advice through the duration of this project; L.P. Gromet, R. Goldsmith, W. Danforth, N. Hamidzada, J. Bursnall and B. Jordan for helpful discussions of the often vexing problems of Avalonian geology, and G.E. Moore for considerable assistance in the forms of valuable discussion, and the very generous donation of field notes, samples, and thin sections. Owens was supported in part by a student research grant from the Geological Society of America and by NSF Grant EAR 8206149 (to O.D. Hermes).

Goldstein acknowledges the field assistance of D. MacLean, K. Linton, and S. Rothwell. Preparation of the manuscript was greatly aided by assistance from E. Rodman and was typed by B. DeGroat. Goldstein's work was funded by the Colgate University Research Council, the Research Corporation (Grant C-1754) and the Connecticut Geological and Natural History Survey.

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Wise, D.U., Dunn, D.E., Engelder, J.T., Geiser, P.A., Hatcher, R.D., Kish, S.A., Odom, A.L., and Schamel, S., 1985, Reply on "fault-related rocks: Suggestions for terminology": *Geology*, v. 13, p. 379.

Starting Point: Communter Parking Lot at exit 80, I395, Norwich, Conn. The parking lot is located immediately west of I395 off Route 82 adjacent to the Sheraton Motor Inn entrance. From parking lot, the trip continues west on Route 82. This trip leaves at 8:00 a.m.

Mileage: Begins from right turn onto Route 82.

<u>Cumulative</u>	<u>Interval</u>	
1.6	1.6	Turn right onto South Road at sign for Camp Tadma.

South Road runs approximately parallel to and slightly north of the Honey Hill fault. Outcrops and cliffs on your left (north) are of thoroughly mylonitized Tatnic Hill Fm. (Biotite Gneiss of Lower Member and Fly Pond Member)

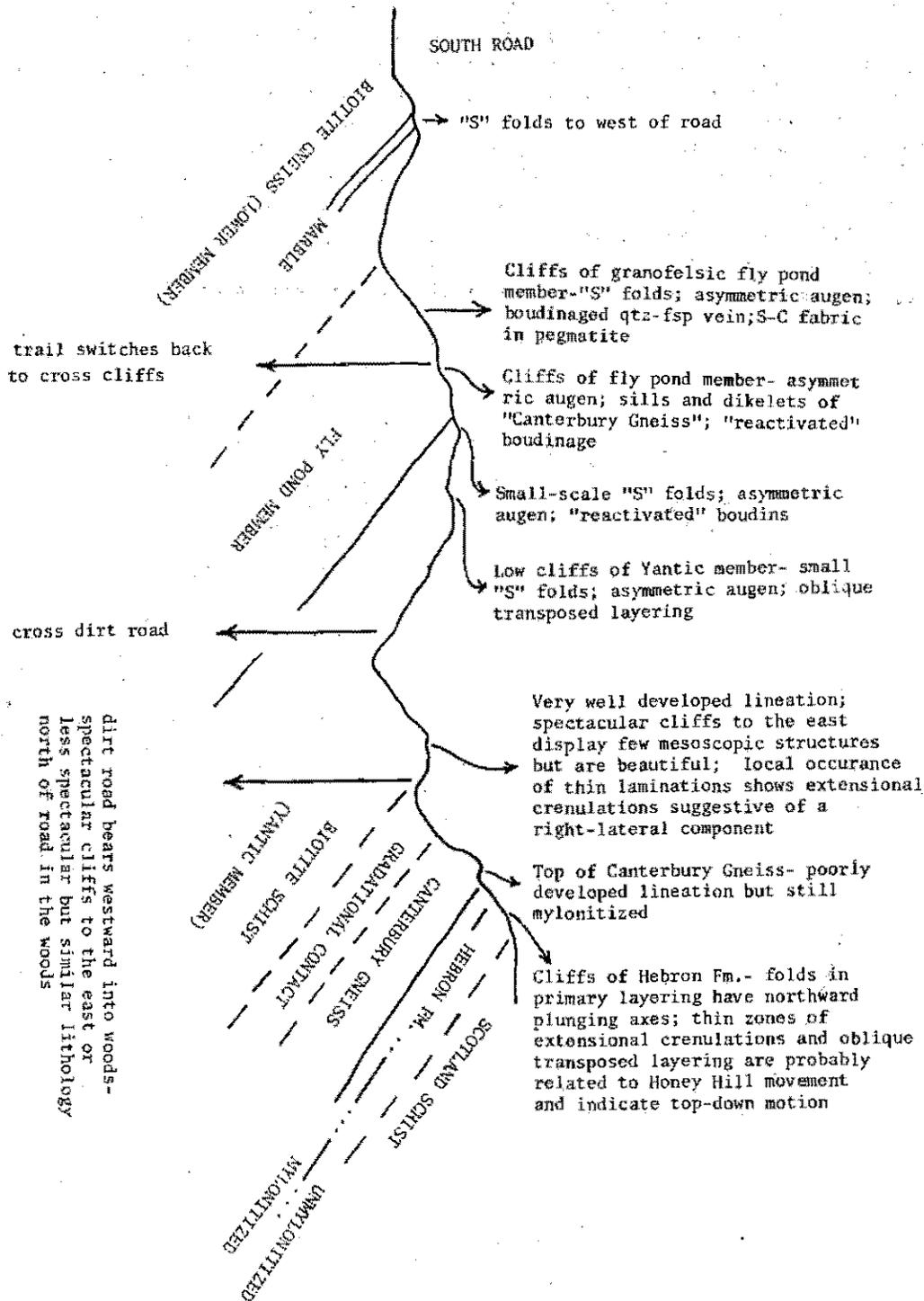
2.5	.9	Intersection with Swan Road and Bishop Road - Continue straight on South Road.
3.35	.85	Stop #1 - Power line crosses South Road; Park along power line road on south side of road.

### STOP #1

Outcrops at parking area are of mylonitized biotite gneiss of the Tatnic Hill Fm. Locally present are folds with axial planes at high angles to the mylonitic foliation which may represent post-mylonitic deformation. Across the small swamp south of South Road are outcrops which Snyder (1964) mapped as plagioclase gneiss south of the Honey Hill fault (Quinebaug Fm.); the trace of the Honey Hill fault occupies the swamp. We will traverse the Honey Hill mylonite zone from close to its base, along the power line road to the north, to its top in the Hebron Fm. Features of interest are noted on the schematic cross section (figure 9). Here the mylonite zone is .7 mi wide, so that this traverse will occupy considerable time, but the mesoscale structures are quite well developed and represent a wide variety of kinematic indicators as well as enigmatic features which can be discussed.

Cliffs immediately south of the power line and north of South Road expose marble of the Fly Pond Member of the Tatnic Hill Fm. Interlayered with the marble are quartzites and these quartzites are folded spectacularly. This is one of the best locations to discuss the origin and significance of folds in shear zones, a topic of considerable debate and uncertainty (see text). Care must be taken at this locality to remember that, because the lineation plunges obliquely

Figure 9. Schematic cross-section of Stop 1 noting mesoscopic structures of interest.



into the outcrop, only a component of fault motion would be parallel to the outcrop face making fold cross sections much less attenuated than if viewed in sections parallel to the lineation. Folds here have a wide variety of geometries: some have axial planes which parallel the mylonitic foliation whereas others have axial planes at a high angle to it. Some are highly attenuated isoclinal similar folds whereas others are smaller amplitude and wavelength buckle folds. It appears as if the quartzite layers were initially boudinaged and then folded. If this boudinage were related to mylonitization, as we believe, then the shear zone boundary would have been inclined at a steeper angle than the layering (with respect to its current orientation). Layering would have rotated toward parallelism with the shear zone boundary and folding could have resulted either from strain heterogeneities localized around boudin or as a result of strain rate variations (Platt, 1983). Fold axis orientation and rotation sense (figure 10) define top-down motion.

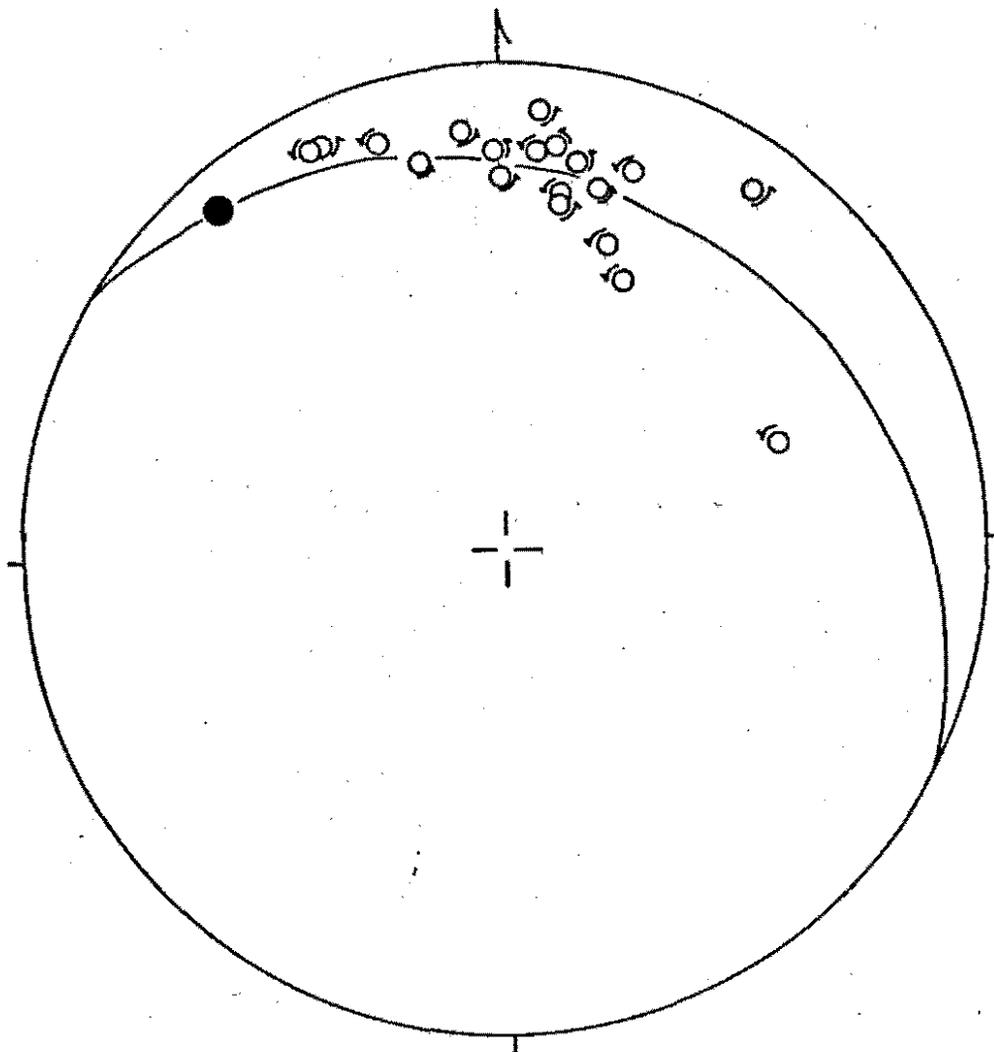


Figure 10. Lower hemisphere, equal-area plot of fold axes measured in the marble of Stop 1. Foliation is shown as the great circle and lineation as the solid dot.

Above the marble is a small pod of biotite gneiss enclosed by marble and calc-silicate of the Fly Pond Member. Within this block are some tenuous thrust-motion indicators including oblique foliations and asymmetric folds. It may be that this block was prohibited from enjoying the top-down motions as a result of its more ductile envelope.

Return to power line and walk northward along power line road. Numerous mesoscopic structures can be seen principally in the traverse over the first large hill; these are noted on figure 9. Continue along road to top of second large hill to observe mylonitized Canterbury Gneiss and spectacular mesoscopic structures in Hebron Fm.

Return to cars. Travel east on South Road.

<u>Cumulative</u>	<u>Interval</u>	
4.2	.85	Intersection with Swan Road and Bishop Road. Park by side of road.

**STOP #2** (Optional)

Natural exposures at the northeastern corner of this intersection are of biotite gneiss of the Tatnic Hill Fm. and are an especially good locality to observe folds in mylonites and to collect samples for microstructures. Oriented thin-sections from this locality display muscovite "fish", oblique quartz grain shapes and asymmetric augen. Asymmetric folds are well developed in quartzofeldspathic layers in outcrops in the woods and immediately north on Bishop Road. The layering may have been pegmatitic dikes, veins or sills. It is interesting to note that fold asymmetry always agrees with microstructural asymmetry.

5.1	.9	Turn left onto Rt. 85
6.9	1.8	Cross under I395
9.0	2.1	Cross bridge over Yantic River into Norwich-Follow signs for Rt. 2E.
9.2	.2	At end of bridge turn right onto Rt. 2E and I2N
9.6	.4	Turn right across bridge-follow signs for Rt. 2E
9.6	.01	Immediately after bridge turn left-Follow signs for Rt. 2E and I2N
9.9	.3	Turn right - Rt. 2E
10.2	.3	Turn left. Follow signs for Rt. 165E

<u>Cumulative</u>	<u>Interval</u>	
15.3	5.1	Intersection with Rt. 164- Continue straight on Rt. 165
19.1	3.8	Turn left onto Gravel Road- Burdick Lane
19.6	.5	Pull over into "cleared" area on left.

**STOP #3**

Exposures on crest of hill to the left are of mylonitized lower plate and were first shown to the authors by Bobby Dixon and described in her guidebook article (Dixon, 1982; Stop #9). Here mylonitized Hope Valley Alaskite (?) displays an exceptional ribbon lineation and post-mylonitic folding. These post-mylonitic folds are uncommon but are present at several localities along the Lake Char fault. We speculate that these folds may be related to late stage displacements on the Lake Char fault.

20.1	.5	Return to Rt. 165 - turn left
21.2	1.1	Turn left onto Rt. 201
21.5	.3	Park on Rt. by waterfall

**STOP #4**

This short stop is to examine, briefly, one of the southernmost exposures of mylonite of the Lake Char fault. Here the Hope Valley Alaskite is mylonitic, displays a N50-60<sup>o</sup>W trending lineation and has top-down microscopic kinematic indicators including oblique quartz grain shape and mica-fish. The obliquity of the muscovite fish can be seen mesoscopically by "fish flash".

Continue north on Rt. 201.

23.3	1.8	Intersection with Route 138 (Dixon's, 1982, Stop #10); Turn left onto Rt. 138 westbound.
27.1	3.8	Intersection with I395, turn right onto 395 northbound
40.1	13	Follow signs for Connecticut Turnpike and Route 6; Bear right (I395 bears left)
44.9	4.8	Park in Diner parking lot... lunch stop

Turn around and return westward on Connecticut Turnpike. The next five stops constitute a cross-section from Rhode Island basement terrane structurally upward to the Lake Char fault. The geology of this traverse is illustrated in figure 11 and 12. Structural data for domains along this traverse are shown in figure 13.

<u>Cumulative</u>	<u>Interval</u>	
45.9	1.0	Pull over by large roadcut

#### **STOP #5**

Domain I, Ponaganset Gneiss (fig. 14a). The lithology seen is typical of the western Ponaganset Gneiss. This is a medium gray, porphyritic granite gneiss. The lineation is seen in the N-S, sub-horizontal elongation and rotation of the 2-4 cm, pink phenocrysts (porphyroclasts). Note the absence of foliation to the extent that there is little or no preferential orientation of phenocrysts perpendicular to lineation. Note that the porphyritic gray granite includes xenoliths of the dark gray tonalite - granodiorite and is cut by the porphyritic, leucocratic granite and aplitic dikes.

46.4	.5	Small roadcut on westbound side of highway, park off shoulder of highway.
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#### **STOP #6 (optional)**

Domain I, Ponaganset Gneiss. This outcrop is within 500 m of Plainfield Formation Quartzite and, as presently defined, the Hope Valley Shear Zone. Here the Ponaganset is weakly to moderately foliated (N trending, W dipping, fig. 13) where the microcline phenocrysts (porphyroclasts) are more thoroughly recrystallized, and these, the quartz aggregates, and the biotite clots are less equant perpendicular to the prominent lineation.

47.2	.8	Large roadcut, park off shoulder of highway.
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#### **STOP #7**

Domain II, Plainfield Formation Quartzite (figure 15). This is one of the larger exposures of the massive, equigranular, metaquartzarenite typical of the quartzite members of the Plainfield Formation. Here foliation occurs as 1) a moderate to strong alignment of 0.5 mm muscovite and biotite lathes, and 2) 1-4 mm horizons of coarser, rectangular, recrystallized quartz grains. The strong N trending lineation persists. In the western part of the outcrop two pelitic layers (biotite schist) outline a isoclinal fold hinge (note enveloping surfaces, figure 15). However, within the enveloping surfaces the quartzite pelite layering has been rotated (or transposed)

A6-32

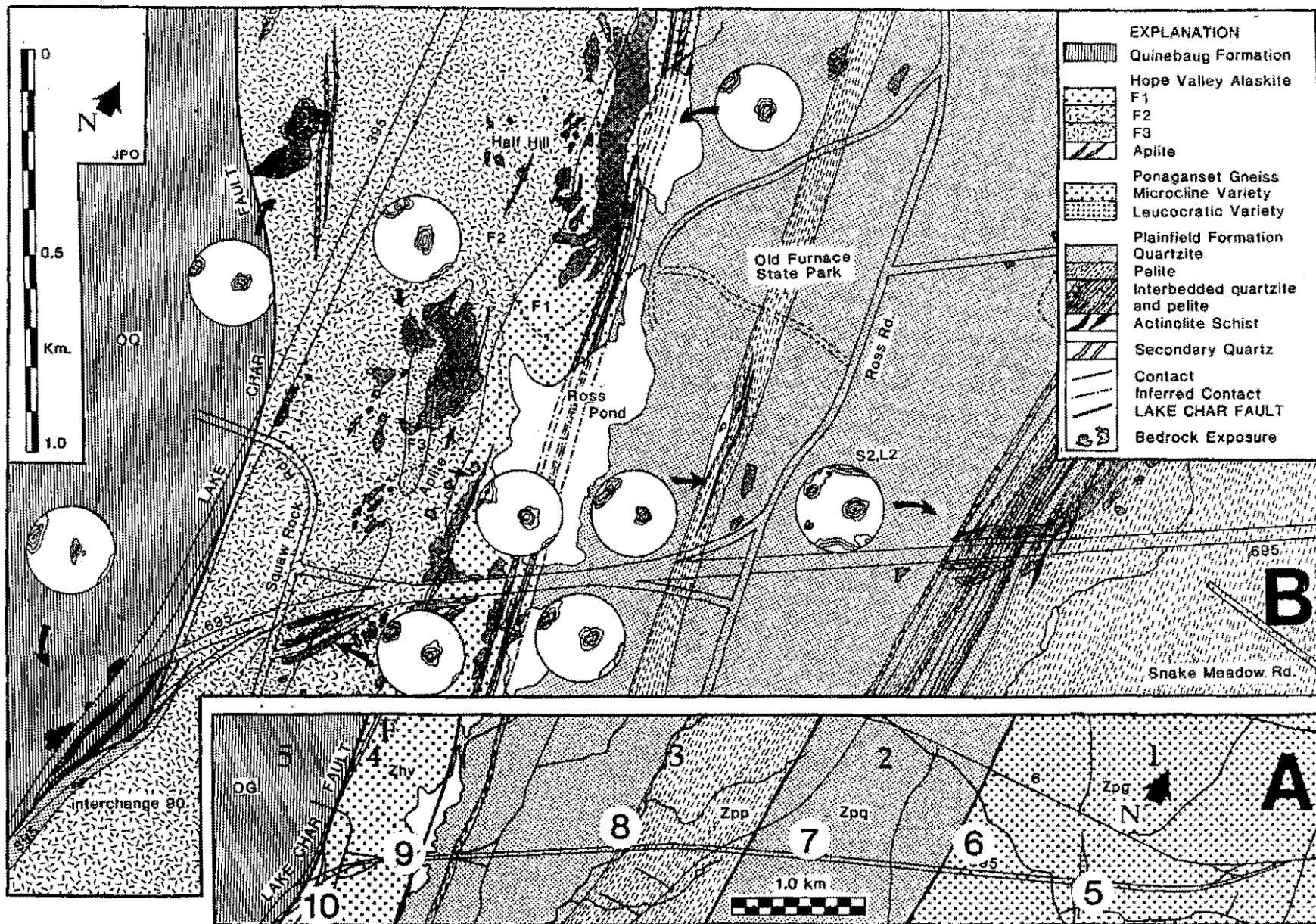


Figure 11. Geologic map of region around Stops 5 through 10, modified after Moore, 1983. Lower hemisphere, equal-area diagrams show the orientation of foliation and lineation, contours are 4.75, 9.5, 14.3, 28.6, and 38.1 percent per one percent area.

A6-33

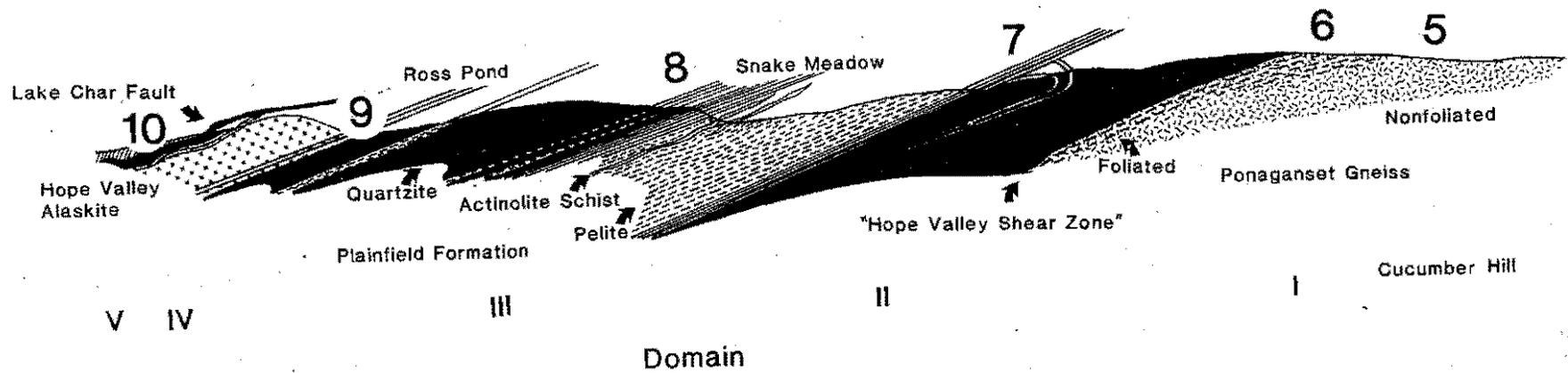


Figure 12. Schematic cross-section of region along the route of Stops 5 through 10.

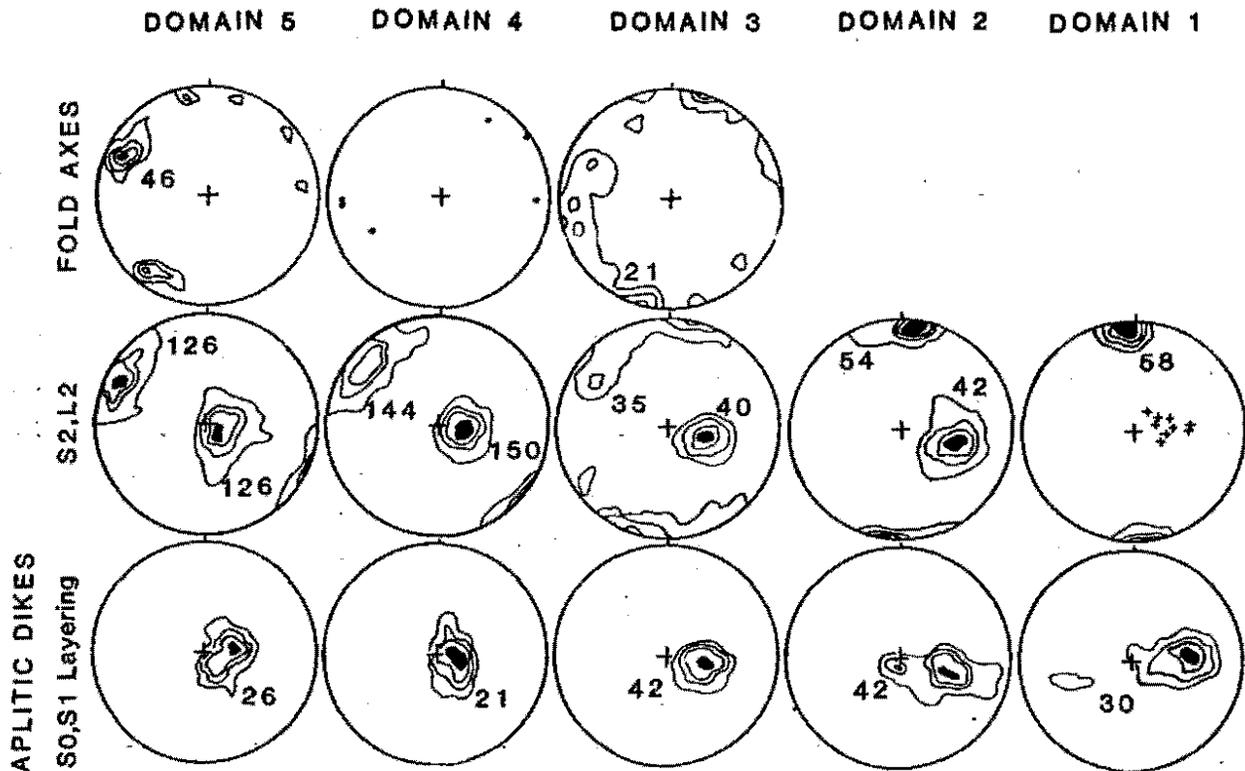
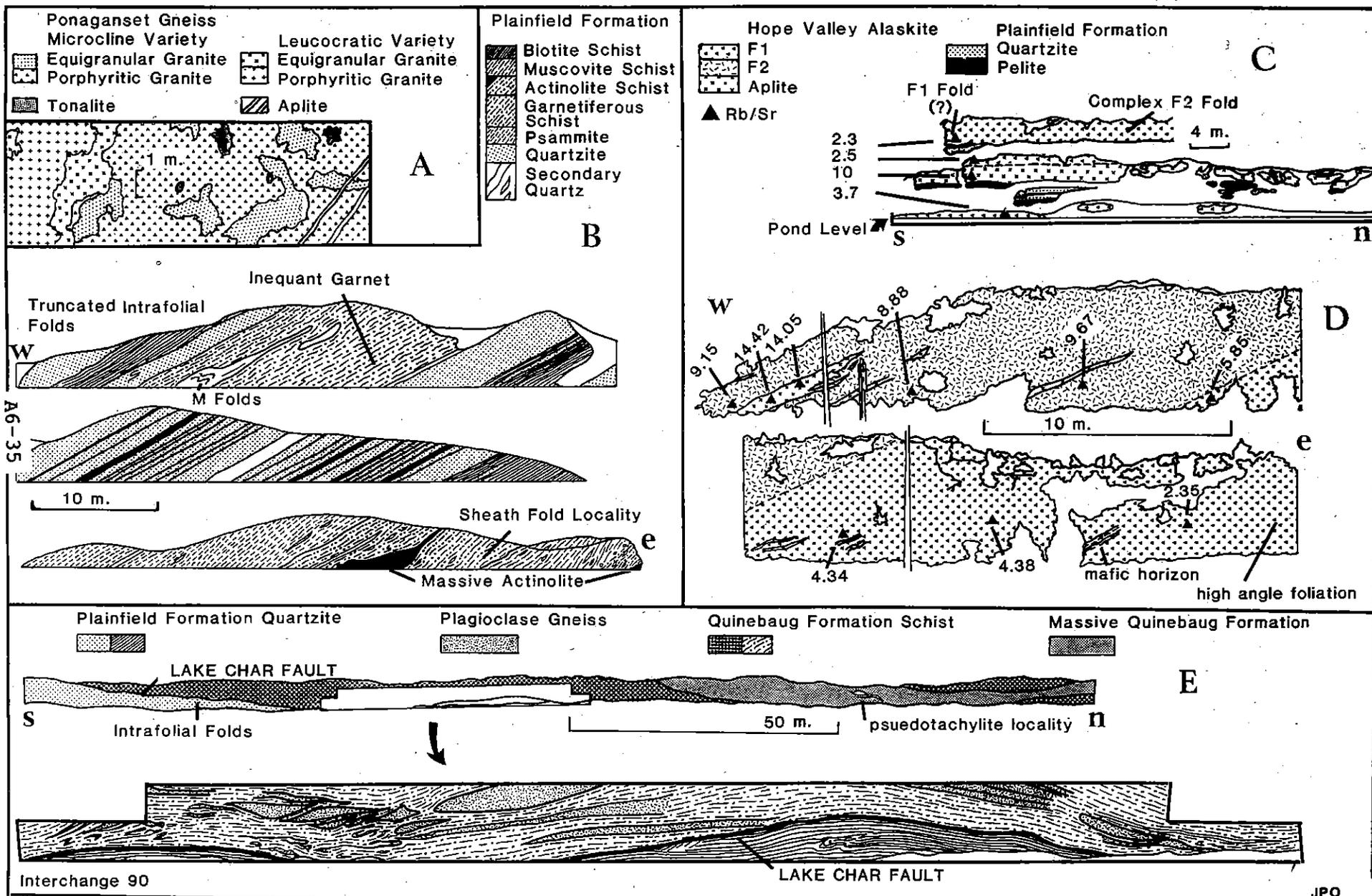


Figure 13. Synopsis of structural data for domains shown in figure 12 plotted or contoured on the Lower hemisphere of equal-area diagrams (S2 and L2 refer to foliation and lineation, respectively), numbers of data are noted, contours are the same as in figure 11.



Figure 14. Sketches of geological relationships at field trip stops along Rt. 695. A - Internal facies of Ponagansett Gneiss, Stop 5; B - Tape and compass map of Plainfield Formation facies at Stop 8, facing north; C - Tape and compass map of interlayered Plainfield Formation and Hope Valley Alaskite at Stop 10, facing west; Internal facies variation in Hope Valley Alaskite at Stop 9, facing north. Note Rb/Sr values increasing to the west and in aplites; E - Tape and compass map of roadcuts at Stop 10 exposing the trace of the Lake Char fault, facing west.



into parallelism with foliation. Note 1) that within the enveloping surfaces, layering is truncated by foliation, indicative of segmentation of layering along foliation; 2) that poles to layering form a partial great circle roughly coincident with poles to foliation (figure 13), and 3) that lineation is crudely axial to this partial great circle.

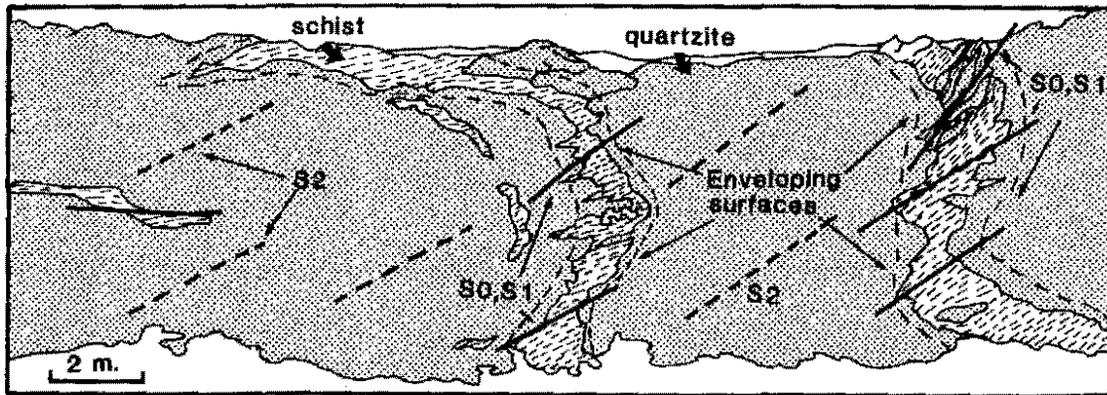


Figure 15. Sketch of roadcut at Stop 7 showing isoclinal fold hinges in bedding of the Plainfield Formation, facing north.

<u>Cumulative</u>	<u>Interval</u>	
48.1	.9	Large roadcut on westbound side of highway

**STOP #8**

Domain III, Plainfield Formation Pelite and Quartzite. In this outcrop some of the Plainfield Formation pelitic member and the gradational transition into a western quartzite member is well exposed (figure 14b). Foliation is well developed in the schists as a sinuous, undulatory surface defined by the preferred orientation of micas and the flattening of quartz and plagioclase, and is developed in the quartzites as at Stop 7. Lineation is well defined in the quartzites as a "mullion" structure and is less well defined in the schists as a preferred orientation of acicular minerals or a tight crenulation of micas in foliation. Asymmetric 1 cm micro-folds in foliation with axes that parallel lineation are quite common. Quartz rods and boudins are common within the schists. All of these structures probably were initiated during the isoclinal folding phase of deformation. An axial planar foliation may have served as a pre-existent slip plane for later shearing. Fold axis parallel crenulations, minor folds, quartz rods, and boudins may have been subsequently rotated into parallelism with the shear-related lineation. Both lineation orientations (N-S, NW) are present in this outcrop and the transition from N-S orientations at the eastern end of the cut to NW at the western end of the cut is apparently abrupt. We have not found the two lineations on the same surface but in the center of the roadcut adjacent foliation surfaces can have lineations with divergent orientations. Also found here are northeast-trending elongation lineations, sheath folds (eastern end of roadcut, figure 14b), truncated intrafolial folds and inequant, flattened garnets.

<u>Cumulative</u>	<u>Interval</u>	
48.9	.8	Continue west on 695, stop at large roadcut just west of Ross Pond. Cross road and walk east along path down to pond south of highway.

**STOP #9**

Domain IV, Hope Valley Alaskite; series of natural exposures from pond shore to the west. Figure 14C is a tape and compass map of these terraces. Note the interlayering of Plainfield Formation quartzites and pelites with the Hope Valley Alaskite. The layering contacts are highly sheared and the layering could be solely tectonic in origin. However, the Hope Valley Alaskite seen here is fine-grained and locally aplitic. This leads to the tenuous speculation that the Hope Valley Alaskite conformably intruded the Plainfield Formation and that the sill (?) contacts have been subsequently tectonically activated. This interlayering can be traced to the N as far as the base of Half Hill

(figure 11b) where a general increase in Hope Valley grain size away from the interlayered zone can be observed. Structural features in the outcrop include 1) the well developed foliation and lineation, 2) the presence of a folded quartz vein, and 3) an asymmetric fold in the mylonitic foliation indicating oblique normal displacement parallel to lineation.

Return to outcrop on W bound side of highway (where vehicles are located).

**Stop 9 continued** - Domain IV, Hope Valley Alaskite. Figure 14C is a photomosaic derived map of this outcrop. Here the gradual transition from Hope Valley Alaskite is observed. The diffuse and arbitrary nature of the contact can be observed. Several aplitic dikes cut foliation. Stained slabs of the Hope Valley Alaskite when cut parallel to lineation and perpendicular to foliation reveal C-S surfaces throughout this domain (figure 16). With one exception they systematically indicate oblique normal displacement.

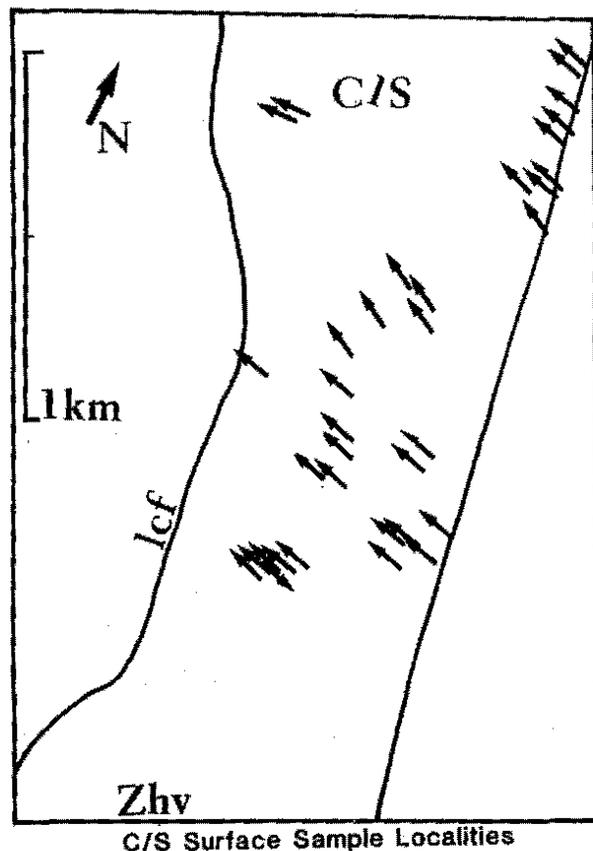


Figure 16. Direction of motion of upper plate of Lake Char fault determined from C and S fabric on stained slabs of Hope Valley Alaskite in domain IV (see figure 11).

<u>Cumulative</u>	<u>Interval</u>	
49.15	.25	Turn left (south) on Squaw Rock Road
49.4	.25	Park in wide shoulder on left side of road by telephone switching box. Walk north to to start of pole and cable guard rail on west side of road. Enter woods walking W and downhill through cleared area. Follow cleared area, through fence, to interchange 90. Walk south along N bound 395/695 past split in highway. Carefully cross highway.

**STOP #10**

Domain V, Quinebaug Formation and Plainfield Formation. Here the Lake Char fault is well exposed (figure 14). The rocks underlying the fault are recrystallized Plainfield Formation quartzites with abundant secondary quartz. Black Quinebaug Formation schist, massive grey Quinebaug Formation, and a coarsely porphyroblastic plagioclase gneiss overlie the fault. The plagioclase gneiss occurs in large tabular lenses which are concentrated at horizons in the Quinebaug Formation schist. These often have diffuse margins, rocks intermediate (in porphyroblast content) between the schist and the plagioclase gneiss are common and lenses are often imbricated (figure 14E). Owens speculates that 1) these represent primary layers of a composition preferential to porphyroblast growth during or before the first phase of metamorphism, 2) that this resulted in a competency contrast, due to which they were boudinaged during isoclinal folding and 3) that they have been rotated and imbricated during mylonitization. Measurements of intrafolial fold axes and rotational sense, from Plainfield Formation quartzites directly beneath the fault, yield a separation angle (figure 17b) consistent with a slip line parallel to the lineation of domain V and indicate oblique normal displacement. To the north in the outcrop plagioclase gneiss layers are cut by pseudotachylites (figure 14E), which have formed in thin seams parallel to foliation and intrude across foliation.

Proceed northward on Squaw Rock Road which turns to the west.

<u>Cumulative</u>	<u>Interval</u>	
50.4	1.0	Turn right onto Green Hollow Rd.

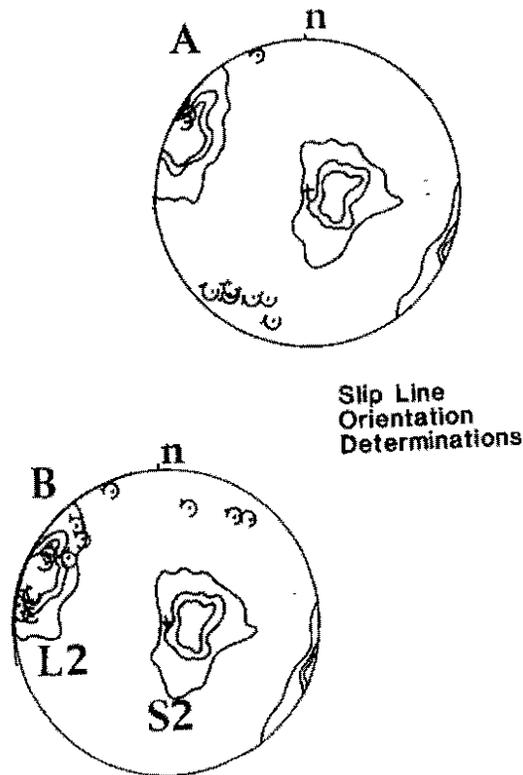


Figure 17. Separation angle determinations of motion direction of Lake Char fault at Stop 10 and an outcrop immediately north of Stop 10; lower hemisphere, equal-area projections.

<u>Cumulative</u>	<u>Interval</u>	
52.4	2.0	Stop sign, go straight across intersection
53.15	.75	Merge with Route 12 in Danielson
53.25	.1	Turn right at stop light
53.6	.35	Turn left onto I395 northbound
61.8	8.2	Take exit 97 for Route 44
61.9	.1	Turn right onto Route 44 eastbound
62.6	.7	Intersection with Route 21, continue on Route 44
64.3	1.7	Intersection with Tucker Hill Road. Park in cleared area at southeast side of intersection

The small roadcuts at this last stop expose several lithologies of the Plainfield formation as well as mesoscopic and microscopic structural evidence of top-down motion. The rocks are quite thoroughly mylonitized and lie within a zone either of fault repetition of the Quinebaug and Plainfield formations or some more complex repetition with mylonitization superimposed upon it. The Plainfield formation here consists of interlayered schist with plagioclase porphyroblasts and massive quartzite. The quartzite displays a spectacular oblique quartz grain shape and the schists, especially where interlayered with quartzites displays asymmetric intrafolial folds (figure 18), extended quartz veins, mica "fish", C and S fabric (figure 7A), and sigmoidal quartz pods all of which indicate top-down motion along the N60W trending mineral lineation. Many quartz veins are present and show crosscutting relationships suggestive of at least three stages of veining, the final one of which is epidote bearing. Another feature of interest here is the local occurrence of fabric forming chlorite.

Return westward on Route 44, I395 is approximately 2.1 miles. To get to the starting point of Goldstein's (1982) field trip to the Lake Char mylonite zone in Webster, Mass. area, which also cites evidence for top-down motion, proceed northwards to exist 100, Wilsonville, Connecticut and follow the directions from there.

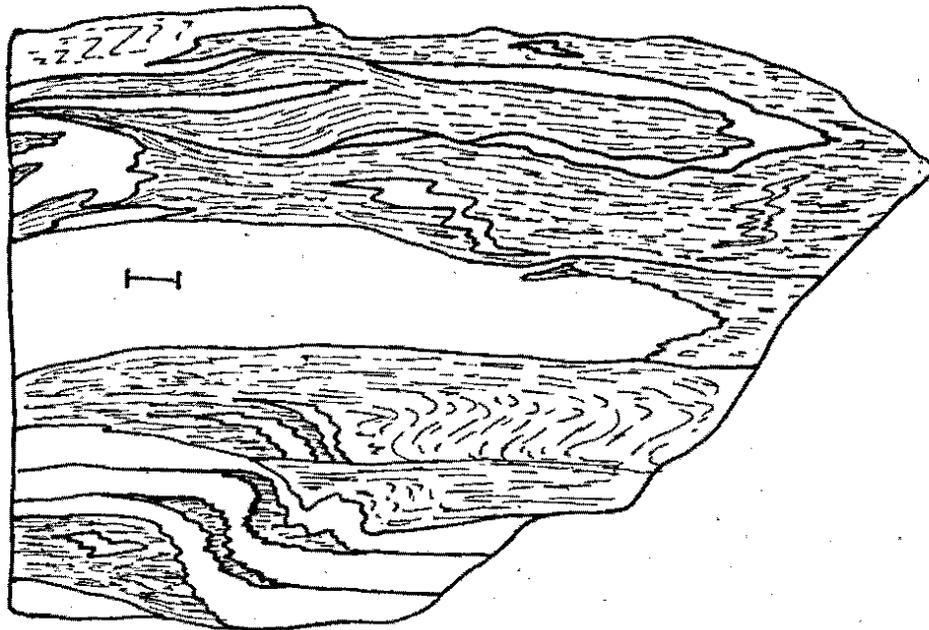


Figure 18. Sketch of folds in Plainfield Formation mylonites at Stop 11. Bar scale is 1 cm, sketched from a slab cut parallel to lineation.



**Trip B-1:** Geology of Southern Connecticut, East-West Transect.  
 Brian J. Skinner and John Rodgers.

This trip crosses the following quadrangles, published quadrangle maps of which are listed below:

	<u>Bedrock Geology</u>	<u>Surficial Geology</u>
<b>New Haven</b> (Stops 1 and 4) (Woodmont)		CG&NHS QR 18 CG&NHS QR18
<b>Branford</b> (Stop 2) (Wallingford; route ticks southwest corner)		CG&NHS QR 14 CG&NHS QR 10
<b>Mount Carmel</b> (Stops 3 and 5)	USGS GQ-199	CG&NHS QR 12
<b>Naugatuck</b> (Stop 6)	CG&NHS QR 9	CG&NHS QR 35
<b>Waterbury</b> (Stop 7)	CG&NHS QR 22	
<b>Thomaston</b> (Stops 8 and 9)		USGS GQ-984

**Stratigraphic Column:** Newark group in Hartford basin (Central Lowland).

Lower Jurassic	Equivalent dikes, according to Philpotts and Martello, ms.
Portland arkose	
"Meriden formation" of Krynine (1950)	
Hampden basalt (3rd flow)	Bridgeport-Pelham dike
East Berlin formation	
Holyoke basalt (2nd flow)	Buttress-Ware dike
Shuttle Meadow formation	
Talcott basalt (1st flow)	Fair Haven and Higganum dikes
Upper Triassic	
New Haven arkose	

**Stratigraphic Column for Rock Units Mentioned in Western Highlands (all age assignments are uncertain)**

Hartland belt	Orange-Milford belt
Silurian and Lower Devonian	
Southington Mountain schist	
• The Straits schist	Wepawaug schist
	(unconformity) Ordovician
Collinsville formation (partly equivalent to Prospect or Harrison gneiss)	Maltby Lakes chlorite schist
	(fault zone) Cambrian
Waterbury gneiss (composite unit)	

**DON'T FORGET TO SET YOUR ODOMETER TO 0.0 BEFORE STARTING OUT!**

(The mileage for this itinerary was measured with a car whose odometer is about 1% fast.)

**Mileage**

- 0.0 Kline Geology Laboratory (town and city of New Haven, New Haven quadrangle).
- 0.2 Exit from Kline parking lot; traffic light. Go straight (east) on Humphrey Street.
- 0.4 Next traffic light; turn right (south) on Orange Street.
- 0.7 Next traffic light; turn left (east) onto ramp, then take right fork onto I-91 South.
- 0.9 Once on I-91, it is necessary to cross at least two lanes of traffic to the left in 0.3 mile in order to take the left ramp for I-95 East.
- 1.6 On I-95, one has a mile to cross back to the right lane before Exit 50 (lane marked exit only). On the way we cross the Mill and Quinnipiac Rivers as they join at the head of New Haven Harbor.

- 2.7 Take Exit 50. Red New Haven arkose (Triassic) is visible to the right from the ramp.
- 3.0 Traffic light; turn right (south) on Woodward Avenue.
- 3.9 Low outcrop of traprock on right, part of a dike connected with the Forbes Bluff sill (see beyond), intrusive into the New Haven arkose.
- 4.5 Optional stop; second parking lot in Fort Hale Park. The hill ahead is held up by part of a saucer-shaped sill of fine-grained gabbro, rather well exposed in Forbes Bluff along the shore.
- 4.7 "T" intersection; turn right (south) on Townsend Avenue.
- 4.9 Route returns to shore of Morris Cove at southeast end of Forbes Bluff, visible to right. Far shore of Morris Cove is underlain by Light House (granite) gneiss beyond Eastern Border Fault of Mesozoic Connecticut basin, here running almost east-west close to that shore.
- 5.5 Traffic light; turn right (southwest) on Light House Road. We are here practically over the Eastern Border Fault, and outcrops appear along the road somewhat farther on.
- 6.1 Enter Light House Point Park, keeping right. Granite in road cuts.
- 6.3 Parking lot for Light House Point. Busses park at Pavilion. (The Lighthouse itself is on the Woodmont quadrangle.)

**STOP 1:**

Outcrops on the shore are the Light House gneiss, here somewhat broken and shattered due to the near proximity of the Eastern Border Fault. Standing on the shore and looking northwest, one sees Morris Cove in the fore-ground to the right. The Eastern Border Fault passes close to the southeastern side of the cove. The Fault then passes southwesterly beneath New Haven Harbor; the gneiss to the south forms a buried escarpment that has been lowered by blasting to provide shipping lanes. Along the fault is a deep valley, now filled with glacial outwash sediments as much as 600 feet deep. Indeed, all of downtown New Haven is built on glacial outwash sands in which there are local lenses rich in organic matter.

On the northern side of Morris Cove stands Forbes Bluff, a small sill of fine-grained gabbro. Beyond the Bluff, East Rock and West Rock, both sills, are visible. The two sills dip gently to the east, having the prevailing dip of the Triassic sediments they intrude. On the far horizon, metamorphic rocks of the Western Highlands can be seen emerging from beneath the Triassic unconformity. The unconformity can be seen in a number of places and will be observed at Stop 4.

Return to traffic light (5.5). (At about 6.95, cars can turn right on Cove Street and cut off corner, then, in less than 0.1 mi., turn right again on South End Road and pick up itinerary at 7.25.)

- 7.1 Traffic light (=5.5). Turn right on Townsend Avenue or South End Road.
- 7.6 Enter town of East Haven and then Woodmont quadrangle. Several outcrops of Light House gneiss nearby.
- 7.8 Stop sign; turn left (east) on Silver Sands Drive.
- 8.4 Turn left (northeast) on Silver Sands Road, soon re-entering New Haven quadrangle.
- 9.2 Enter Branford quadrangle. Gneiss in quarry to left (northwest) of road is badly shattered, being close to border fault.
- 9.3 Stop sign; continue straight (east). We are here following a low ridge over the Light House gneiss just south of the border fault, which lies at the north foot of the ridge.
- 9.5 Stop sign; continue ahead, but soon bear left.
- 9.8 Stop sign; turn left (east) on Route 142. For the next two miles, the route runs through large outcrops of gneiss or granite.
- 10.4 Optional stop to see Light House gneiss.
- 10.5 Cross East River and enter town of Branford.
- 10.9- Road follows shore of Pages Cove; gneiss crops out on  
11.1 headlands and islands.
- 12.6 Traffic light at U.S. 1; we have returned to the border fault. Turn right (east) under RR overpass. Beyond overpass road turns northeast and starts to follow border fault. Prepare to stop.

12.8 STOP 2:

U.S. 1 here follows the Eastern Border Fault. To the west, a fragment of the Talcott Flow, with its distinctive pillow structures, lies in the hanging wall. Across Route 1, east of the fault, the Branford (granite) gneiss is extensively fractured, sheared and hydrothermally altered. Alteration extends for several hundred meters from the fault; feldspars form clays and epidotes, ferromagnesian minerals form serpentine and chlorite, and well formed quartz crystals are common.

Continue northeast on U.S. 1, which soon diverges eastward from border fault into gneiss (Branford gneiss cutting metasediments of Waterford group). Pass traffic light and blinker.

- 13.4 Quarry on right shows sheared Branford gneiss, here including some schist, intruded by dikes of granite pegmatite and granite, both probably marginal to the Stony Creek granite.
- 13.5 Second traffic light; turn left (north) on Cedar Street.
- 13.6 During construction of apartment complex to left, the border fault zone was uncovered; the dip was nearly vertical, in contrast to its 55' dip as recorded north of Lake Quonnipaug in Guilford and also in Manchester. In both those places, the fault trends nearly north-south, whereas here it trends northeast; strike-slip movement may be more important here.
- 13.7 Optional stop, at or just beyond underpass of I-95. Cedar Street here is exactly on the border fault. Entrance ramps to east, especially north of I-95, show highly sheared and partly silicified gneiss (Branford gneiss). Cuts on I-95 at end of entrance ramp to west show maroon sediments.
- 13.75 Turn left onto entrance ramp onto I-95 West.
- 14.0 Outcrops of basalt; same slice as at Stop 2. Beyond, road enters basin of Portland arkose dipping generally toward the border fault and framed by the three lava flows.
- 15.7 Tollbooths on Conn. Turnpike (I-95). The Hampden (3rd) flow, dipping east, passes under the eastern part of the tollbooth plaza without cropping out, but it is exposed in the hills to the north.
- 16.1 Lake Saltonstall, in valley over East Berlin formation. Enter town of East Haven. The big cut just beyond on the right (north) is in the Holyoke (2nd) flow, dipping east, at the south end of Saltonstall Ridge, but the highway and railroad pass along a cross-fault zone that produces a left offset. The little ridge on the south, east of the small bridge over the highway, is the

offset portion of the Holyoke flow; moreover, the shape of Lake Saltonstall at this point reflects the offset (see Figure 1). During the construction of I-95, John Sanders and students in the Yale Field Geology course mapped the roadbed and vicinity and showed that the fault zone is composite, being made of several nearly vertical, east-west faults, not all with left offset. (See also description of Stop 11, Trip C-1.)

- 16.6 Cuts in Talcott (1st) flow, showing pillows, vesicular basalt, and a dike (feeder?). (See description of optional stop in Trip C-1, mileage 76.8.)
- 16.8 Enter New Haven quadrangle.
- 17.5 Enter town (city) of New Haven.
- 17.8 Poor outcrops of New Haven arkose on right (where not hidden by concrete walls). Parallel road to north (old U.S. 1) exposes the same, cut by dikes of the Fair Haven swarm.
- 19.2 Take exit 48 (right exit) onto I-91; follow it to Exit 10.
- 20.1 East Rock sill is prominent ahead; we pass just right of it.
- 21.5 Cross Quinnipiac River. Hills ahead are underlain by dikes of Fair Haven swarm. Big cut beyond apartment complex shows two dikes intersecting at right angles, and much baked arkose.
- 22.7 Enter town of North Haven, then Branford quadrangle.
- 24.0 After Exit 9, Mt. Carmel or the Sleeping Giant is visible straight ahead; it is a large stock of gabbro.
- 25.3 Take Exit 10 onto Route 40. As the road bends left and crosses I-91, the Sleeping Giant is evident off to the right. For a short distance we are on the Wallingford quadrangle but then enter the Mount Carmel quadrangle.
- 27.6 Stop at sign marking Hamden town line.

**STOP 3:**

A sequence of easterly dipping, Triassic flood-plain sediments. Each unit is graded and each is presumed to represent a single flood event. Pebbles in the graded units can be identified with units to the east of the Eastern Border Fault, here more than 10 kilometers away. The green color at the top of each graded unit may be due to a local accumulation of very fine-grained organic matter acting as a reducing agent.

Continue northwest on Route 40.

- 28.4 Traffic light at intersection with Route 10 (Whitney Ave.). Turn left (south).
- 29.8 Village of Centerville. Route 10 turns right, but we continue south on Whitney Avenue.
- 30.1 Entrance to Wilbur Cross Parkway, not available for busses.
- 30.4 Enter New Haven quadrangle.
- 31.9 Bridge over Whitney Lake (on Mill River).
- 33.0 Lake Whitney on left (east). Whitney Peak at north end of East Rock sill across lake; Mill Rock (dike) exposed on right at 33.1. The dam at Lake Whitney is on the site of the original dam for water power for the works of Eli Whitney, the inventor of standardized parts and the cotton gin. An industrial museum has been established at the site, the location of his gun factory. The dam was located where Mill River had a natural waterfall over the Mill Rock dike, which connects the East Rock sill, visible to the left (east), through Mill Rock, to the right (west), and Pine Rock to the south end of the West Rock sill in West Rock. The dike is exposed on the right (west) side of Whitney Avenue.

Stay in right lane and watch for traffic light.

- 33.2 Turn right (west) on Armory Street, following south side of Mill Rock.
- 33.6 "T" intersection; turn left (south) on Prospect Street.
- 33.9 Turn right (west) on Goodrich Street; continue past two stop signs (Winchester and Newhall Aves.) and one traffic light (Shelton Ave.).
- 34.7 Abandoned railway track and second traffic light. Turn right (north) on Dixwell Avenue.
- 34.9 Traffic light; turn left (west) on Arch Street.
- 35.2 Traffic light; turn left (southwest) on Fitch Street.
- 35.4 Enter town (city) of New Haven. We are driving by Southern Connecticut State University.
- 35.7 Turn right (northwest) on Wintergreen Avenue. Pine Rock is visible on right. It is formed by a westward continuation of the Mill Rock dike, but quarrying excavations over the years have shown it to be a complex of small intrusions. To the left is West Rock, the south end of the major West Rock sill, which dips east with the enclosing strata of the New Haven arkose.
- 36.1 The road cut at the west tip of Pine Rock exposes the rounded top

of the dike; the columnar joints form a fan. The large area beyond on the right was a large gravel pit in glacial outwash forming a delta that was built north by part of West River, flowing eastward and northward around the end of West Rock.

- 36.35 Bear left, then turn sharp left (36.4) onto Springside Avenue; road goes south along the east foot of the dip slope of West Rock.
- 36.9 West Rock terminates just to the west; the termination has been made more abrupt by quarrying in the last century. The east corner of the quarry is worth visiting (from here); one sees the top of the sill and overlying baked arkose, both cut by the same columnar joints. The upper surface of the West Rock sill north of here is quite irregular; apparently small dikes projected from it in the direction of Pine Rock, but the actual source of Pine Rock must have been farther down dip, still in the subsurface.
- 37.2 Traffic light; turn right (west) on Blake Avenue.
- 37.5 First of two traffic lights; turn right (northwest) on Valley Road. Impressive view of south face of West Rock. The bottom of the sill is not parallel to bedding here but climbs southward toward the sill's termination (as can be seen clearly when leaves are off the trees).
- 37.9 New bridge over West River. In 1981, the West River flooded following torrential local rains. The bridge was washed away and all houses in the immediate area were flooded. A kilometer downstream, where the West River crosses Whalley Avenue, factories and stores were flooded, and the road surface was uprooted when several meters of water flowed across it.
- 38.3 Traffic light; turn left (southwest) on East Ramsdell Street.
- 38.5 Traffic light; turn right (northwest) on Whalley Avenue.
- 38.7 Take left fork (Route 63, not 69).
- 39.0 Pass under Wilbur Cross Parkway and immediately turn left on first entrance road for Amity Shopping Center; proceed 0.15 mi. past south end of shops to cliff face; turn right and drive behind shops as far as possible.

**39.3**      **STOP 4:**

This is a classic outcrop. The metamorphic rocks of the Western Highlands lie unconformably below the Triassic sediments of the Connecticut Graben. The basement rocks belong to the Maltby Lakes formation (formerly known as the Milford Chlorite Schist), a unit believed to be largely metavolcanic. The angular unconformity on which the coarse

Triassic arkosic grits and conglomerates lie, dips gently toward the east. Across the valley to the east, the West Rock Sill (gabbro) is visible.

The nearly bare rock surface above the stop is approximately the unconformable surface, exhumed by erosion by Pleistocene time, glacially smoothed and covered by till, then exhumed again by man. On the far side of the surface, the Maltby Lakes formation is cut by a dike of fine-grained gabbro, the Buttress dike. The dike, which continues for many kilometers both north and south, can be seen to cut the West Rock sill about 250 meters north of the gap where the Wilbur Cross Parkway tunnels through the sill; it forms a prominent spur, called The Buttress, on the west or scarp face of West Rock. (The gap is not eroded over a fault but over a zone of intense jointing that caused a great deal of trouble during the building of the tunnel.)

Return to Route 63 and turn left (northwest) (39.6). Enter town of Woodbridge.

- 39.8**      Traffic light; turn right (east) on Lucy Street.
- 40.0**      Traffic light; turn left (north) on Litchfield Pike (Route 69). We now follow valley of West River, between scarp face of West Rock on the right, and exhumed pre-Triassic erosion surface on left.
- 42.1**      Dam of Lake Dawson on right. During excavation for this dam, the unconformity of the Triassic on the older rocks (here the Wepauwaug schist) was well exposed, dipping east.
- 42.6**      Enter Mount Carmel quadrangle.
- 42.7**      Cement kiln on left at junction of Dillon Road. Continue straight on Route 69, but we will return here. Route 69 now starts to climb up left side of valley and to cut into Wepauwaug schist (with layers of dirty carbonate rock).
- 43.3**      Stop just short of end of guard rail, opposite south end of large roadcut on left in Wepauwaug schist, here a phyllite.

STOP 5:

The Wepawaug schist here is at chlorite grade and shows three principal rock types:

- 1) muscovite-chlorite-quartz-plagioclase phyllite with more than 40% mica;
- 2) quartz-plagioclase-muscovite-chlorite phyllite with less than 20% mica; and
- 3) calcite-dolomite-muscovite-quartz brown-weathering limestone.

It also includes lenses of "Woodbridge granite", made of plagioclase, quartz, muscovite, and minor K-feldspar, probably felsic tuff. According to Dieterich (1968a, b), the dominant foliation is closely spaced F<sub>2</sub> axial-plane crenulation cleavage; minor F<sub>2</sub> folds are visible in the northern part of the exposure. Streak lineations at an angle to the F<sub>2</sub> axes are F<sub>1</sub> lineations parallel to F<sub>1</sub> fold axes (difficult to see in the natural outcrop but visible in thin-section). F<sub>4</sub> kink-bands are prominent. F<sub>3</sub> has not been seen here.

Make U-turn and return to corner of Dillon Road.

- 43.9 Corner of Dillon Road (= 42.3); turn right in front of cement kiln. The "limestone" burned in the kiln came from a quarry on the hill behind, cut in a large lense of the very impure carbonate rock in the Wepawaug schist, like that seen at Stop 5 (rock type 3).

In this area Fritts (GQ-199) mapped many small bodies of "Woodbridge granite"; these have turned out to be lenses of felsic metavolcanics (Jean Bahr; unpublished senior essay at Yale). Unfortunately no good exposures are accessible to a large group.

- 44.4 Enter New Haven quadrangle.
- 45.3 "T" intersection with Route 63 (Amity Road); turn right (north).
- 45.4 Road ticks southwest corner of Mount Carmel quadrangle and enters Naugatuck quadrangle.
- 45.7 Intersection of Routes 63 and 67; turn left (west) on 67.
- 46.1 Under thick glacial till (note nice drumlin to north), we cross East Derby fault, from relatively lower grade Wepawaug schist into relatively higher grade but probably correlative The Straits schist; the latter was called Cooks Pond schist by Crowley (QR 24) and Derby Hill schist by Fritts (GQ-426 and 199).

- 46.6 Ridge ahead marks edge of Prospect (Harrison) gneiss; along the border is the distinctive Pumpkin Ground porphyritic member, which we will not see here; we will see comparable rocks ("horse-tooth gneiss") at Stop 6. Road follows Bladens River through ridge.
- 48.4 Enter town of Seymour. We are here crossing the Bridgeport synform, a true syncline for Crowley and Hall, an overturned anticline for Dieterich and Rodgers. Beyond, to right, a large gravel pit in stratified drift.
- 49.6 Entrance ramp for Route 8 North. Turn right (north) and join Route 8. The outcrops at the right, at the end of and beyond the ramp, are fairly typical Prospect gneiss, here on the west limb of the Bridgeport synform, though only a few layers show the coarse "horse-tooth" K-feldspars. Much fancy folding of layers and veins.
- 50.3 We now enter the "jelly roll", first worked out by Dieterich (1968a, b). See Figure 1. A deep isoclinal, compound, probably recumbent and west-vergent syncline of The Straits schist forms the jelly layer in cake made of older gneisses (Collinsville and Taine Mountain formations), probably equivalent, in part at least, to Prospect (Harrison) gneiss. The syncline was then rolled up by two later east-vergent fold phases, and the roll itself was then bent over the rising Waterbury dome (north of us) as if over someone's knee. Erosion now gives us two oblique-sections of the jelly roll, north and south of the dome; we are here in the southern oblique-section. The Straits schist is a resistant rock and makes ridges. At and north of Seymour village the upper layer of jelly is crossed three times in a mile by the sinuous Naugatuck River, the two southern crossings (behind us on the left) marked by distinct riffles which were exploited for water power - hence the village. The lower (west) contact of this layer used to be well exposed along Route 8 at 50.3, where Rosemary Vidale studied in detail the reaction between calcite-bearing layers and aluminous The Straits Schist, matching the natural occurrences by experimental work, but the outcrop was entirely destroyed in the most recent relocation of Route 8. The outcrop is described by Vidale in Burger and Hewitt (1968, Trip D-1, p. 15-18; see also Vidale, 1969). The cliff high on the south end of the Hill ahead to the right (Rock Rimmon) shows The Straits schist and the same contact. The road cuts beginning at the Beacon Falls town line (50.5) are in the underlying Collinsville gneiss between the upper and middle layers of jelly, the latter cropping out on the lower slopes of the hills to the left across the river.

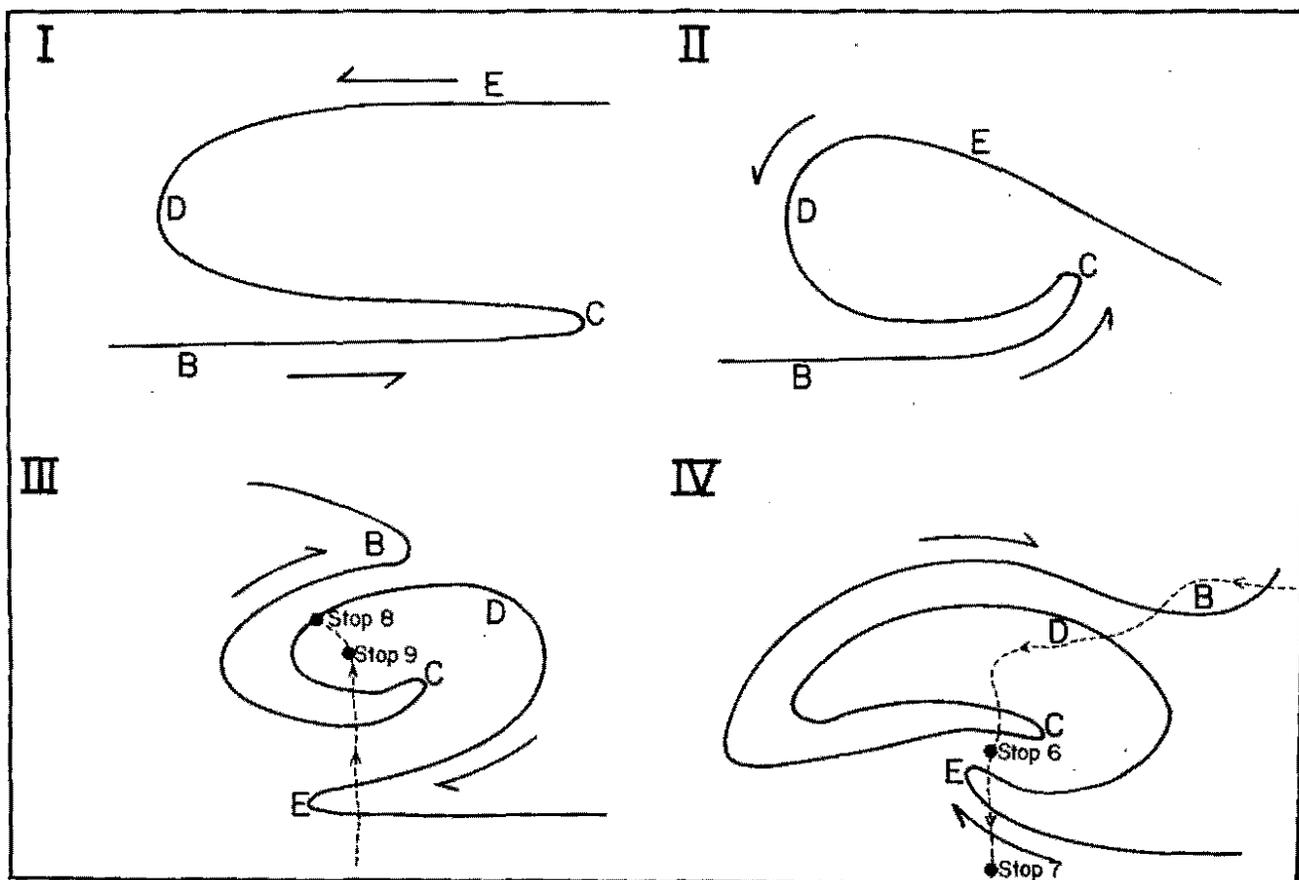


Figure 1. Conceptual scheme to produce the jelly-roll structure, inspired by but somewhat simplified from cross-sections in Dieterich (1968b, Figs. 4 and 5).

- I.  $F_1$  - west-vergent recumbent folding; contact from C to D becomes inverted.
- II. Continued west vergence; hooking of C.
- III.  $F_2$  - east-vergent folding; contact from B to E is steepened and in large part inverted and then flattened, producing present form of northern oblique-section of the roll.
- IV.  $F_3$  - anticline B is carried on eastward and overturned to become Bridgeport synform, east of southern oblique-section of roll. Trip route is very generalized.

- 52.7- Large outcrops on both sides of Route 8 within bend of the  
 52.9 Naugatuck River west of Beacon Falls village. The middle layer of jelly, The Straits schist, forms a separate outcrop on the left (west) side of the road (at 52.5) south of the main outcrops and can be traced to outcrops on old Route 8 in the village and about a mile farther east, where it ends, wrapped around by "cake", itself wrapped around by jelly connecting the upper layer (50.3) with the lower layer, which forms the prominent gorge of the Naugatuck River just ahead of us (53.1 to 54.8). The large outcrops here are in the layer of cake between the middle and lower layers of jelly. Stop near north end of outcrop, short of end of guard rail.

### STOP 6

These new outcrops are some of the finest known of the "horse-tooth" gneiss, filled with large Carlsbad twins of K-feldspar. Carr mapped them as Prospect gneiss, and they certainly fit its Pumpkin Ground member; here, however, we now assign them to the Collinsville formation. As both units are now interpreted as largely felsic metavolcanics of Ordovician age, the difference is not important.

Continue north on Route 8.

- 53.1- Gorge of Naugatuck River through the type belt of The Straits  
 54.8 schist, here the lower layer of jelly in the jelly roll. The Straits is a narrow pass in this ridge 3 miles to the east, where Route 63 passes through on its way from New Haven to Naugatuck. The rock is coarse two-mica schist, almost always slightly graphitic and highly aluminous (garnet, staurolite, usually kyanite), full of pegmatite bodies. The original rock was probably an euxinic black shale.
- 54.9 Enter town (borough) of Naugatuck. Collinsville formation beneath The Straits schist is exposed in low cuts on the railroad across the river, but the locality is not easy to reach. We now approach the Waterbury dome, which Dietsch's work (see trip A-1 in this guidebook) shows is even more complex than we feared. First we pass the bordering rocks (the lowest layer of cake below the lowest jelly), not exposed at all along our route, and then, at about 57.2, we enter the "core" rocks in the eastern of the three antiforms into which Dietsch has dissolved the dome.
- 57.3 Enter Waterbury quadrangle.
- 58.0 Take Exit 29 onto Main Street north.

58.3 Traffic light; turn right and proceed up hill past outcrops. The busses will let us out at the top of the outcrops, then turn around and pick us up at the bottom, or in the parking lot across Main Street.

58.5 STOP 7:

Typical outcrop of some of the rock types in the Waterbury gneiss and the surrounding units, including ultramafic pods, in an intensely sliced up zone cutting across the eastern antiform.

Return to traffic light; turn left on Main Street.

58.7 Immediately take entrance ramp to right onto Route 8 North.

60.0 Rejoin Route 8 and enter town (city) of Waterbury. Several large outcrops on left are in "core" rocks of Waterbury dome. Hill across river in the center of the city, marked by large cross, is a body of orthogneiss (cut by intrusions of several ages) that appears to form an "inner core" within the Waterbury gneiss.

61.6 Enter complex of roads where Route 8 crosses I-84. Stay on Route 8 north, avoiding two right forks and taking third.

63.7 About here we leave the "core" rocks and re-enter the cover rocks, the lowest layer of cake.

64.3 Exit 36; stay on Route 8. Road now starts to climb hill on west side of river. Outcrops on second prominent spur across river (northeast) display contact between Collinsville gneiss and The Straits schist, the latter being part of lowest layer of jelly. As shown by Gates and Martin (QR 22), the jelly syncline is here doubled back on itself across a west-plunging antiform, over which the lowest layer of cake is folded back into the main jelly roll. We are entering the northern oblique-section of the roll.

64.7 Enter town of Watertown.

65.5-  
65.9 Across crest of hill, road passes through the antiform of The Straits schist. White patch near north end of outcrop left (west) of road, close to end of southbound entrance ramp, is kaolin produced by preglacial weathering of a pegmatite body and preserved from glacial erosion. Beyond crest of hill, road returns into Collinsville formation showing much folding.

67.2 Road crosses an isolated lens of The Straits schist within the Collinsville, a stray bit of jelly.

67.5 Enter Thomaston quadrangle.

68.0 Outcrop on left (west) side of southbound lane shows south contact of The Straits in middle layer of jelly. As at 52.5,

this layer extends about a mile northeast of road and pinches out, wrapped around by cake and then the outside layer of jelly.

- 68.8 Outcrop on left (west) shows north contact of middle layer of jelly. This outcrop is very much worth a stop but must be visited from the southbound lane; we describe it beyond (mileage 74.9).
- 69.9 Enter town of Thomaston; beginning of Reynolds Bridge outcrop on left; this is Stop 9 of our itinerary.
- 70.2 Take Exit 38 off Route 8.
- 70.4 At end of ramp, turn left (west) under Route 8.
- 70.7 Traffic light; turn left (west) on Route 6.
- 71.1 Traffic light; jog left, then right (west) on Route 109.
- 71.9 Outcrop on right will be Stop 8.

72.0 STOP 8:

Parking area on left for dry dam across Branch Brook; turn in and park. Outcrops are on north side of Route 109. The Straits schist underlain down the hill by Collinsville formation. Coticule (quartz-spessartine rock), probably a metamorphosed manganiferous chert, can be found in the lowest layers of The Straits. The contact is the base of the upper layer of jelly in the jelly roll and can be followed east and then south to the top of the lower layer at 65.9, or south and then east to the north contact of the middle layer at 68.8.

Turn around and retrace route to traffic light at 70.7 (third light).

- 73.3 Traffic light (=70.7); turn right and then park at entrance to ramp onto Route 8 south (which we will take after the stop) to view the outcrops along the ramp and beyond.

### STOP 9:

Reynolds Bridge outcrop in "Reynolds Bridge gneiss", a particularly disturbed part of Collinsville formation. Original rock was probably interlayered felsic and mafic volcanics; it was probably folded and metamorphosed once in the Taconic orogeny, then again in the Acadian orogeny when it was caught up in the core of the jelly roll. There has been considerable later retrogression, marked by clots of chlorite and epidote. About where ramp joins main Route 8, a late aplite dike cuts cleanly across all preceding structure; Seidemann (1980) obtained a Rb/Sr date of  $345 \pm 11$  Ma for this dike and similar K/Ar dates for both the dike and the surrounding gneiss. The dike is not symmetrical and hence appears to have cooled in an inclined position, perhaps with nearly its present south dip. Somewhat farther along, a thinner dike with a gentle north dip also cuts across the older structure and contains large xenoliths whose source along the wallrock contact can be readily seen.

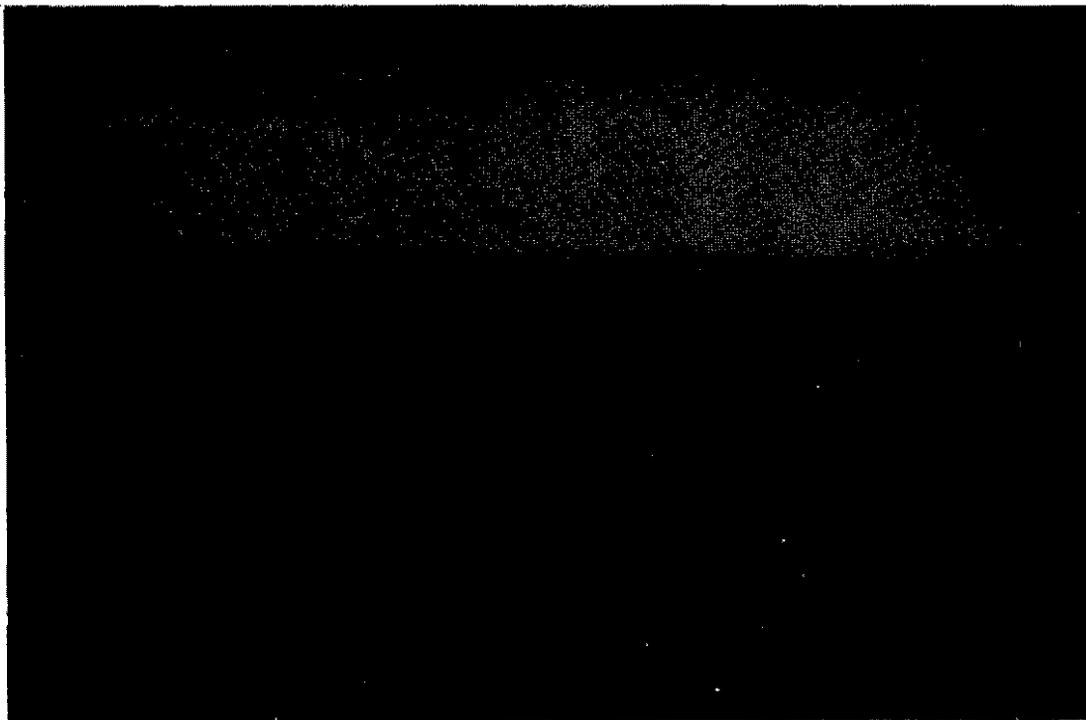
After Stop 9, proceed onto Route 8 south.

- 74.9** Optional Stop. Park where shoulder widens after first outcrops and before beginning of solid outcrop. The rocks at road level north of the parking spot belong to Collinsville formation, those to the south to The Straits schist. At road level, the contact dips steeply south, but traced upward it passes through vertical to north-dipping, then through horizontal to gently south-dipping, beyond which it forms a broad synform. A large aplite dike cuts through all the other rocks near the trough of the synform. Fairly large pods of marble follow part of the contact, especially where it descends to road level. The Straits schist contains tightly folded layers of quartz (once chert?) and cotecule (once manganiferous chert?); some of these layers return on themselves in evident interference structures.

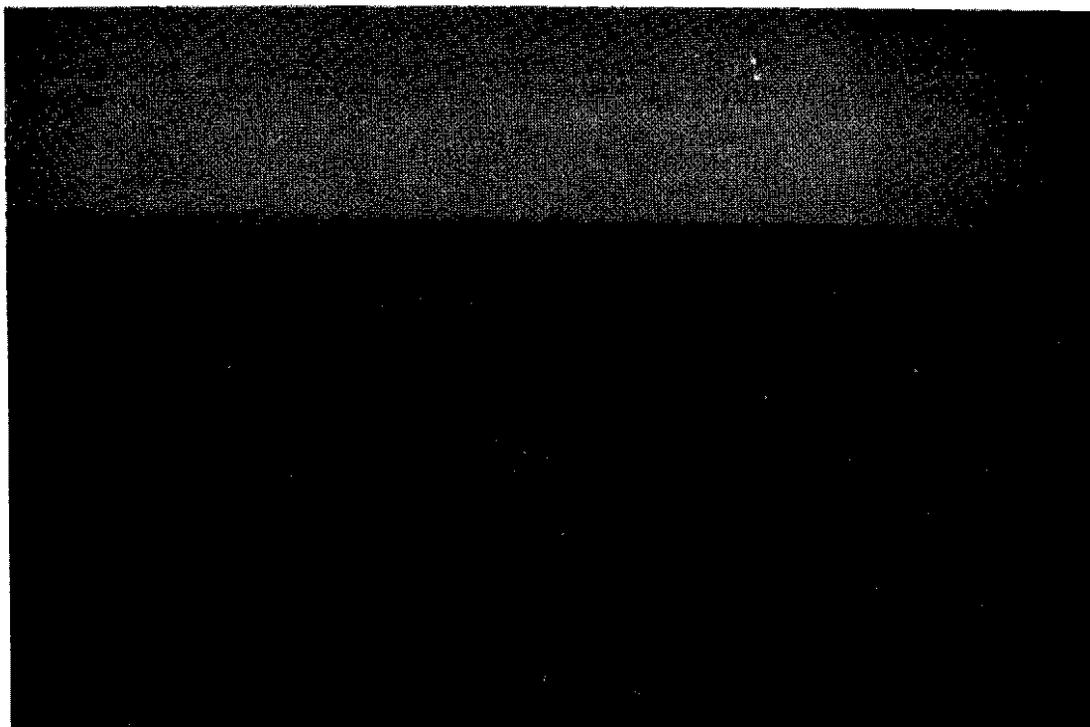
To return to New Haven, continue south on Route 8 to Exit 26 in Naugatuck; there pick up Route 63 south and follow it to New Haven. About 3 miles beyond the turn, Route 63 passes through the Straits, a defile in the lower layer of jelly, which there is turning northward around the southeast corner of the Waterbury dome and can be followed north and then west to the lower layer north of Waterbury (65.5 of itinerary). Distance from Stop 9 to Kline Geology Laboratory is about 30 miles.

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Looking to the northwest at the outcrops of the Holyoke (center) and Talcott Flows (left), Branford, Connecticut. The Holyoke flow is offset by two faults - Route 1 follows the trace of the near fault, Interstate 95 the trace of the second fault. Lake Saltonstall is dammed behind the dip slope of the Holyoke Flow.



West Rock sill, looking north from the Westville area of New Haven, Connecticut. The sill dips shallowly to the east. To the left of the field of view, rocks of the Western Highlands crop out; the escarpment marks the unconformity between the Paleozoic metamorphic rocks and the overlying Mesozoic sedimentary rocks.

STRATIGRAPHY AND STRUCTURAL GEOLOGY  
IN THE BETHEL AREA, SOUTHWESTERN CONNECTICUT

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INTRODUCTION

Detailed stratigraphy and structural geology of the Bethel area, along the east-central edge of the Fordham Terrane (Hall, 1980) will be examined on this field trip. The Bethel area straddles Cameron's Line, a major tectonic and stratigraphic boundary interpreted to be a thrust fault. The general stratigraphy and structural history of this area is correlated with the stratigraphy and structural framework proposed for the White Plains-Glenville area of southeastern New York and southwestern Connecticut by Hall (1968a, 1976, 1980). This correlation is based on detailed mapping, structural analysis, and petrographic study of the bedrock in the Bethel area.

REGIONAL SETTING

The Bethel area (Fig. 1) is situated at the east-central edge of the Fordham Terrane, a northeast trending belt of Precambrian basement rocks that are unconformably overlain by metamorphosed autochthonous rocks, that in turn are overridden by allochthonous Cambrian-Ordovician cover rocks. The eastern part of the Bethel area is underlain by metamorphosed allochthonous Lower Paleozoic rocks known as "eastern region cover rocks" (Hall, 1980), and associated mafic and ultramafic plutonic igneous rocks. Other metamorphosed Paleozoic intrusive rocks, such as the Siscowit Granite Gneiss and related rocks are also present in the region (Figs. 1 and 2).

The Bethel area underwent a complex tectonic history including the effects of the Grenvillian, Teconian, Acadian, and possibly Alleghenian orogenies. Rocks in the western part of the area are in the sillimanite-K-feldspar zone of Paleozoic regional metamorphism but the grade of metamorphism decreases eastward in the Bethel area to the kyanite zone. The metamorphic history of the rocks is complex, as Acadian metamorphic effects overlap Taconian metamorphic effects. Furthermore, Precambrian rocks in the Bethel area contain evidence of relict granulite facies metamorphism and are intimately involved in all phases of Paleozoic deformation and metamorphism along with the cover rocks.

STRATIGRAPHY

Three main subdivisions of stratified bedrock are present in the

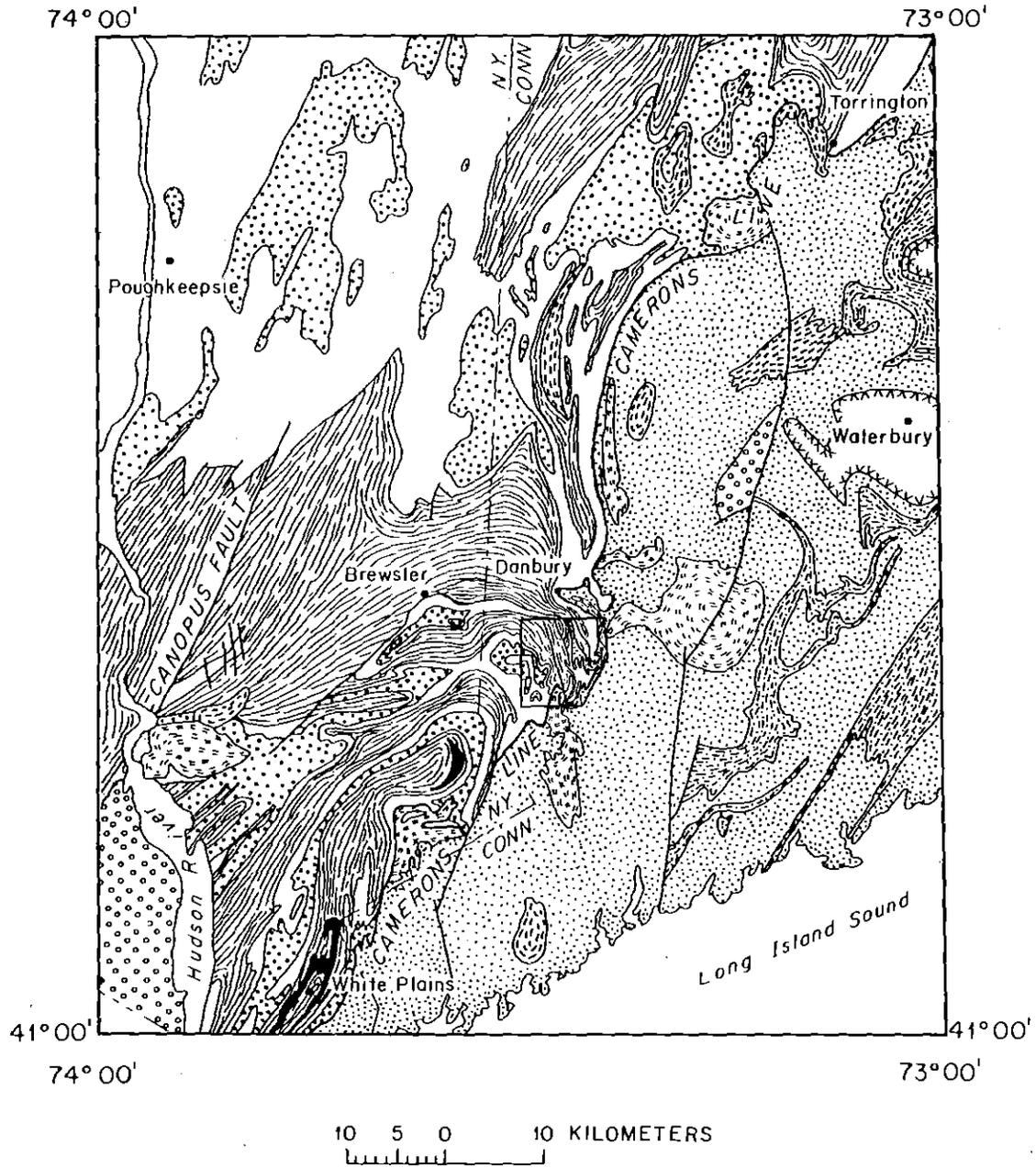
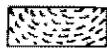


Figure 1. Generalized geologic map of southwestern Connecticut and adjacent New York (modified from Hall, 1980). The Bethel area is outlined south of Danbury.

*Explanation*

INTRUSIVE ROCKS

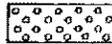


FELSIC PLUTONS



MAFIC PLUTONS

STRATIGRAPHIC UNITS



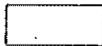
TRIASSIC - JURASSIC



SILURIAN - DEVONIAN

*Western Region*

*Eastern Region*



CAMBRIAN - ORDOVICIAN  
Autochthonous  
Cover Rocks



ALLOCHTHONOUS  
Cover Rocks

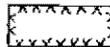


CAMBRIAN - ORDOVICIAN  
Cover Rocks

BASEMENT UNITS



AVALONIAN AGE BASEMENT  
Yonkers and Pound Ridge Gneisses



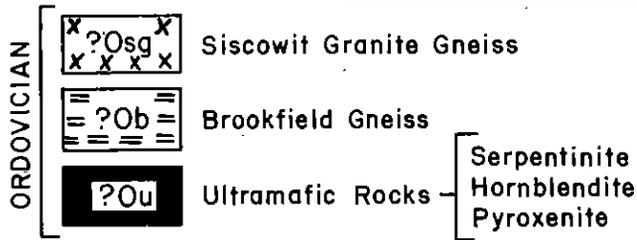
PRECAMBRIAN -  
LOWER ORDOVICIAN  
Basement Gneisses  
of Uncertain Age



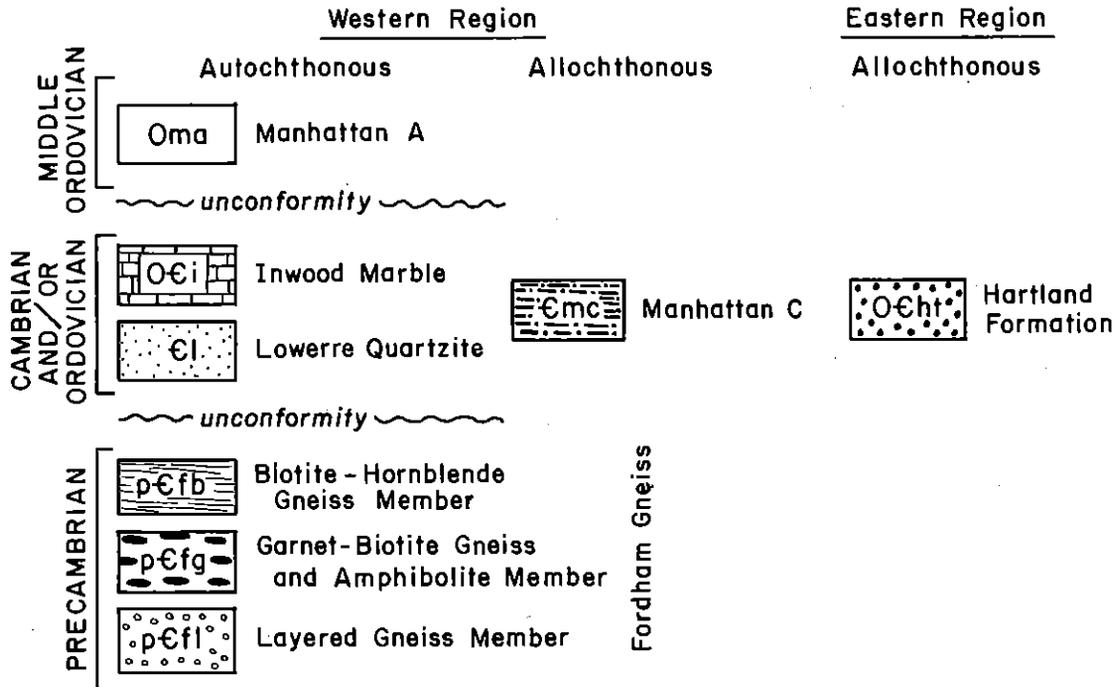
GRENVILLE BASEMENT

# Explanation

## INTRUSIVE ROCKS



## STRATIGRAPHIC UNITS



## SYMBOLS

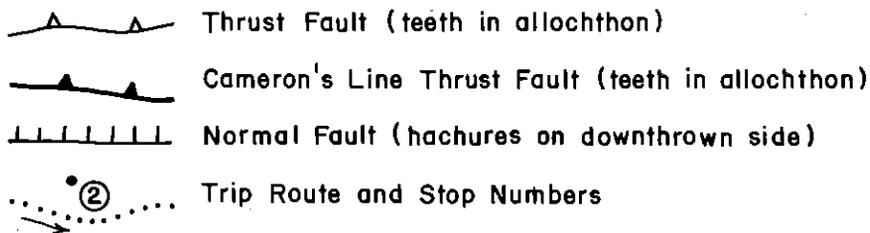


Figure 2. Generalized geologic map of the Bethel area.



Bethel area. These subdivisions are: 1) Precambrian (Grenvillian) gneisses, amphibolites, calc-silicate rocks, and marbles, 2) Cambrian and Ordovician autochthonous and allochthonous rocks west of Cameron's Line, 3) Cambrian and Ordovician allochthonous rocks east of Cameron's Line (Fig. 2).

The Precambrian Fordham Gneiss has been subdivided into three members in the Bethel area (Fig. 2). Their age sequence, from oldest to youngest, is uncertain and the sequence shown on the geologic map was chosen arbitrarily. The members of the Fordham Gneiss mapped in the Bethel area are the Layered Gneiss Member, the Garnet-Biotite Gneiss and Amphibolite Member and the Biotite-Hornblende Gneiss Member. Within these members are other thin mappable horizons including distinctive garnetiferous gneisses, calc-silicate rocks and marbles, not shown in figure 2. The Biotite-Hornblende Gneiss Member is physically continuous with rocks mapped in the adjoining Ridgefield area (Brandon, 1981). Although there are lithic similarities between the members of the Fordham Gneiss in the Bethel area with subdivisions of the Fordham Gneiss elsewhere (Hall, 1968b, 1968c; Alavi, 1975; Brandon, 1981) in southeastern New York and southwestern Connecticut, no specific correlation between these units is made.

The Fordham Gneiss is interpreted by Hall (1968a, p. 121) as "a predominantly clastic and volcanic eugeosynclinal sequence that has undergone high grade metamorphism". Although no truncation of contacts between the three members of the Fordham Gneiss by the angular unconformity at the base of the Paleozoic rocks can be demonstrated in the Bethel area, mappable rock types within the Biotite-Hornblende Gneiss Member (not shown on figure 2) are truncated by the basal Paleozoic unconformity. Hall (1968a) has recognized multiple fold patterns defined by the map patterns of members of the Fordham Gneiss that are truncated by this angular unconformity.

Cambrian and Ordovician autochthonous and allochthonous rocks west of Cameron's Line are interpreted to have been deposited on continental and/or transitional crust. The Fordham Gneiss is unconformably overlain by Cambrian quartzites, granulites, and schists of the Lowerre Quartzite (Stop 1) and the Cambrian-Ordovician Inwood Marble (Stops 2 and 7). The eastern exposures of the Lowerre Quartzite consist of sillimanite-garnet-biotite schist whereas western exposures are predominantly clean quartzites and biotite-quartz-feldspar granulites. It has been suggested (Hall, et al., 1975; Jackson and Hall, 1982) that these different rocks in the Lowerre represent a facies change from west to east and may represent a partial time-transgressive sequence with the rocks in the east deposited prior to those in the west. The Inwood Marble has been divided into two members in the Bethel area, not shown separately in Figure 2. There is a basal member, Inwood A, of thick bedded clean dolomite marble, and an upper member, Inwood B, consisting of interlayered thin bedded clean dolomite marble, silicate-rich dolomite marble, and lesser calc-schists and brown-weathering

quartzites. Middle Ordovician calcite marble and rusty-weathering schists and granulites of Manhattan A (Stop 3) unconformably overlie basement and western region autochthonous rocks.

The western region autochthonous stratigraphy is physically overlain by an allochthonous sequence of schists, schistose-gneisses, gneisses, granulites, and lesser quartzites, interpreted to be Cambrian and/or Ordovician. These rocks, Manhattan C, (Stop 4) are thought to be an eastern facies of the basal clastics and carbonate bank sequence deposited primarily on Grenville basement (Hall, et al., 1975; Hall, 1980).

A major tectonic boundary, Cameron's Line (Rodgers, et al., 1956; Hall, 1980; Merguerian, 1983) separates western region cover rocks and basement described above from eastern region cover rocks and associated mafic and ultramafic plutonic rocks (Fig. 1). Eastern region cover rocks are represented in the Bethel area by the Hartland Formation, a sequence of amphibolites, schists, granulites and gneisses thought to have been deposited on oceanic crust in the Cambrian and Ordovician. The Hartland has been subdivided into three mappable units in the Bethel area but these are not shown separately on the geologic map (Fig. 2).

#### STRUCTURAL GEOLOGY

The geologic map pattern (Fig. 2) is the result of the Precambrian deformation and a sequence of thrusting, followed by multiple folding during the Paleozoic (Fig. 3). The Bethel area has also been affected by post-metamorphic high-angle faults, that are probably as young as Mesozoic.

The Precambrian Fordham Gneiss in the Bethel area, and elsewhere in the Fordham Terrane, underwent at least one phase of intense deformation and associated granulite facies metamorphism during the Grenvillian Orogeny. This deformation produced folds, associated axial plane foliations and other structural features that are truncated by the basal Paleozoic angular unconformity.

The first phase of Paleozoic deformation involved large-scale west directed thrusting that may have been accompanied by local isoclinal folding, particularly in the allochthonous rocks. During this phase of deformation Manhattan C was thrust onto the autochthonous sequence. Apparently, rocks traditionally referred to as part of the autochthonous sequence may also have undergone some thrusting (Figs. 2 and 3). Both the thrust fault that lies beneath Manhattan C and thrust faults within the "autochthonous" sequence are truncated by the Cameron's Line Thrust Fault, which floors the thrust sheet that consists of the Hartland Formation and mafic and ultramafic igneous rocks intruded into the Hartland Formation. The Cameron's Line Thrust Fault has been folded by at least 4 phases of Paleozoic deformation

and, in places where it dips steeply, it appears to have been reactivated with post-metamorphic movement along high angle faults that are as young as Mesozoic. Evidence for thrusting along Cameron's Line includes truncation of units above and below the fault, juxtaposition of rocks interpreted to be older or equivalent in age atop those thought to be younger or the same age and mylonitic rocks at fault zones.

The first phase of major Paleozoic folding occurred after thrusting along Cameron's Line and involved isoclinal folding of basement, cover rocks on both sides of Cameron's Line, and mafic and ultramafic rocks associated with eastern cover rocks. Foliation produced at this time is the dominant foliation in the rocks, although locally, intense subsequent deformation has produced younger foliations that are locally dominant. A second phase of isoclinal folding refolded earlier folds and produced a well developed axial planar foliation. The third phase of Paleozoic folding produced tight to isoclinal folds and was accompanied by a well developed axial planar foliation and a strong lineation. A fourth phase of Paleozoic folding involved open folds with a locally developed axial planar cleavage, which is a crenulation cleavage in the schistose rocks.

Post-metamorphic high-angle faulting may locally involve reactivation of movement along Cameron's Line where it is steeply dipping.

#### METAMORPHISM

The Precambrian Fordham Gneiss contains evidence for granulite facies metamorphism, presumably associated with Precambrian deformation during the Grenvillian Orogeny. The assemblage orthopyroxene-clinopyroxene-hornblende-plagioclase is present in local mafic rocks in the Biotite-Hornblende Gneiss Member. This assemblage is exclusive to rocks in the Fordham Gneiss and is interpreted to be evidence for relict granulite facies regional metamorphism during the Precambrian.

Peak regional metamorphism in the Paleozoic cover rocks in the Bethel area at sillimanite-K-feldspar grade. The grade of Paleozoic regional metamorphism decreases from west to east and the rocks at the east edge of the area are in the kyanite zone of regional metamorphism. Evidence for retrograde metamorphism includes the alteration of biotite to Fe-rich and Mg-rich chlorite and alteration of forsterite to serpentine in calc-silicate rock.

#### ACKNOWLEDGEMENTS

We gratefully acknowledge the financial support of the Connecticut Geological and Natural History Survey. We wish to thank the Town of Ridgefield Recreation Department, the Limekiln Swimming Association, Robert Pattison, Peter Jennings, Frederick Timm, the Danbury Conservation Committee, the Vanderbilt Chemical Company and the

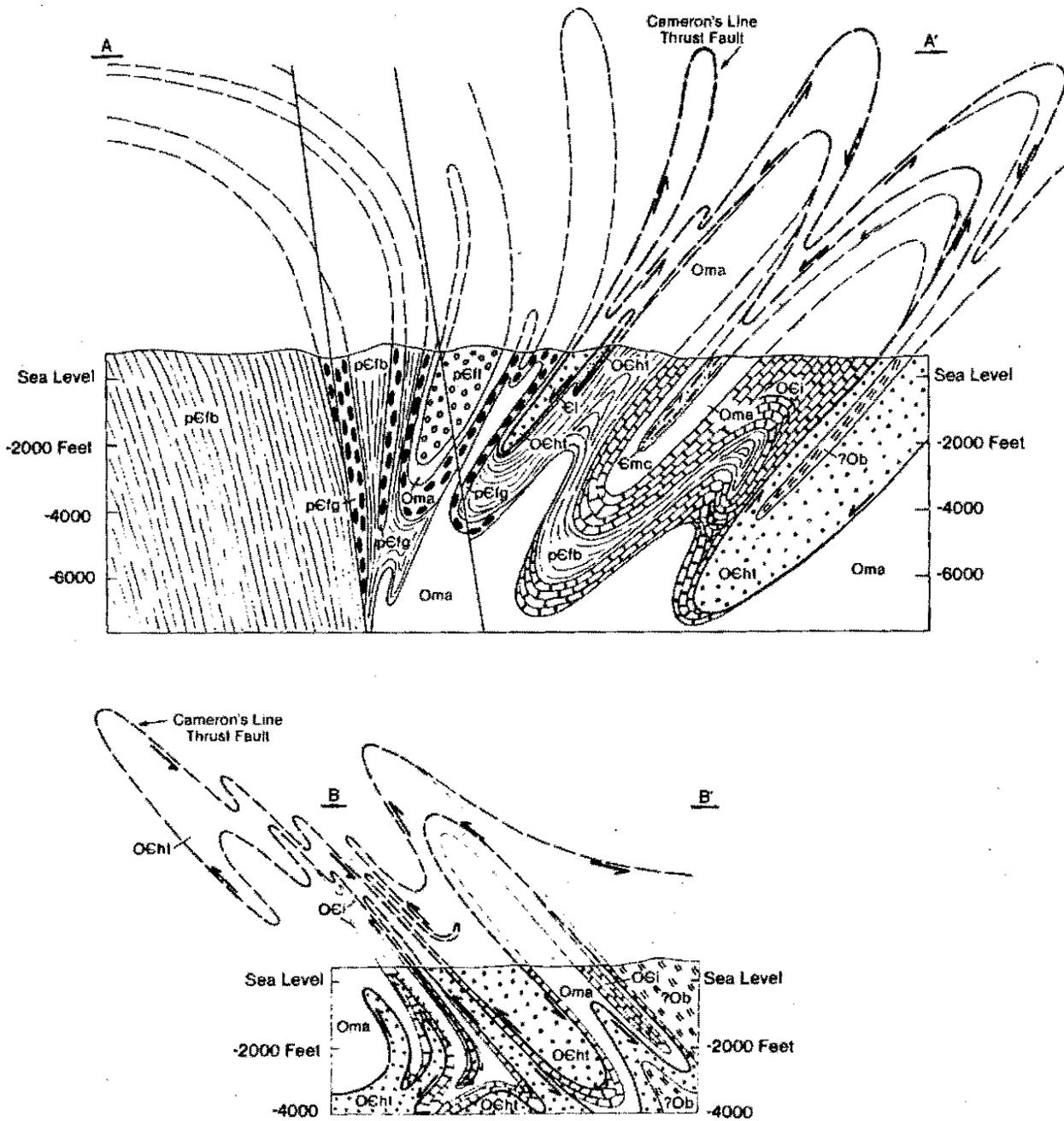


Figure 3. Generalized structure sections A-A' and B-B', located on figure 2.

Consolidated Controls Corporation for their cooperation.

ROAD LOG

From the New Haven, Connecticut area follow Route #34 west to the Merritt Parkway (Route #15). Drive southwest along the Merritt Parkway to Route #7 in Norwalk. Proceed north along Route #7. Approximately 4 miles north of Branchville turn right at the intersection of Route #7 and Great Pond Road, where there is a traffic light. This intersection is 0.75 miles south of the intersection of Route #7 and Route #35 in Ridgefield.

The assembly point is the Martin Park parking lot, located on the north side of Great Pond Road, approximately 0.2 miles east of its intersection with Route #7.

Stop 1. Great Pond, Ridgefield. At this stop we will examine rocks in the vicinity of the angular unconformity between the Precambrian Fordham Gneiss and the Cambrian Lowerre Quartzite along the south shore of Great Pond. The Biotite-Hornblende Gneiss Member of the Fordham Gneiss is unconformably overlain by tan- and rusty-weathering, biotite-quartz-feldspar granulites of the Lowerre Quartzite. The rocks are overturned here due to folding so that the Lowerre Quartzite lies beneath the Fordham Gneiss, dipping moderately northwest.

Rock types in the Fordham Gneiss at this outcrop include gray biotite-hornblende gneiss, (biotite) -quartz-feldspar gneiss and black amphibolite. Discordant and concordant granitic rocks including pegmatite are common, and vein quartz is locally abundant.

The unconformity can only be located to within 2 feet, about 15 feet south of the telephone pole at the south end of the beach, due to limited exposure and to the presence of pink biotite-quartz-feldspar granitic gneiss which obscures the actual unconformity. These particular granitic rocks crosscut the angular unconformity and thus are interpreted to be Paleozoic. However, some granitic rocks at this outcrop are exclusive to the Precambrian rocks and are truncated by the unconformity. The unconformity strikes approximately N20°E and dips 30° to 35° northwest.

The Lowerre Quartzite at this outcrop consists of light-gray- to tan-weathering, biotite-quartz-feldspar granulite. A minor amount of rusty-weathering biotite-quartz-feldspar granulite that contains minor sulfides is present on the west edge of the pathway leading from the parking lot to the beach.

Orientations of compositional layers in the Fordham Gneiss differ considerably from place due to complex folding. The dominant orientation of compositional layering is N63°W and dips 73° northeast. However, near the unconformity, the layering in the

basement gneisses appears to have been rotated into parallelism with bedding (and Paleozoic first phase axial surface foliation) in the Lowerre Quartzite, presumably due to intense Paleozoic deformation. Minor folds in the gneisses include a set of folds that have axial surfaces that strike  $N45^{\circ}W$  and dip  $64^{\circ}SW$ . Fold axes of these folds plunge  $24^{\circ}$ ,  $N62^{\circ}W$ . Concordant granitic rocks within the basement gneisses are folded by this set of isoclinal folds. These folds are apparently refolded by a later set of tight to isoclinal folds that have axial surfaces that strike  $N45^{\circ}E$  to  $N75^{\circ}E$  and dip  $30^{\circ}$  to  $35^{\circ}NW$ . The fold axes of these later folds plunge  $28^{\circ}$  to  $32^{\circ}$ ,  $N55^{\circ}W$  to  $N58^{\circ}W$  and a strong quartz lineation is parallel to the axes. These later folds are evident in both the Fordham Gneiss and the Lowerre Quartzite and are interpreted to be third phase Paleozoic folds. Although most of the granitic rocks at this outcrop appear to be folded by these later folds, a granite dike crosscuts them at one location.

A thin mylonite zone, 1 to 2 feet thick, is present in the Fordham near the northeast end of the outcrop. This zone contains a mylonitic pink pegmatite vein. The mylonite zone strikes  $N70^{\circ}E$  and dips  $75^{\circ}NW$  and displays prominent mineral lineation that plunges  $N71^{\circ}W$  at  $25^{\circ}$ .

Mileage		
<u>Total</u>	<u>Interval</u>	
0.0	-	Leave the parking lot and turn left (east) on Great Pond Road. Follow Great Pond Road (also called Picketts Ridge Road) east to its intersection with George Hull Hill Road.
1.4	1.4	Bear right at this intersection, staying on Picketts Ridge Road. Continue south to the intersection of Picketts Ridge Road and Sympaug Turnpike, where there is a stop sign.
2.2	0.8	Turn left onto Sympaug Turnpike and cross the bridge over the railroad tracks. Turn left after crossing the bridge and proceed northeast along Sympaug Turnpike.
3.1	0.9	Outcrop of "West Redding Garnet Rock" (Balk, 1936, map), a contact metamorphic rock predominantly composed of grossular and diopside, is present on the southeast side of the road.
3.4	0.3	Stop sign at West Redding Post Office. Bear right onto Side Cut Road.

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| 3.5 | 0.1 | Bear left on Side Cut Road and continue to its intersection with Route #53.                              |
| 4.0 | 0.5 | Turn left (north) onto Route #53 and proceed to driveway for Friendly Wood and Wire Fence Company.       |
| 5.0 | 1.0 | Turn left into driveway and proceed through main gate to the parking lot at the west edge of the quarry. |
| 5.4 | 0.4 | Parking lot at Limekiln Swimming Association Quarry.   |

Stop 2. Limekiln Swimming Association Quarry, Bethel (Fig. 2).

At this stop we will examine Members A and B of the Inwood Marble in the quarry. We will also make a short traverse along the quarry road to view the Cameron's Line Thrust Fault, which at this locality brings the Hartland Formation in thrust fault contact with Member B of the Inwood Marble.

The contact between Inwood A and Inwood B is exposed at the entrance to the quarry, with Inwood A lying east of and below Inwood B. Inwood A consists of thick bedded, gray to light-gray, clean dolomite marble. Some beds contain sparse phlogopite and iron sulfides. A 2-inch thick dark-gray quartzite is present in Inwood A at this locality.

Inwood B consists of thin bedded, gray- or dark-gray-weathering, gray dolomite marble. Some beds contain abundant phlogopite and green serpentine after forsterite. Iron sulfide minerals are also present in some beds. Clean gray dolomite marbles are characteristically interbedded with silicate-rich dolomite marbles in beds up to 10-12 inches thick. Calc-schist layers are present in Inwood B along the southwest inner wall of the quarry.

Bedding in the Inwood Marble strikes N5°W to N5°E and dips 44° to 46° west. An isoclinal fold in bedding, interpreted to be a first phase Paleozoic fold, is present on the east facing wall on the north side of the quarry entrance. This fold has an axial surface that strikes N-S, and dips 45° west and a fold axis that plunges 44°, N65°W. The 2 inch thick dark-gray quartzite in Inwood A outlines outstanding fold digitations.

We will leave the quarry and travel south along the road by foot. At the hairpin curve in the road 50 feet west of the inner gate we will examine the rocks in the vicinity of the Cameron's Line Thrust Fault. Because of folding this thrust fault is present in two places at the hairpin curve in the road. At this locality the Cameron's Line Thrust

Fault brings the Hartland Formation into contact with Inwood B.

The Hartland Formation here consists of sillimanite-garnet-biotite schist, black biotite-schistose-granulite and thick bedded black amphibolite. The rocks are intensely sheared and this shearing is interpreted to have taken place as the Hartland Formation was thrust over more competent rock prior to the movement which finally placed it in thrust contact with Inwood B. Amphibolite beds in the Hartland strike  $N32^{\circ}E$  and dip  $42^{\circ}$  west. Minor folds in Inwood B have an axial surface orientation that strikes  $N17^{\circ}E$  and dips  $42^{\circ}$  west and fold axes that plunge  $42^{\circ}$ ,  $N26^{\circ}W$ . Both dextral and sinistral folds, presumably of the same order, are present in the outcrop at the base of the slope on the south side of the hairpin curve. These tight to isoclinal folds are beautifully outlined by the clean dolomite and silicate-mineral-rich dolomite beds of Inwood B.

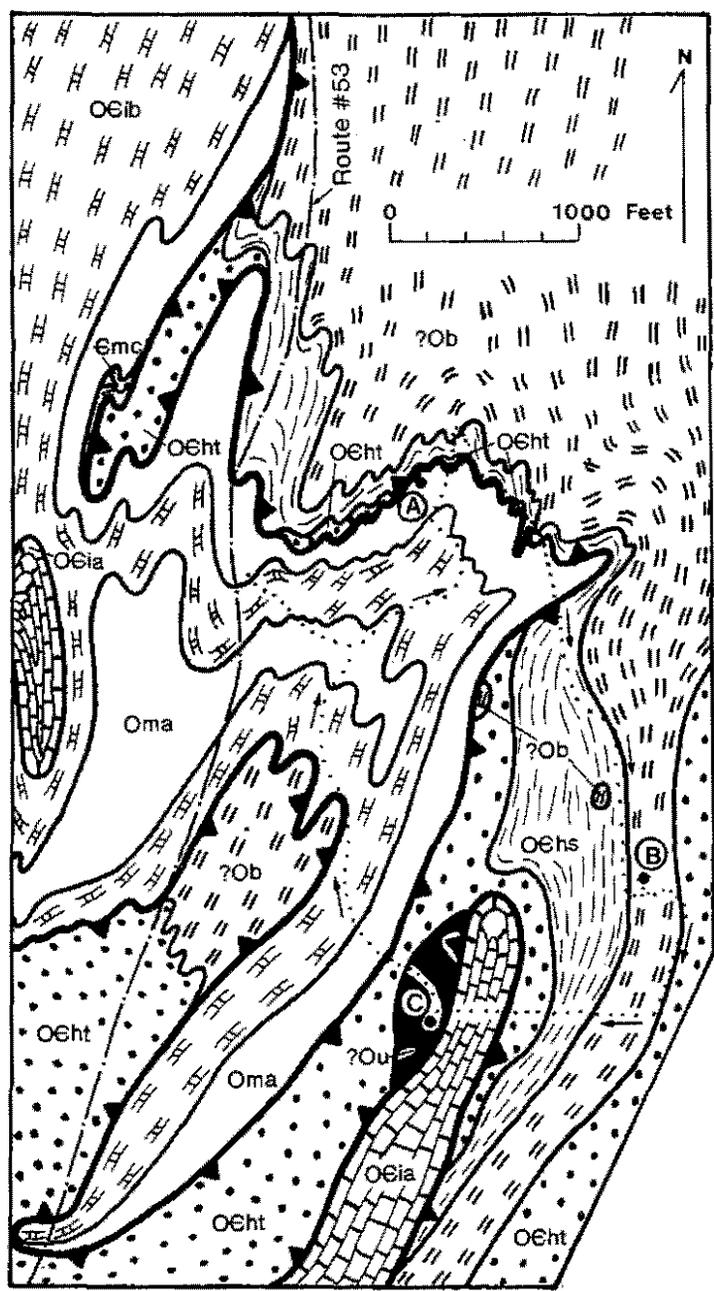
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|-----|-----|---|
| 5.0 | -   | Return to the vehicles and retrace the route to Route #53.  |
| 5.4 | 0.4 | CAUTION. BE CAREFUL WHEN CROSSING ROUTE #53. Turn left (north) onto Route #53 and proceed 0.1 mile to the driveway marked "Pattison" on the right (east). |
| 5.5 | 0.1 | Turn right into the driveway and park on the field north of the driveway.   |

Stop 3. Pattison's Spur, Bald Rock, and Serpentinite Hill Traverse, Bethel-Redding. This stop includes a traverse along the Cameron's Line Thrust Fault and shows some of the rocks of the autochthonous section beneath Cameron's Line. We will also examine the Hartland Formation, the Brookfield Gneiss and an ultramafic body, all of which are above the thrust fault. Proceed east along the trail that leads from the parking area to the base of the cliffs to the east (Fig. 4).

Station A (Fig. 4) - The Manhattan A Marble Member is exposed at the base of this spur and consists of light-gray-weathering, white to light-gray calcite marble with minor iron sulfides and phlogopite. Interbedded with the marble at this locality are rusty-weathering calcareous granulites and a 10-12 feet thick lens of brown-weathering, sillimanite-garnet-biotite schist, biotite schist, and tan- to brown-weathering calcite marble. Rusty-weathering, siliceous granulites in beds of 1 to 2 inches thick, and thin calc-silicate rocks are also present. A mappable body of white granite is present in Manhattan A Marble here, and it contains inclusions of schist and marble. Orientations of compositional layering ranges from strikes of  $N40^{\circ}E$  to  $N50^{\circ}E$  and dips of  $41^{\circ}$ - $51^{\circ}$  NW.

Proceed up the spur from outcrops of Manhattan A to outcrops at

B2-14



*Explanation*  
INTRUSIVE ROCKS

- ORDOVICIAN
- Brookfield Gneiss
  - Ultramafic Rocks - Serpentine

STRATIGRAPHIC UNITS

- |                            |                        | <u>Western Region</u> |                        | <u>Eastern Region</u>          |
|----------------------------|------------------------|-----------------------|------------------------|--------------------------------|
|                            |                        | <u>Autochthonous</u>  | <u>Allochthonous</u>   | <u>Allochthonous</u>           |
| MIDDLE ORDOVICIAN          | <b>Oma</b> Manhattan A |                       |                        |                                |
|                            | unconformity           |                       |                        |                                |
| CAMBRIAN AND/OR ORDOVICIAN | <b>OEib</b> Member B   |                       |                        |                                |
|                            | <b>OEia</b> Member A   |                       | <b>Emc</b> Manhattan C | <b>OEht</b> Hartland Formation |
|                            |                        |                       |                        | <b>OEhs</b> Spotted Granulite  |

SYMBOLS

- Thrust Fault (teeth in allochthon)
- Cameron's Line Thrust Fault (teeth in allochthon)
- Traverse Route and Station Letters

Figure 4. Geologic map showing traverse for Stop 3.

the top of the spur. A section of the Hartland Formation 40 feet thick is exposed here. The Cameron's Line Thrust Fault can be located to within 20 feet and is obscured by lack of outcrop. The Hartland Formation immediately above Cameron's Line consists of approximately 20 feet of thin bedded, light-gray biotite granulite, black amphibolite, and gray pinpoint-garnet-biotite granulite. A section of gray (sillimanite)-garnet-biotite spotted granulite with distinctive 1/8 inch crystals of white plagioclase is 20 feet thick and lies east of these various Hartland rocks which are overturned due to folding. This sub-unit is interpreted to be a sheared rock in the Hartland Formation, owing its texture to movement along the Cameron's Line Thrust Fault and/or tectonic movement of the overlying Brookfield Gneiss relative to the Hartland Formation.

Continue up the spur to the contact between the spotted granulite unit of the Hartland and the Brookfield Gneiss. Of particular interest here is that the degree of shearing within the Brookfield is greatest closer to Cameron's Line. This contact is interpreted to have been an original intrusive contact along which there has apparently been fault movement.

Compositional layering at the Brookfield-Hartland contact strikes N5°E and dips 57° west. Tight folds in the Hartland have an axial surface orientation that strikes N21°E and dips 55°NW and fold axes that plunge 42°, N72°W. These folds are interpreted to be third phase folds and to have developed during Acadian deformation.

The Brookfield Gneiss consists of gray (hornblende) -biotite-quartz-feldspar gneiss with distinctive megacrysts of K-feldspar and plagioclase that are up to 2 inches long. Compositional layering within the Brookfield strikes N31°E and dips 43°NW.

Proceed south to Bald Rock (Fig. 4, Station B). The Bald Rock outcrop consists of the Brookfield Gneiss with abundant feldspar megacrysts. Inclusions of biotite-hornblende gneiss up to 6 feet long and 4-5 inches wide are present within the Brookfield. Bald Rock affords a good view of the city of Danbury to the north.

Proceed south and then west, across an anticline in Cameron's Line to Station C at Serpentine Hill (Fig. 4). The dull green, highly jointed rock exposed on this small hill is a very coarse grained serpentinite body that lies immediately above the Cameron's Line Thrust Fault. Compositional layering at the north end of the body strikes N10°E to N25°E and dips 73°W. Joints strike N40°W and dip 50° SW and are closely spaced.

Proceed northwest to the base of the hill, and then north. Note the isolated body of Brookfield Gneiss intrusive into the Hartland Formation along this final leg of the traverse. Return to the

vehicles.

LUNCH

- |     |     |  |
|-----|-----|--|
| 5.4 | -   | Proceed north along Route #53 to Reservoir St. in Bethel.  |
| 7.7 | 2.3 | Turn left (west) onto Reservoir St.  |
| 8.0 | 0.3 | Stop sign at intersection of Reservoir St. and Bethpage Drive. Outcrop of Manhattan A Marble Member on the right (north) side of the road. |
| 8.2 | 0.2 | View of Bogus Mountain (Precambrian Fordham Gneiss) to the left (south).   |
| 8.9 | 0.7 | Turn right (east) into the Nature Center and follow the dirt road to the parking lot on the right.   |
| 9.1 | 0.2 | Parking lot on right (south) side of Nature Center Road.   |

Stop 4. Nature Center Traverse. At this stop we will examine several rock types in Manhattan C exposed on a small hill south of the parking lot. Rock types in Manhattan C here include gray-weathering, sillimanite-garnet-biotite schistose gneiss, dark-gray to black, fissile sillimanite-garnet-biotite schist with sillimanite nodules and garnet-sillimanite-biotite schistose granulite.

Compositional layering in Manhattan C strikes N27°W to N30°W and dips 57° to 68° SW. Biotite and sillimanite lineations plunge 54°, N57°W at one location.

Contact relations with nearby Manhattan A are obscured by cover. The two hills to the west and south are underlain by calcite marble and rusty-weathering schist of Manhattan A.

These outcrops of Manhattan C are located in a doubly plunging first phase syncline that extends into the adjoining Danbury quadrangle to the north. The fold is interpreted to be downward closing and plunging northwest at both the south and north ends. The curvature of the fold hinge is extreme, with both north and south closures plunging moderately northwest. This curvature is interpreted to have occurred during the first phase of Paleozoic folding due to differential movement during the folding and not by a subsequent deformation, thus it is what is commonly referred to as a sheath fold.

- |     |   |  |
|-----|---|--|
| 9.1 | - | Return to the vehicles and retrace route |
|-----|---|--|

- to the paved road (Mountainville Avenue).
- |      |     |  |
|------|-----|--|
| 9.3  | 0.2 | Turn right (north) onto Mountainville Avenue.  |
| 9.5  | 0.2 | View to west of Thomas Mountain which is underlain by the Biotite-Hornblende Gneiss Member of the Fordham Gneiss. The Lower Paleozoic unconformity is located at the east foot of Thomas Mountain. |
| 10.2 | 0.7 | Turn left onto Southern Boulevard.   |
| 10.4 | 0.2 | Bear left onto Brushy Hill Road  |
| 11.6 | 1.2 | Turn right onto Old Post Road.   |
| 11.8 | 0.2 | Turn right onto Deal Drive and proceed up the hill. Outcrops of the Garnet-Biotite Gneiss and Amphibolite Member of the Fordham Gneiss on both sides of Deal Drive.                                |
| 12.0 | 0.2 | Proceed around the circle and retrace route back to Old Post Road  |
| 12.3 | 0.3 | Park on the right (east) side of Old Post Road. The outcrops for Stop 5 are located on the east side of Brushy Hill Road, opposite from its intersection with Old Post Road.                       |

Stop 5. Brushy Hill Road Pyroxenite, Danbury. At this stop we will examine a small body of biotite pyroxenite exposed near the base of the Hartland Formation, above the Cameron's Line Thrust Fault. The pyroxenite body is at least 50 feet long and is bordered on the east by schists of the Hartland Formation and a white granitic body. Petrographic study of the pyroxenite reveals that it is composed of augite, hornblende, biotite and minor plagioclase. This body is interpreted to have been intruded into the Hartland Formation prior to or during(?) the initial movement along Cameron's Line.

The rocks here are located on the west limb of a major second phase fold in Cameron's Line which refolds a major first phase syncline in Cameron's Line.

Nearby outcrops of the Hartland Formation consist of gray sillimanite-garnet-biotite gneiss. Compositional layering in this gneiss strikes N5°E and dips 68° west. It is uncertain whether the contact between the Hartland and the pyroxenite is entirely intrusive, tectonic, or in part both. Another ultramafic body, a hornblendite, is located approximately 1.5 miles south of here in a similar structural position.

- |      |     |  |
|------|-----|--|
| 12.3 | -   | Return to vehicles and proceed north to the intersection of Old Post Road and Brushy Hill Road.          |
| 12.4 | 0.1 | Turn left (north) onto Brushy Hill Road and retrace route to Southern Boulevard.                         |
| 13.7 | 1.3 | Bear right onto Southern Boulevard and continue northeast to its intersection with Mountainville Avenue. |
| 13.9 | 0.2 | Turn right (south) onto Mountainville Avenue.  |
| 14.9 | 1.0 | Bear left onto Reservoir Street.   |
| 15.4 | 0.5 | Stop sign at Bethpage Drive. Proceed straight along Reservoir Street.                                    |
| 15.7 | 0.3 | Stop sign at Grassy Plain Street (Route #53). Turn right (south) onto Grassy Plain Street.               |
| 16.0 | 0.3 | Turn left onto South Street.   |
| 16.1 | 0.1 | Stop sign at Blackman Street. Bear right along South Street.   |
| 16.3 | 0.2 | Turn right onto Taylor Avenue. Cross railroad tracks.  |
| 16.5 | 0.2 | Turn right into driveway for Vanderbilt Chemical Company.  |
| 16.5 | -   | Park in Vanderbilt Chemical Company parking lot.   |

Stop 6. Brookfield Gneiss at the Vanderbilt Chemical Company, Bethel.  
 At this stop we will examine the Cameron's Line Thrust Fault at a point where the Brookfield Gneiss is in thrust contact with Inwood B of the autochthonous section. The Brookfield hill is the southern extension of the rocks at Shelter Rock in the Danbury Quadrangle (Clarke, 1958). Because of folding here, the Brookfield dips west beneath Inwood B. The Brookfield consists of gray biotite-hornblende gneiss with microcline megacrysts. Inclusions in the Brookfield are present near the west edge of the outcrop and consist of thin layers of garnet-biotite granulite and black amphibolite of the Hartland Formation. These layers of the Hartland Formation and the Brookfield Gneiss are folded though it is uncertain whether these folds occurred during one of the main phases of Paleozoic folding or whether the folds were produced in relation to thrust faulting during movement along the

Cameron's Line Thrust Fault. The Brookfield is extremely sheared here and the degree of shearing and production of mylonitic rocks appears to be a function of distance from Cameron's Line. The shearing within the Brookfield is interpreted to have been the result of thrusting against other competent rocks during transport into thrust contact with Inwood B. This mylonitic texture within the Brookfield is folded by folds that are generally interpreted to be third phase Paleozoic folds, however there are numerous candidates for earlier phase folds in the outcrop. Some of these folds may have developed as early as the thrusting along Cameron's Line. The general orientation of the axial surfaces of the folds is N30°E, 43°NW. The fold hinge lines and mineral lineations have a general plunge of 40°, N50°W. There are several gullies that are parallel to the strike of the Brookfield at this outcrop that may be the result of weathering of subsidiary shear zones within the Brookfield. The movement sense along the Cameron's Line Thrust Fault as deduced by asymmetric feldspar megacrysts and the angular relationship between s- and c-surfaces indicates west-directed thrusting of the Brookfield.

- |      |     |   |
|------|-----|---|
| 16.5 | -   | Return to the vehicles. Turn left onto Taylor Avenue.   |
| 16.7 | 0.2 | Turn right onto South Street. Cross railroad tracks.  |
| 16.8 | 0.1 | Turn left onto Depot Place.   |
| 16.9 | 0.1 | Jog left and right at the traffic light at the intersection of Depot Place and Greenwood Avenue.                                    |
| 17.0 | 0.1 | Continue past Bethel Post Office on the left.   |
| 17.1 | 0.1 | Turn left into the parking lot on the west side of the road opposite the Consolidated Controls Corporation plant on Wooster Street. |

Stop 7. Cameron's Line Thrust Fault at the Consolidated Controls Corporation, Bethel. At this stop we will view the Cameron's Line Thrust Fault where the Hartland Formation lies in thrust fault contact with Inwood B of the autochthonous section. Apparently the Hartland Formation is truncated by Cameron's Line between this outcrop and the Brookfield at Stop 6, 0.75 miles to the south (Fig. 2).

The Hartland Formation exposed here consists of brown-weathering biotite-quartz-feldspar granulite that has a mylonitic texture along its contact with Inwood B. The production of this mylonitic texture is interpreted to be the result of thrusting of the Hartland against other competent rocks prior to their emplacement against Inwood B.

Inwood B consists of thin bedded, clean dolomite marble and silicate-mineral-rich dolomite marble with minor brown- and rusty-weathering quartzites. Accessory minerals in Inwood B marble include phlogopite, serpentine after forsterite, and some pyrite.

The outcrop dramatically shows at least three phases of folding involving the Cameron's Line Thrust Fault. Because of folding, Inwood B overlies the Hartland Formation in most of the outcrop, dipping moderately west.

Abundant minor folds, most of which are third phase folds, are present in this exposure. A poorly developed cleavage is parallel to the axial surface of the third phase folds and a pronounced mineral lineation is parallel to their hinge lines. Earlier isoclinal folds, inferred to be first phase folds, are refolded by third phase folds. Although these earlier folds may be second phase deformational features their axial planar foliation appears to be the first phase regional foliation. Therefore they are interpreted as first phase folds. The map scale fold is also inferred to be a first phase fold because of its relationship to these minor isoclinal folds.

There are later minor folds present that are open and have axial surface orientations that strike N62°W to N90°W and dip 55° to 65° NE to N. These open folds have a weakly developed axial surface foliation and are interpreted to be fourth phase folds. These folds may be associated with map scale folds that are reflected by broad changes in the regional map pattern of the rocks in southeastern New York and southwestern Connecticut.

- |      |     |   |
|------|-----|---|
| 17.1 | -   | To return to the Yale University Campus, exit from the parking lot and proceed to the traffic light at Greenwood Avenue.  |
| 17.3 | 0.2 | Turn left onto Greenwood Avenue (Route #302).   |
| 17.6 | 0.3 | Traffic light at intersection of Greenwood Avenue and Chestnut Street.  |
| 17.9 | 0.3 | Bear left, staying on Greenwood Avenue (Route #302).  |
| 18.4 | 0.5 | Traffic light at the intersection of Route #302 and Route #58 (Putnam Park Road). Turn right (south) onto Route #58. Follow Route #58 south through Redding, Aspetuck, and on to Fairfield. |

Intersection of Route #58 and the Merritt Parkway (Route #15). Proceed northeast along the Merritt to the exit for Route #34 east, which leads to New Haven.

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The Timing and Nature of the Paleozoic Deformation  
in the Northern part of the Manhattan Prong, Southeast New York

by  
Patrick W.G. Brock and Pamela C. Brock

### INTRODUCTION

The Paleozoic rocks of the northern part of the Manhattan Prong in the Croton Falls and Peach Lake map areas have been repeatedly deformed and metamorphosed. The main objectives of this field trip are 1) to demonstrate the age relationships between the deformational events, 2) to show examples of those characteristics that help distinguish each deformation and its associated metamorphism and so permit the tracing and recognition of these events in other parts of the area, 3) to point out the constraints that we have so far on the ages of the deformational events. In addition we will point out those places where we feel it necessary to reinterpret the stratigraphic relationships determined by earlier workers.

#### Regional Setting

The Manhattan Prong of southeastern New York and western Connecticut is underlain by polydeformed metasedimentary and metaigneous rocks of Proterozoic Y (Grenvillian) through Lower Paleozoic age. It is bounded on the north and west by the Grenvillian rocks of the Hudson Highlands, to the southwest by the Mesozoic rocks of the Newark Basin, and to the east by the Cambro-Ordovician rocks of the Hartland Terrane (see Figure 1). Early mapping carried out by Merrill (1896), Fettke (1914) and Fluhr (1950), among others, defined general structural trends as well as the three dominant units: Fordham Gneiss, Inwood Marble, and Manhattan Schist.

#### STRATIGRAPHY

The oldest rocks in the Manhattan Prong are the metasedimentary and metaigneous rocks of the Fordham Gneiss. Rb-Sr whole rock data indicate that at least part of the Fordham Gneiss is about 1,350 m.y. old (Mose, 1982). Zircon studies have also suggested Proterozoic ages for parts of the Fordham (Grauert and Hall, 1973). These rocks underwent Grenvillian metamorphism, which peaked at about 1,100 m.y. (Mose, 1982) and reached at least lower granulite facies in the Fordham (Brock and Brock, 1983).

In the central Manhattan Prong, Grenvillian units of the Fordham are truncated by the Yonkers Granite Gneiss (Hall, 1980), which has yielded Rb-Sr whole-rock ages of  $563 \pm 30$  m.y. (Long, 1969) and  $530 \pm 43$  m.y. (Mose, 1981). The Yonkers Gneiss has been mapped as part of the Fordham, and is always found below the recognized Paleozoic stratigraphic units. The absence of inherited ages in its zircons suggests that it originated as an igneous body (Grauert and Hall, 1973). A similar unit, the Pound Ridge Granite Gneiss, is found in the Fordham of the northern Manhattan Prong; it has been dated at  $583 \pm 19$  m.y. by Mose and Hayes (1975), who suggested that both

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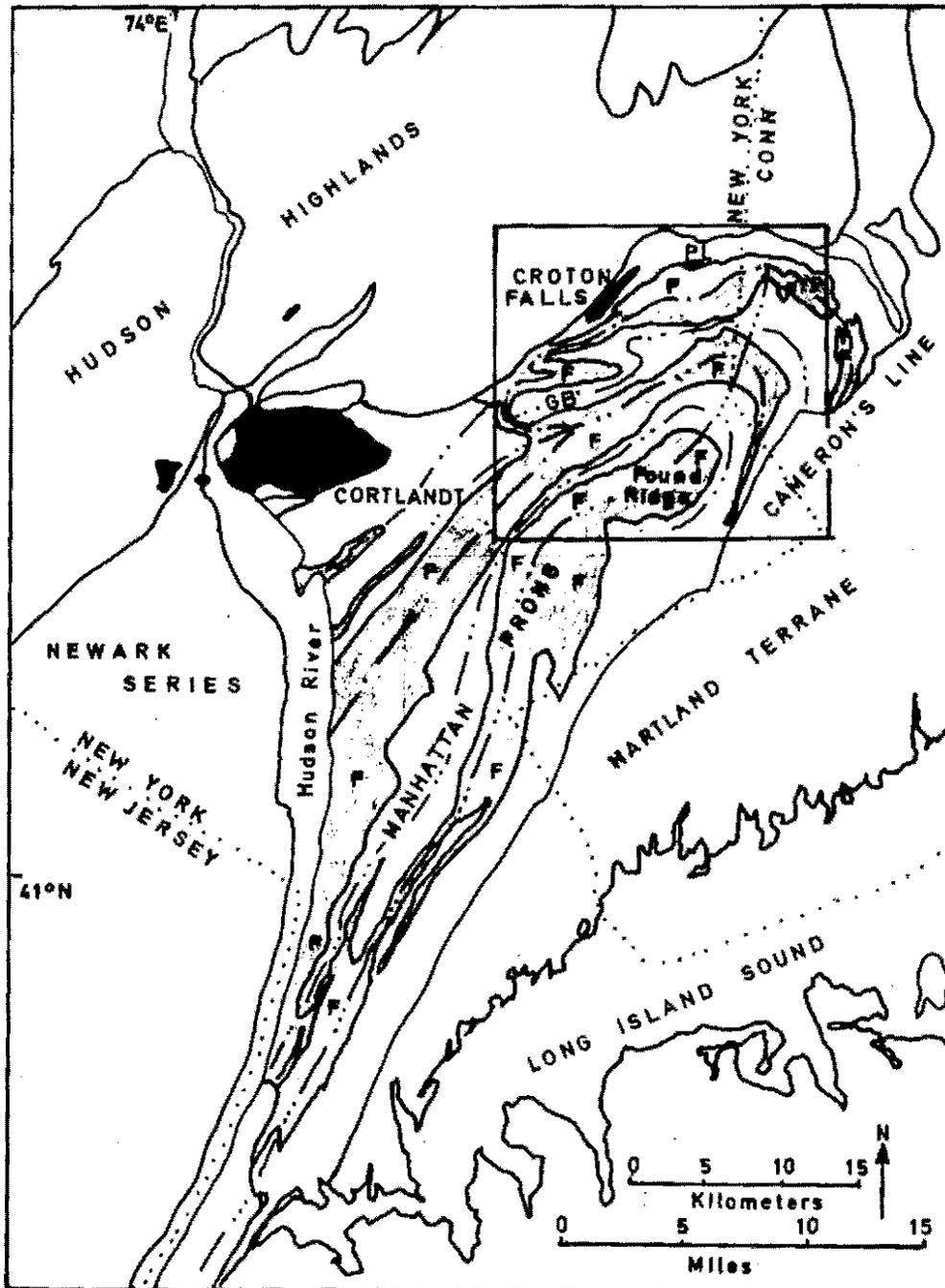
\* All Rb-Sr and K-Ar ages have been recalculated using the recommended decay constants of Steiger and Jager, 1977.

FIGURE 1: Location Map showing the Manhattan Prong in relation to surrounding rocks.

Mafic complexes are shown in black. Fordham Gneiss is marked with F's.

PL = Peach Lake mafic complex. GB = Goldens Bridge.

Box encloses the area shown in more detail in figure 2.



it and the Yonkers Gneiss are best interpreted as metarhyolite bodies (an assessment we agree with). Mose, Eckelmann, and Hall (1979) subsequently favored a plutonic origin for the Yonkers on the basis of zircon morphology studies.

The stratigraphic position of the irregularly present Lowerre Quartzite, between the Fordham Gneiss and Inwood Marble, was established by Norton (1959). It has been correlated with basal sandstones of late Precambrian to early Cambrian age such as the Dalton and Poughquag Quartzites. The Inwood Marble is thought to represent carbonate bank deposition, and is correlated with the Cambro-Ordovician Stockbridge Group. The regional Mid-Ordovician unconformity that truncates the carbonate bank rocks was traced into the area by Hall (1966). Marble is found near the base of the Manhattan Schist, which overlies the unconformity; pelmatozoa from this marble, thought to be Mid-Ordovician in age, have been described by Ratcliffe (1968b) and Ratcliffe and Knowles (1969). The basal unit of the Manhattan Schist, consisting primarily of calcitic marble and sulphitic schist, has been correlated with the Mid-Ordovician Walloomsac Formation of Dutchess County and western Connecticut. Thus, the general geologic relationships found on the Lower Paleozoic platform of eastern North America are all evident in the Manhattan Prong: Grenvillian basement; late Precambrian igneous rocks; basal sandstone; carbonate bank sequence; unconformity; Mid-Ordovician calcitic carbonates; and overlying pelite (Figure 2).

Hall (1968b), making an analogy with the Taconic area to the north, proposed that the structurally higher Manhattan units ("B" and "C") were allochthonous, emplaced as early thrust slices or gravity slides. By this reasoning, Manhattan B and C should be Cambrian or older in age, and may be comparable to the Nassau Formation, the Everett Schist, and/or the Hoosac Formation. Some additional support for this point of view is derived from the Rb-Sr ages of  $554 \pm 49$  m.y. yielded by upper Manhattan metasediments (Mose and Merguerian, 1985). Hall's interpretation has been adopted on the new Connecticut state map (Rodgers, 1982), where the lower Manhattan (unit "A") is shown as the Walloomsac Formation, and only the upper units ("B" and "C") are still called Manhattan Schist. In this study, the single designation Manhattan Schist has been retained because Walloomsac-type rocks have been confirmed at only a few scattered locations and we are not yet convinced of their continuity in this area. The bulk of the Manhattan Schist present here belongs to the upper (?allochthonous) unit.

Although Hall (1968b) subdivided rocks of the Fordham Gneiss into traceable map units in the Glenville and White Plains areas, we have not yet attempted to trace stratigraphy in the Precambrian except for those rocks that lie immediately below the late Precambrian "unconformity". Here, some intriguing relationships have been recognized:

1. As previously discussed, large bodies of igneous rock of very late Precambrian age are found a short distance below the "unconformity". We feel these rocks are probably metarhyolites and we have found extensive amphibolites in a similar stratigraphic position in several places.

2. A sequence of feldspathic wackes is found below Lowerre Quartzite and Inwood Marble in a number of places. These rocks show no evidence of Precambrian deformation, and, indeed, appear conformable with the overlying Paleozoic strata. In several places amphibolite (basalt flows?) is inter-layered with the metasedimentary rocks.

FIGURE 2. Geologic Map of the Croton Falls and Peach Lake sheets, and adjacent areas modified from Prucha et al., 1968, Fisher et al., 1970, Alavi, 1977, and Rodgers, 1982

Explanation of Symbols

- = Stop location
- x = Age determination location

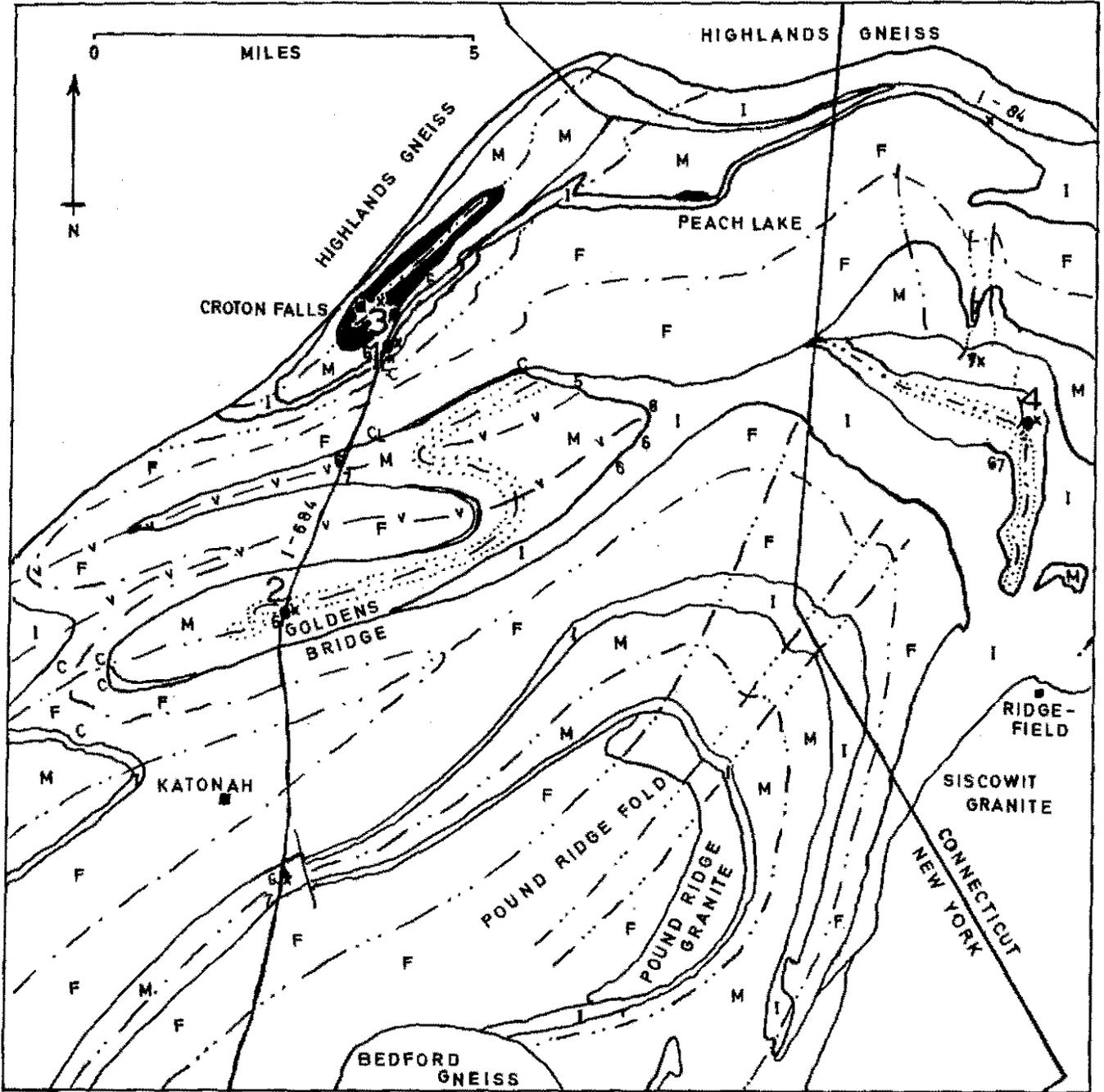
DEFORMATIONAL EVENTS

- 7 = Zone of D<sub>7</sub> thrusting
- 6 = Zone of D<sub>6</sub> ductile shearing
- v -- v -- v -- F<sub>5</sub> Axial Trace

- |             |                            |   |
|-------------|----------------------------|---|
| — ... — ... | F <sub>3</sub> Axial Trace | Note: These axial traces are derived partly from mapping, and partly from construction of cross-section of the Manhattan Prong. They are required by the cross-sections, but their geographic locations are in places only approximate. |
| — .. — .. — | F <sub>2</sub> Axial Trace |   |
| — . — . —   | F <sub>1</sub> Axial Trace |   |

ROCK UNITS

Granites (Too small to show on this scale) Croton Falls and Peach Lake Mafic Complexes ..... Schematic representation of Hartland Formation along the F <sub>2</sub> axial trace that we think is responsible for its presence M = Manhattan Schist (Not subdivided) I = Inwood Marble L = Lowerre Quartzite C = Slabby-weathering unit underlying Lowerre Quartzite F = Fordham Gneiss Highlands Gneiss	Devonian and Mississippian ? Mid-Ordovician ?Cambro-Ordovician b+c: ? Cambrian a: early Mid-Ordovician Cambro-Ordovician Early Cambrian ? Late Precambrian or early Cambrian Grenvillian
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B3-5

We believe the simplest interpretation for these rocks to be that they were deposited in a late Precambrian rift basin, perhaps before the opening of Iapetus. The association of feldspathic clastic sediments with a suite of volcanic rocks suggests a rifted margin, and has parallels in some Taconic allochthons (e.g., in the Nassau and Hoosac Formations) and elsewhere in the Taconics (e.g., in Newfoundland; Strong and Williams, 1972). Ratcliffe has found probable Catocin age mafic dikes in the Hudson Highlands (Ratcliffe, 1983), and has recently determined that the geochemistry of the mafic volcanic rocks in the Nassau Formation suggests a continental rift environment (Ratcliffe, 1985).

Another stratigraphic "package" we have identified has structural rather than paleoenvironmental implications. Rocks of distinctive appearance have been recognized within "Manhattan" belts in three localities; these rocks form a suite including interbedded schist and granulite (resembling turbidites), thick beds of fine-grained quartzo-feldspathic metasediments, and amphibolites. These rocks bear a striking resemblance to parts of the "Hartland Terrane", and are here mapped as Hartland; they are distinct from the poorly bedded, more pelitic Manhattan Schist. Hartland rocks have been interpreted as eugeosynclinal, and the boundary between them and North American shelf and slope facies (Cameron's Line) has been interpreted as a zone of major tectonic convergence (e.g. Rodgers, Gates, and Rosenfeld, 1959). Recently, Cameron's Line has been interpreted as a ductile fault deep within a west-facing accretionary prism, along which North American shelf and slope facies and Hartland rocks were juxtaposed (Merguerian, 1983). Thus these previously unrecognized occurrences of Hartland have tectonic implications that must be considered in reconstructing the Taconic history of the Manhattan Prong.

#### LITHOLOGIC DESCRIPTIONS OF THE ROCK UNITS IN THE CROTON FALLS AREA

The rock units present in the Croton Falls / Peach Lake map areas include the Highlands Gneiss, Fordham Gneiss, Pound Ridge Granitic Gneiss (and related rocks), Lowerre Quartzite, Inwood Marble, Manhattan Schist, and outliers of the Hartland Formation. In addition there are intermediate to ultramafic intrusive rocks of the Croton Falls and Peach Lake Complexes, and numerous granites of various ages.

Each of the formations contains a wide variety of rock types (see Table 1), and all have been repeatedly deformed and metamorphosed. Their present appearance is commonly more dependent on their local tectonic/metamorphic history than on their original characteristics. Positive identification of the formation to which the rocks of a particular outcrop belong usually rests on the proportions of the different rock types present. Thus, identification can be problematic for small outcrops and small fault-bounded blocks. What follows is a brief summary of what appear to be the most useful field, textural, and mineralogical criteria for distinguishing between the formations in the Croton Falls area.

TABLE 1  
Estimated Proportions of Various Rock Types in the Manhattan Prong

XXX = dominant    xx = major    x = minor

Rock Type	Highlands Gneiss	Fordham Gneiss	Pound Ridge Granitic Gneisses & Related Rx	Lowerre Quartzite	Inwood Marble	Manhattan Lower unit	Schist Upper unit	Hartland Formation
Pelites	xx	x		x	x	XXX	XXX	xx
Wackes	x	x	x	x	x	xx	xx	xx
Quartz-free Calc-Sil	x	x			XXX	x		xx
Qtz-bearing Calc-Sil	x	x			xx	xx		xx
"Limy Sandstones"	x	x				xx		xx
Arkosic Rocks	xx	xx	xx	XXX	x	xx	xx	xx
Quartzo- Feldspathic Rocks	XXX	XXX	xx	x		x	x	xx
Intermediate Rocks and Amphibolites	xx	XXX	xx				x	xx
Mafic Rocks	xx	x	x				x	x
Ultramafic Rocks	x	x						
Granitic Rocks	x	x	x		x	x	x	x

## Proterozoic Y Units (Grenvillian)

The Highlands Gneiss includes meta-igneous and meta-sedimentary rocks of various compositions that were metamorphosed under granulite facies conditions. Although they were somewhat remetamorphosed and deformed during the later Paleozoic orogenic events, they were not as modified as were the rocks of the Manhattan Prong. Many of the rocks in the Highlands are massive, relatively homogeneous, and relatively poorly foliated. Some still look recognizably igneous (for instance, some of the pyroxene-bearing granites). However, the Highlands Gneiss also includes layered quartzo-feldspathic gneisses, amphibolites, occasional small calc-silicate layers, and belts of sulfide-bearing pelitic schist.

The Fordham Gneiss is generally more distinctly layered and foliated than the Highlands Gneiss. These differences are attributed to the fact that the Fordham Gneiss has been repeatedly reformed and remetamorphosed during the Paleozoic. Relict pods or blocks of homogeneous rocks that still retain their granulite facies assemblages are present throughout the Fordham, but they are relatively small, and cannot be traced for more than a few tens of feet. The layering in the Fordham Gneiss is commonly irregular, and in many places appears to be related to migmatization associated with the Paleozoic metamorphism: Coarse-grained quartzo-feldspathic leucosomes are separated from the neighboring hornblende-bearing gneiss by biotite-rich selvages. Where significant biotite is present, the rocks can be distinctly schistose. Layered quartzo-feldspathic gneisses predominate, but the Fordham Gneiss also includes amphibolites, pelitic and calc-silicate assemblages, and relict blocks of mafic to ultramafic rocks.

## ? Late Precambrian to Early Cambrian Units

The Pound Ridge Granitic Gneiss is relatively homogeneous and strongly foliated. Around its edges the rocks are migmatized (Scotford, 1956). Associated with the granitic gneiss are numerous layers of relatively fine-grained, distinctly foliated amphibolite. Other rocks that appear to lie stratigraphically below the Paleozoic metasedimentary rocks and above the Grenvillian basement include migmatized amphibolites, various quartzo-feldspathic rock types, and a few calc-silicate and slightly pelitic rocks. They differ from the underlying Fordham gneiss in their slabby fracture pattern: layers are more continuous, boundaries are straighter and, after exposure, the rocks tend to weather and break along the planar lithologic boundaries. Where the quartzo-feldspathic members have been migmatized or granitized the resultant rock is relatively homogeneous granitic gneiss (in contrast to the lit-par-lit layering that characteristically develops in the Fordham Gneiss). Migmatization of the amphibolites commonly produces a brecciated rock.

A detailed study of these rocks is being undertaken, and a more complete description of their characteristics will become available when the study has been completed. At the moment our preferred interpretation is that they and the associated Pound Ridge Granitic Gneiss represent a meta-sedimentary/ metavolcanic suite that underlies the Inwood Marble and unconformably overlies the Grenvillian Fordham Gneiss. Earlier workers have placed the unconformity immediately below the sporadically distributed outcrops of Lowerre Quartzite or Dalton Formation: we feel that it should be placed below the Pound Ridge Gneiss and its related rocks.

At stop #1 Lowerre Quartzite overlies the slabby-weathering unit described above. It is thin to massively bedded, and characteristically retains its fine, even grain size through all the ensuing metamorphism. It is commonly arkosic, and the feldspar is almost entirely potassic. (In contrast, feldspathic quartzites in the lower Manhattan Schist contain little if any K-feldspar. Note: this generalization is based on examination of several dozen etched and stained chips of Lowerre and Manhattan quartzites from many localities supplemented by a few thin sections.)

#### Cambro-Ordovician Units

The Inwood Marble consists largely of foliated, impure dolomitic marbles, with locally well-developed layering. The layers contain different proportions of the calc-silicate minerals olivine, diopside, tremolite, phlogopite, scapolite, and/or epidote. The marbles are irregularly distributed, and the thickness of the unit varies significantly. Greater than normal thickness is present on many fold noses, and the unit is completely missing along some fold limbs.

The Manhattan Schist is a composite unit: the lower part (the Walloom-sac Formation) consists of interbedded sulfide-bearing pelites, fine-grained quartz-rich rocks with calc-silicate affinities, and occasional impure marble layers. The upper (probably allochthonous) unit, is more extensive, more homogeneous, less layered, and generally richer in quartz and feldspar than the lower. In many places, metamorphic segregation has produced distinct, if irregular layering in which quartzo-feldspathic leucosomes and extremely aluminous melanosomes (or "restites") have taken the place of the more homogeneous quartzo-feldspathic pelites. Scattered fine-grained amphibolites are present at or near the base of this unit.

The Hartland Formation as we see it in the Croton Falls area, consists of generally distinctly layered fine-grained quartzo-feldspathic granofelsic rocks interlayered with pelitic schists and occasional fine-grained foliated and/or lineated amphibolites. In places the interlayered granofelses and schists appear to reflect original graded bedding. Elsewhere extensive granofelses are present in which only minor intercalated pelites are present.

#### Intrusive Rocks of Paleozoic Age

The rocks of the Croton Falls and Peach Lake Complexes range in composition from dunite, pyroxenite, and pegmatitic hornblendite through gabbro and diorite, to perthite syenite. Intrusion of the ultramafic rocks produced emery (corundum-magnetite+garnet) assemblages in the adjacent Manhattan Schist, and in xenoliths of the schist that were incorporated into the magma.

Granites of several different ages intrude the rocks of the Manhattan Prong. The youngest ones are the easiest ones to recognize: they contain primary muscovite and/or tourmaline, and in several places contain distinctly zoned plagioclase ( $An_{32}$  to  $An_{20}$ ). They range from fine grained to pegmatitic, are generally rich in K-feldspar, and commonly produce muscovite-rich selvages in the adjacent host rocks. In many places tourmaline is present along the contacts or extending out from the contacts parallel

to the older schistosity. In a number of places kyanite and staurolite concentrations or pods up to 4 inches long have formed in the host rocks and along fractures in the host rocks adjacent to the granite intrusions. Although they have been deformed in places by the latest two deformational events, they do not generally look as deformed and recrystallized as the older granites. Young granites from three localities have been dated by the Rb-Sr isochron method. They give ages of  $335 \pm 13$  m.y. at the Trattoria outcrop (Stop #1),  $343 \pm 9$  m.y. at the Goldens Bridge outcrop (Stop # 2), and  $358 \pm 9$  m.y. at an outcrop near Katonah (see Figure 2), (Brock et al., 1985).

One set of older granites that intrudes the Croton Falls mafic rocks has been dated by the Rb-Sr isochron method at  $387 \pm 37$  m.y. (Brock and Mose, 1979). These granites contain no tourmaline and no primary muscovite, contain more plagioclase, and are generally more zoned, more layered, and more foliated than the younger granites. Where the older granites intrude the emeries in the xenoliths in the Croton Falls Complex, they produce biotite-garnet selvages in contrast to the muscovite selvages produced by the younger granites in the same area.

Additional older granites are present in other parts of the area, but their age relationships are not as certain, and they have not yet been dated. Like the Croton Falls granites, they contain neither tourmaline nor primary muscovite.

#### Rock Compositions based on Mineral Proportions

In order to compare mineralogical characteristics of the different formations, the rocks were subdivided into compositional categories, and the mineralogy of rocks of similar composition from each formation were compared. (See Table 2.)

The compositional categories used in Tables 1 and 2 are intended to approximately follow distinctive primary compositions, but since the extensive metamorphism can have modified those compositions either by segregation, anatexis, or metasomatism, the mineralogical categories are not perfectly analogous to their original counterparts. In general, anatexis will have reduced the proportion of quartz and/or feldspar in the rocks, and the resulting "restite" will be relatively rich in garnet, sillimanite, biotite and other Fe-Mg- +Al minerals that are not readily taken up by the granitic melt. The ultimate "restites" are the emery assemblages in the xenoliths in the Croton Falls Complex, and the garnet-sillimanite-biotite-magnetite rocks with virtually no quartz and feldspar that are found in the regionally metamorphosed pelites elsewhere in the area.

Opaque minerals are not separated on the Table 2. In most cases magnetite (or titanomagnetite) predominates, but in a number of cases reflected light shows several black opaque phases to be present together with some complex intergrowths. In the emery deposits a number of Fe-Ti oxide phases are present. These include members of the ilmenite-hematite and magnetite-ulvospinel solid solution series'. Primary pyrite is present in the sulfidic schists, and secondary pyrite is common on and alongside late fractures in many rocks.

The estimates of percentages of minerals present are based on thin sections (and some hand specimens) selected to be "representative" and to show the ranges of composition present. The percentages have been rounded because in many of these rocks percentages vary both along and across strike, and even a precise determination of mineral percentages in a particular thin section can serve as no more than a gross approximation of those in adjacent rocks.

The rules used to classify the compositional categories in thin sections are outlined below.

### I Rocks Containing Alumino-Silicate Minerals

>5% Alumino-silicates = Pelite

0.1-5% Alumino-silicates = "Wacke" [Note: some of the "wackes" contain less than 30% quartz, which would be unusual in normal wackes. It is likely that some quartz was lost by anatexis, but that the low, but significant alumino-silicate content still indicates a wacke origin].

### II Rocks Containing Carbonate and/or Calc-Silicate Minerals

>15% Carbonate or calc-silicate minerals:-

a. Quartz-free Marble or Calcsilicate Rock

b. Quartz-bearing Marble or Calcsilicate Rock

[The normal calc-silicate minerals in these rocks are diopside, tremolite, scapolite, phlogopite, epidote, and greater than normal amounts of sphene: chondrodite and vesuvianite are present in Fordham marbles in one locality].

0.1-15% carbonate and/or calc-silicate minerals (and usually relatively quartz-rich) = "Limey Sandstone"

### III Rocks Containing Neither Alumino-Silicate nor Calc-Silicate Minerals

>30% Quartz and no alumino-silicate or calc-silicate minerals = Arkosic Rock [As in the "wackes", the relatively low silica content is attributed to segregation or anatexis].

15-30% Quartz = Quartzo-Feldspathic Rocks

a. with >5% K-Feldspar

b. with <5% K-Feldspar

5-15% Quartz = "Quartz-bearing Amphibolites and other Intermediate Rocks"

0.1-5% Quartz, and relatively sodic plagioclase ( $<An_{40}$ ) = Amphibolites and other Intermediate Rocks

<5 % Quartz, and relatively calcic plagioclase ( $>An_{40}$ ) = Mafic Rocks

<5% Quartz and <5% Feldspar = Ultramafic Rocks

TABLE 2  
MINERAL CONTENTS OF EQUIVALENT ROCK TYPES IN THE DIFFERENT FORMATIONS  
IN THE CROTON FALLS AREA  
(See explanatory notes at the bottom of the third page of the table)

ARKOSIC ROCKS																	
Formation	Quartz	K-Felds	Perth	Plag	Oliv	0-Px	C-Px	0-Amp	Cum	Hbl	Biot	Muscovite	Chlor	Gar	Epi	AL	Op
												Early ?	Late				
Fordham n=9	30-60 0	0-60 0	0	15-60	0	0-15 0	0-15 0	0	0-30 0	0-15	1-30 15-30	0	0	0-15 0	10-15	10-15	.1-5 .1-1
Highlands n=2	30-60	0-1	0	15-60	0	0	0	10	0	0-5	15-15	0	0	0	10-1	10-1	0-1 1-30
Hartland n=4	30-60 .1-1	0-1 0	0-1 0	5-60 30-60	0	0	0	10	0	0	15-30	0-15 0	0-15 0	0	10-5 0	10-15	.1-5 .1-1
Manhattan n=12	30-90 30-60	0	0	.1-60 15-30	0	0	0	10	0	0-15 0	15-30	0-60 0	0-30 0-5	0	10-15 15-15	10-5 0	0-15 1-5
Lowerre n=1	40	35		1							15	2	1	16			1
QUARTZ-FELDSPATHIC ROCKS WITH MORE THAN 5% K-FELDSPAR																	
Formation	Quartz	K-Felds	Perth	Plag	Oliv	0-Px	C-Px	0-Amp	Cum	Hbl	Biot	Muscovite	Chlor	Gar	Epi	AL	Op
												Early ?	Late				
Fordham n=6	15-30	5-30 5-15	.1-5	5-60 30-60	0	0-15 0	0-15	10	0	.1-30 5-15	10-15 15-15	0	0-1 0		10-15 0	0-1 0	.1-5 1-5
Highlands n=7	15-30	15-60	0-5 0	5-60 15-30	0	0	0	10	0	0-15 0	1.1-30 15-15	0-5 0	0-1 0		10-5 11-5	0-1 .1-1	0-5
Hartland n=1	20	35		15	0	0	0	10	0	0	15	0	0	15			.5
Manhattan n=3	15-30	5-90	0-60	1-30 15-30	0	0	0	10	0	0	10-30	0-1 .1-1	0-15 0	0-15 0	10-5 0		0-5 1-5
QUARTZ-FELDSPATHIC ROCKS WITH LESS THAN 5% K-FELDSPAR																	
Formation	Quartz	K-Felds	Perth	Plag	Oliv	0-Px	C-Px	0-Amp	Cum	Hbl	Biot	Muscovite	Chlor	Gar	Epi	AL	Op
												Early ?	Late				
Fordham n=16	15-30	0-1 0	0-5 0	15-60	0	0-60 0-15	0-60 0-15	10	0	0-30 5-15	10-60 15-30	0	0	10-15	10-1 0		0-30 1-5
Highlands n=6	15-30	0-5 0	0	15-60 30-60	0	0	0	10	0	0-30 1-30	15-60 15-30	0	0	10-5 0	10-15		.1-5
Hartland n=3	15-30			30-70 30-60	0	0	0	10	0	0	15-30	0	0-15	10-1 0	10-5 0	0-1 .1-1	.1-5 1-5
Manhattan n=3	15-30	0	0-1 0	1-60	0	0	0	10	0	0	15-30	0	0-30 0	11-15 15-15	10 0	0-1 0	1-15
AMPHIBOLITES AND RELATED ROCKS WITH LESS THAN 5% QUARTZ																	
Formation	Quartz	K-Felds	Perth	Plag	Oliv	0-Px	C-Px	0-Amp	Cum	Hbl	Biot	Muscovite	Chlor	Gar	Epi	AL	Op
												Early ?	Late				
Fordham n=7	0-5 1-5	0-1 0	0-5 0	5-60 30-60	0	0-15 0	0-30 0	10	0-30 0	5-90 30-90	1-30	0	0	0-5 0	10	10-1	0-5 1-5
Highlands n=6	0-5	0-5 0	0	30-90	0	0	0-30 15-30	10	0	1-60 5-30	1.1-30 15-30	0	0	0	10	10-5 0	.1-5 1-3
Hartland n=2	1-5			15-60	0	0	0	10	0	30-90	1-5	0	0	0	10		1-15
Manhattan n=4	.1-5 1-5			15-30	0	0	0	10	0	0-90 60-90	1.1-90	0	0	0-5	10	10-15 0	0-5 1-5

## AMPHIBOLITES AND RELATED ROCKS WITH 5-15% QUARTZ

Formation	Quartz	K-Felds	Perth	Plag	Oliv	C-Px	Amph	Cum	Hbl	Biot	Muscovite Early ?	Late	Chlor	Gar	Epi	AL	Op	
Fordham n=16	15-15 0-1	0-30 0-5	0-15 0-5	5-60 30-60	0	0-60 5-15	0-30 5-15	10	0-15 0	0-30 1-5	11-60 15-30	0	0	0-5 0	10-15 0	10-15 0	0-5 1-5	
Highlands n=4	15-15	0-30	0-30	5-60 30-60	0	0	0-30	10	0	0-30	11-30	0	0		10-1 0	10-5 0	0-1 0	0-5 1-5
Hartland n=2	5-30	0	0	5-30	0	0	0	10	0	30-60	11-1	0	0	0	10	11-15	0	1-15
Manhattan n=5	15-15	0	0	5-30 5-15	0	0	0-15 0	10	0-15 0	0-60 0	10-30 15-30	0-15 0	0-15 0		11-15 15-15	10-1 0		1-15 1-5

## MAFIC ROCKS

Formation	Quartz	K-Felds	Perth	Plag	Oliv	C-Px	Amph	Cum	Hbl	Biot	Muscovite Early ?	Late	Chlor	Gar	Epi	AL	Op	
Fordham n=1	0	0	0	2	0	0	0	10	0	90	15	0	0	1			2	
Highlands n=8	0-5 1-5	0-5 0	0	5-60 30-60	0	0	0-60	10	0	5-90 30-60	10-30 0	0	0		10-15 0	10-1 0	0-1 0	1-15 1-5
Hartland n=1	2	0	0	40	0	0	.5	10	0	60	10	0	0	0	1.5		.5	
Manhattan n=2	1-5	0	0	15-30	0	0	0	10	0	60-90	11-15	0	0	0	10			1-15

## QUARTZ-FREE CARBONATE AND CALCSILICATE ROCKS

Formation	Quartz	K-Felds	Perth	Plag	Oliv	C-Px	Hbl	Biot	Musc	Chlor	Gar	Epi	AL	Scap	Carb	Sph	Op	Serpi
Fordham n=1					18	15	15	12		4					45		.5	10
Hartland n=	x		x		x	x					x	x	x	x	x	x		
Inwood n=6	0	0-1 0	0	0-1	11-30 11-5	0-30 1-5	0-1 0	5-30 15-30	0-15 0			0-1 0			30-90		0-1 1-1	1-30 5-15
Manhattan n=1	0	0	0	0	10	.5	0	10		.5					80		.5	

## QUARTZ-BEARING CARBONATE AND CALC-SILICATE ROCKS

Formation	Quartz	K-Felds	Perth	Plag	Oliv	C-Px	Hbl	Biot	Musc	Chlor	Gar	Epi	AL	Scap	Carb	Sph	Op	Serpi
Fordham n=2	10-40	5-25		0-10	0	10-40	12-10	10-5						0-30	.5-10	0-5	0-1	
Highlands n=2	15-30	0-30	0-5	5-30	0	15-60	11-5	0-15				10-30				1-5	1-5	
Inwood n=4	1-15	5-90 5-15	0	0-15 5-15	0	5-30 15-30	1-15 5-15	0-15 0				0-15		0-5 0	.1-60 15-30	0-15 1-5	.1-5	
Manhattan n=10	1-30 5-30	0-30 0+10	0	5-60 5-30	0	5-60 5-30	1-30 5-15	0-30 5-15	0-1 0	0-15 0		0-15 5-15		0-30 0-15	0-15 0	.1-15 1-5	.1-15 1-5	

## "LIMEY SANDSTONES": relatively Quartz-rich rocks containing calc-silicate minerals

Formation	Quartz	K-Felds	Perth	Plag	Oliv	C-Px	Hbl	Biot	Musc	Chlor	Gar	Epi	AL	Scap	Carb	Sph	Op
Fordham n=1	25	15		15		15	13	13				15				8	.5
Highlands n=1	35	30		.5		15	15					14				3	2
Hartland n=1	30	20		20				120				17				2	.5
Manhattan n=6	5-60 15-30	0-15 0	0-1 0	15-30		0-15 5-15	1-30 1-5	5-30 15-30				0-15 1-15	0-5 0			0-5 0	0-15 1-5

## PELITES

Formation	Quartz K-Felds Perth Plag				O-Amp	Biot	Muscovite		Chlor	Gar	Cord	Staur	Kyan	Sillimanite			Op
							Early ?	Late						Fib	Acic	Pris	
Fordham n=7	5-60 15-30	0-60 5-15	0-15 5-15	1-30		15-30 15-30		0-30 0	0-15 0	5-30 5-15			0-5 0	1-30 5-15	.1-30 1-5	0-5 1-5	1.1-5 1-5
Highlands n=1				30									20		.5		1.5
Hartland n=4	.1-60		0-5 0	.1-30		115-60 115-30	0-30 0	0-30 5-15		.1-30 .1-1		0-15 0	0	0-15 5-15	.1-5 .1-1	0-5 0	1.1-15 1-5
Manhattan n=47	0-60	0-15	0-90 .1-15	0-60		10-60 11-30	0-1 0	0-30 0	0-30 1-15	0-15	0-60 5-60	0-60 0	0-15 0	0-90 0	0-60 1-15	0-60 .1-30	0-15 1-5

## MANHATTAN PELITES

Assemblage	Quartz K-Felds Perth Plag				O-Amp	Biot	Muscovite		Chlor	Gar	Cord	Staur	Kyan	Sillimanite			Op
							Early ?	Late						Fib	Acic	Pris	
Cord-bearing n=6	1-90 15-30	0-5 .1-1	0-30 5-15	0-30	0-5 0	1-30 5-15	0	0-15 1-5	0-5 0	5-60 15-30	.1-60	0	0-1 0	0-60 5-15	1-15 1-5	0-60	.1-15 1-5
Perthite- bearing n=42	1-60 5-30	.1-15 .1-1	.1-90 5-60	.1-15	0	1-60 115-30	0	0-15 1-5	0-15 0	0-60 5-60		0	0-1 0	0-60 .1-1	.1-15 1-5	0-30 1-30	0-5 1-5
K-feldspar- sill n=15	1-60 15-60	.1-15	0	1-60 5-30	0	1-60 5-30	0-30 0	0-15 5-15	0-15 0	1-30 15-30		0	0	0-15 .1-1	.1-15 5-15	0-30	.1-5 1-5
Kyanite- bearing n=17	1-90 5-30	0-5 0	0-30 0	.1-60 5-30	0	1-30 5-30	0-1 0	0-30 .1-1	0-1 0	.1-60 5-15		0	.1-90 1-15	0-15 5-15	.1-15 0-5	0-30	.1-15 1-5
Staurolite + kyanite n=14	0-60 1-30	0	0	0-60 5-15	0	.1-60 5-30	0-1 0	0-30 .1-1	0-30 0	0-60 15-30		.1-30 5-15	.1-90 5-30	0-15 .1-1	.1-30 1-5	0-15 0	0-15 1-5
Staurolite n=6	1-30 15-30	0	0	1-30 15-30	0	1-30 15-30	0-60	0-60	0-1 0	5-30		1-30	0	0-5 0	1-15 .1-1	0-5 0	.1-15 1-1
Sill-Musc. No ky, st, crd. n=8	1-90	0	0	1-60 15-30	0	1-60 15-30	.1-30 5-15	0-15 0	0	.1-15 5-15		0	0	0-15 0	.1-5 .1-1	0-15	.1-15 1-5
Sill-late Musc n=11	15-60 15-30	0	0	.1-30 5-30	0	5-60 5-30	0	.1-30	0-30 0	5-60		0	0	0-60 1-5	.1-60 1-5	0-30 1-5	.1-15 1-5
Sill. No Musc n=4	1-60 15-30	0	0	0-5	0	.1-60	0	0	0	15-60		0	0	0-30	.1-60	0-60	0-5 1-5

The upper numbers in each box show the maximum range of percentages of the minerals observed.

The lower numbers show the normal range.

## Explanation of abbreviations

K-felds = K-feldspar; Perth = Perthite; Plag = Plagioclase; Oliv = Olivine; O-Px = Hypersthene;  
 C-Px = Clinopyroxene; O-Amp = Gedrite; Cum = Cummingtonite/Grunerite;  
 Hbl = Hornblende (including end-members Tremolite and Hastingsite); Biot = Biotite;  
 Muscovite: Early ? Late: ? = uncertain age relationships, or both early and late Muscovite present;  
 Chlor = Chlorite; Gar = Garnet; Epi = Epidote; Al = Allanite; Scap = Scapolite; Carb = Carbonates;  
 Sph = Sphene; Op = Opaques; Sillimanite: Fib = Fibrous; Acic = Acicular; Pris = Prismatic  
 Cord = Cordierite; Staur = Staurolite; Kyan = Kyanite;

The most striking, and potentially the most useful mineralogical differences, include:-

1. The arkosic and quartzo-feldspathic rocks of the Highlands and Fordham Gneisses routinely contain hornblende and commonly clino- and orthopyroxenes. In contrast the rocks of equivalent composition in the Manhattan and Hartland Formations contain no amphiboles or pyroxenes but only biotite (+garnet). In addition these rocks in the Highlands and Fordham commonly contain more K-feldspar than their equivalents in the younger formations.

2. Amphibolites in the Highlands and Fordham Gneisses tend to be medium grained and unfoliated or only poorly foliated, and generally contain at least some surviving pyroxene. In contrast, amphibolites in the younger formations tend to be finer grained, show distinct planar and/or linear parallelism of the amphiboles, and contain no pyroxene. One exception to this generalization is an amphibolite with calc-silicate affinities in the Hartland Formation (outside the Croton Falls area) that contains diopside and epidote and is only poorly foliated.

3. The Lowerre Quartzite characteristically contains a significant amount of K-feldspar and little or no plagioclase, whereas the sandy units in the overlying Manhattan contain significant plagioclase and little or no K-feldspar.

#### SEQUENCE OF DEFORMATIONAL EVENTS

The sequence of events that we have worked out in the Croton Falls area is summarized in Table 3. Much of our work has been focussed on the Manhattan Schist in order to avoid the added confusion of the pre-Taconian structures.

$F_1$  folds are rarely preserved on outcrop scale. Where they are found they are recognized by the fact that no discernible older foliation is deformed by the folding. In these instances the axial plane foliation ( $S_1$ ) is seen to correspond to the dominant schistosity or foliation in the outcrop, and we therefore assume that in most (but not all) places the dominant foliation that we see being deformed by younger events is the  $S_1$  foliation. Because the earliest events have been overprinted by all the younger deformation and recrystallization, we cannot tell what grade of metamorphism was associated with the  $D_1$  deformation. However, since coarse-grained quartzo-feldspathic segregations are deformed by  $D_2$  deformation, it is clear that the grade of metamorphism was high enough for these segregations to form.

$D_2$  deformation also is only rarely demonstratable in outcrop. One of the clearest examples will be pointed out at stop 1. Here the  $D_2$  is strongly cataclastic, and older coarse-grained quartzo-feldspathic segregations have been sheared into flasers. Elsewhere, isoclinal folds are present that fold the  $S_1$  schistosity and are themselves refolded by the ubiquitous  $F_3$  folding. The grade of metamorphism associated with this folding reached at least sillimanite grade since  $F_3$  folds bend and break pre-existing sillimanite. In the eastern part of the Peach Lake map sheet

TABLE 3. PROVISIONAL GEOLOGIC HISTORY OF THE CROTON FALLS AREA, SOUTHEAST NEW YORK

GEOLOGIC AGE	FOLD EVENT	FORMATIONS, ROCK TYPES, ETC	META EVENT	FACIES/DISTINCTIVE ASSEMBLAGES	RADIOMETRIC AGE (in M.Y.)	COMMENTS
Mesozoic		Faulting, and some folding.				In adjacent areas
	D7	Late thrust faulting.		Muscovite	294±13 309±6	K-Ar on Biotite & Muscovite Seiderman & Brock (unpub. data)
		"Latest metamorphic event."			318±6 329±13	White Plains area, Grauert and Hall, 1974. K-Ar on Biotite, Clark and Kulp, 1968
Carboniferous	D6	Ductile shearing and related folding. Deforms Boldens Bridge & Katonah Bntes.  Boldens Bridge & Katonah Granites. Tourmalinized fractures.	M5	Musc-Staur-Kyan-Gar (+Bil)  Tourmaline-Musc-(+Kyanite) Chlorite-Musc-Gar (Staur)	335-358	White Plains area, Grauert and Hall, 1974. K-Ar on Biotite, Clark and Kulp, 1968 NE/Mod NW. NW-side-down movement  Brock et al., 1984 + Margarite in metamorphosed marbles of the Croton Falls mafic complex. Brock & Brock 1981.
(Acadian)	F5	Close to tight folding. ?Doming. Deforms Croton Falls Granites.	M4	Musc-Kyan-Staur in west rising to Kf-Bil in E & W		Brock and Mose, 1979. + recent field work.
Devonian		Croton Falls Granite.		No Muscovite in pelitic selvages	387±34	Brock and Mose, 1979.
(Taconian2)	F4	Close reclined folds. Quartz-K-Feldspar, and Qtz-Gar-Bil segregations. Deforms F3 segregations.			442	Brock et al., 1985 (F4 segregations)
	F3	Ubiquitous tight to isoclinal folds. Bisial-planar Quartz-K-Feldspar segre- gations. Deforms F2 flaggers.	M3	K-Feldspar-Billimanite		
Mid- to late Ordovician		Croton Falls Complex. Ultramafics + Diorite. Cortlandt Complex. Ultramafics, Norite + "		Emery Emery	423 430±34	Brock and Brock, 1981. Ratcliffe et al., 1982. No-Sm Downnick and Bass, 1981
Mid- Ordovician (Taconian1)	F2	Isoclinal folds, Cataclastic flaggers, Presumed recumbent, associated with major thrusts.	M2	Billimanite rising to Kf-Gar-Cord in E		
	F1	Isoclinal folds. Quartzo-feldspathic segregations, Porphyroblastic Garnet. Presumed recumbent, associated with thrusting/sliding.	(N1)			
Cambro- Ordovician		Manhattan Schist. Upper unit predominantly pelites and wackes, but with some amphibolites near base. Upper unit was probably tectonically emplaced early in the Taconic Orogeny  Basal unit includes carbonates, limy sandstones and graphitic, sulphitic schists.  Unconformity. - - - u - - - u - - - u - - - u - - - u - - - u - - - u - - - u - - -  Inwood Marble. Dolomitic marbles near base, becoming more calcitic near top.  Lower Quartzite. Quartzite and arkosic wackes.			554±29	Hall, 1968. Mose & Marguerian, 1985 lower Middle Ordovician fossils in basal beds Ratcliffe and Knowles, 1968
Proterozoic Z		Yonkers & Pound Ridge Granite Gneisses, and related metasedimentary and metavolcanic rocks  Unconformity - - - u - - - u - - - u - - - u - - - u - - - u - - - u - - -			583±19 563±30 538±43	Mose and Hayes, 1976. Long, 1969. Mose, 1981.
Precambrian (Granvillian)		Highlands & Fordham Gneisses. Two episodes of granitic intrusion.				
	IF(-1)	At least two episodes of folding	M(-1)	Hornblende Granulite?	1914±31	Helenek and Mose, 1976.
	IF(-2)	and metamorphism.	M(-2)	Pyroxene Granulite Hypersthene-K-Feldspar	1165±44	Helenek and Mose, 1976. Brock and Brock, 1983.
Proterozoic Y		Deposition of sediments and volcanics.			1350	Mose, 1982

Radiometric ages are Rb-Sr isochron ages unless otherwise specified

the grade was higher: the flasers include cordierite-garnet-perthite-sillimanite assemblages. Although rocks of this grade have been reported in Grenvillian metamorphism in the Highlands and Fordham Gneisses, and in Acadian metamorphism in south-central Massachusetts (Tracy and Robinson, 1980, and Robinson *et al.*, 1982), this is the first such occurrence that we know of in Taconian metamorphism in the northern Appalachians.

On the map scale, a number of major  $F_1$  and  $F_2$  fold closures and associated thrusts are required in order to explain the map distribution patterns of the stratigraphic units. When we have attempted to reconstruct the geometry of the early folding of the Prong using the structural and stratigraphic constraints that we have collected both here and in other parts of the Prong, as well as those that are available in the literature, a number of solutions have always proved possible. Figures 1 and 2 show the approximate locations of the major  $F_1$  and  $F_2$  closures required in the reconstruction that comes closest to fitting all the facts as we know them. Because the dominant folding in the Croton Falls/Peach Lake region is moderately-plunging upright  $F_3$  folding, a cross section through this part of the Prong looks like a slightly fore-shortened map-view of the area; for this reason, no cross section is presented here. More work is still needed to trace and delimit these early structures in detail. The apparent amplitude of these folds implies that both the  $F_1$  and  $F_2$  were recumbent.

$D_2$  flasers and associated sillimanite prisms are folded and broken by upright, tight to isoclinal  $F_3$  folds. The latter folds are characteristically visible on outcrop scale throughout most of the western half of the map area, and are associated over almost the entire area with K-feldspar-sillimanite grade metamorphism. Generally, the  $S_3$  axial plane cleavage is only weakly developed. At stop 1, the  $F_3$  folds are associated with quartz-K-feldspar segregations that are aligned approximately parallel to the axial planes of the folds. In the eastern part of the area fewer recognizable  $F_3$  folds are present on outcrop scale, because fewer parasitic folds developed on the nose of the major  $F_3$  fold that passes through the Pound Ridge area.

The Croton Falls mafic rocks were intruded after the  $D_2$  deformation but before the  $F_3$  folding: small  $F_3$  folds can be seen folding original igneous layering in previously undeformed (unfoliated) mafic rocks. The dominant foliation in the mafic rocks is parallel to the axial planar foliation in these folds. In addition, emery assemblages that were formed in Manhattan Schist alongside and incorporated within the mafic rocks were locally remetamorphosed during the  $F_3$  deformation prior to the emplacement of the Croton Falls granites.

The  $F_3$  folds at Stop 1 are refolded by close, reclined  $F_4$  folds. These folds commonly have rounded noses on the broad scale, and small-scale chevron folds in pelitic layers as they pass round the noses of the larger folds. A distinct axial plane cleavage is developed in some places, and sillimanite has grown along this cleavage as it passes through the  $S_3$  quartz-K-feldspar segregations. The  $F_4$  folding is relatively restricted in its distribution, and probably has less tectonic significance than any of the other deformational events. However, it is extremely useful since its associated quartz-feldspar-garnet-sillimanite segregations that cross-cut  $F_3$  hinge surfaces, are amenable to Rb-Sr age determination analysis (See discussion of age constraints below).

$S_3$  foliation in the Croton Falls mafic rocks has controlled the emplacement of the Croton Falls granites that give an Rb-Sr isochron age of  $387 \pm 34$  m.y. (Brock and Mose, 1979).

The  $F_5$  fold event is complex. The Croton Falls granites are cut by  $S_5$  cleavage, and in a few places are visibly folded by  $F_5$  folds. Outside the Croton Falls complex the  $S_5$  is sporadically developed in distinct zones through much of the area. In the western part of the area the  $S_5$  is readily recognizable because it is associated with kyanite-staurolite grade metamorphism, and thus produces visible down-grading in the K-feldspar-sillimanite schists. To the north and east, however, the grade of metamorphism associated with it increases to K-feldspar-sillimanite grade, and positive identification of  $F_5$  folds becomes more difficult. However, enough places have been found where a full sequence of events can be worked out, and it can be shown that folding of this age is present, and that it takes several different forms. In the north  $D_5$  ductile shearing deforms  $F_2$ ,  $F_3$ , and  $F_4$  folds, and in the east,  $D_5$  appears to entail doming that is responsible for the inflection in the  $F_3$  axial trace as it passes through the Pound Ridge area, and some large-scale drag folding on the limbs of the "dome" (see Figure 2).

The  $S_5$  cleavage and all older structures are cut by tourmaline, two-mica granites and associated tourmalinized fracture surfaces, and these granites are themselves deformed by two episodes of ductile shearing, one syntectonic with the intrusion of the dikes, and the other later. The dikes are in most cases small (less than 4 feet thick) but a few irregularly-shaped bodies up to 45 feet across are also present. The tourmalinized surfaces, and the majority of the more regular dikes, strike northeast and dip moderately to the northwest, and their emplacement was largely controlled by a fracture system with this orientation. The older shear zones ( $D_6$ ) have a similar orientation and the associated slip has a down-dip component, suggesting an extensional regime. The coarse-grained muscovite-kyanite-staurolite rocks that are associated with these dikes and shear zones indicate that water-rich fluids circulated along the shear zones through an otherwise dried out K-feldspar-sillimanite terrane.

The later thrusting ( $D_7$ ) has similar attitudes, and in many places reuses pre-existing planes of weakness. Where they affect rocks already rehydrated during  $D_6$  deformation, new skeletal staurolite is seen growing in the shear zone. In other cases, however, where dry rocks are deformed, textural changes (to mylonite) are striking, but the associated mineral assemblages are relatively little changed from the original, and it is not clear what pressure and temperature conditions prevailed during the deformation.

The age relationships between the deformational/metamorphic events are preserved because the later events are not evenly developed throughout the area:  $D_6$  and  $D_7$  in particular are developed only in restricted zones separated by large areas that show little or no evidence of young deformation.  $D_5$  is more widespread, but even here, the normal pattern is of older K-feldspar-sillimanite assemblages surviving between the muscovite-bearing  $S_5$  cleavage surfaces and shear zones. Other striking examples of survival of older metamorphic assemblages and their associated textures include the granulite assemblages in the Fordham Gneiss that have survived through all the Paleozoic deformation and metamorphism, and the scattered patches of

emery in the Croton Falls complex that have survived with only minor modification through all the post  $F_2$  events.

Age constraints on the timing of the deformational events are provided by a number of Rb-Sr whole rock isochron age determinations (see Figure 3). A determination on the  $F_4$  quartz-feldspar-garnet-sillimanite segregation that crosscuts  $F_3$  hinge surfaces gives an age of at least 442 m.y. (or 462 m.y. if one sample with very low Sr content is discarded as unreliable). This date restricts the four early deformational events to the Taconic Orogeny (Brock et al., 1985)

The  $D_5$  is bracketed by the  $387 \pm 34$  m.y. Croton Falls granites (Brock and Mose, 1979) and the 335 to 358 m.y. two-mica granites (Brock et al., 1985). Even though the error bars on the Croton Falls granite age are large, the  $D_5$  is restricted to Devonian age.

The the two-mica granites appear to have been intruded during the  $D_6$  deformation: the  $D_6$  fracture surfaces control the emplacement of the two-mica granites and  $D_6$  fabrics are locally developed within in them. The  $D_6$  deformation therefore extends into the Mississippian. The  $D_7$  is clearly younger: it deforms the granites, breaks and bends the associated tourmaline and makes flasers of the kyanite pods. K-Ar age determinations on biotite carried out by Kulp and Long from a zone that we map as a  $D_6$  gives an age of  $329 \pm 13$  m.y., and preliminary analyses by Seidermann and Brock on biotite from two  $D_7$  shear zones give ages of  $309 \pm 8$  and  $294 \pm 13$  m.y. (2 sigma) We hope to get more K-Ar ages from biotites from the shear zones in the near future.

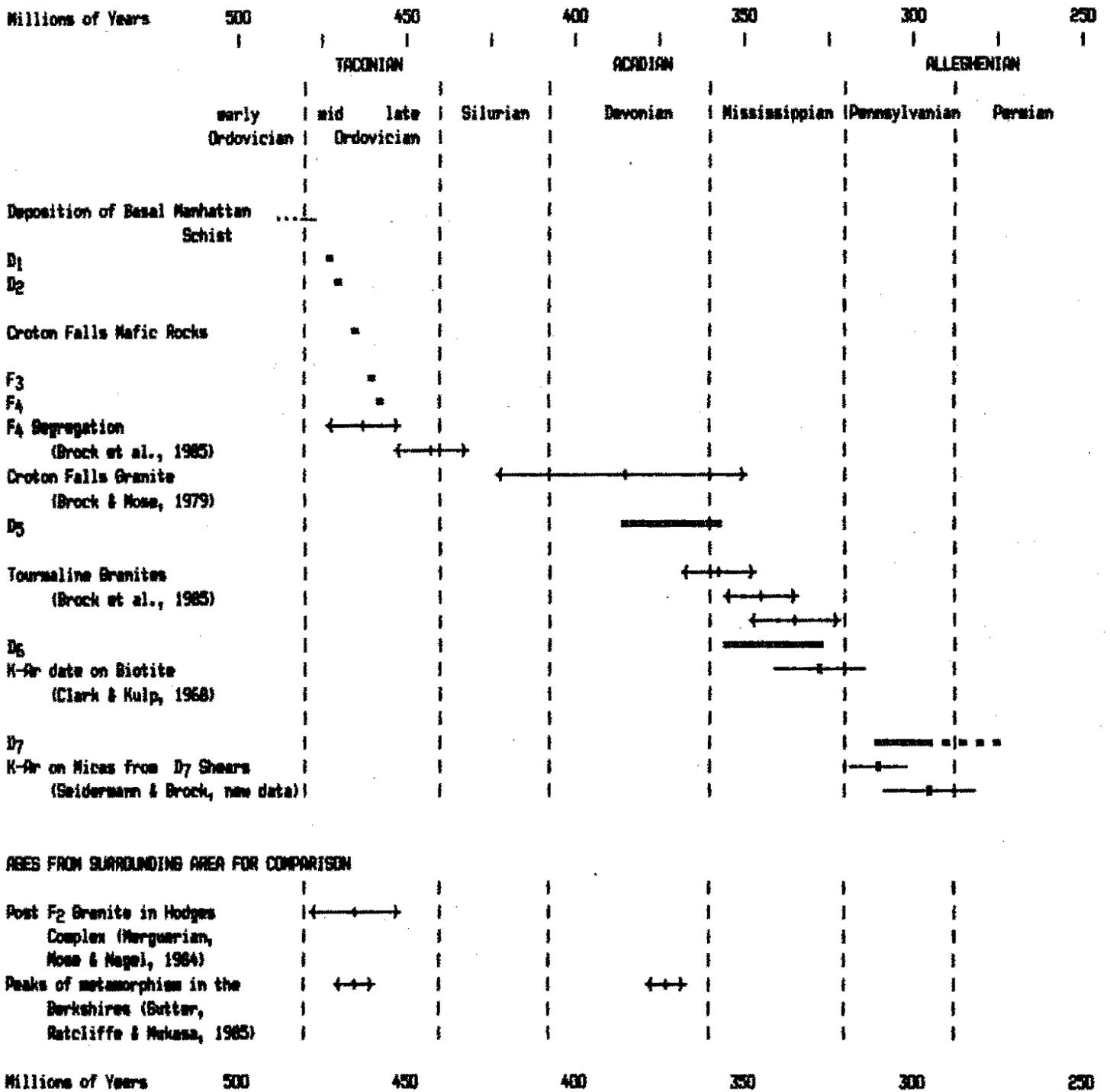
### Summary

The Taconic Orogeny in the Croton Falls area includes two early episodes of large-scale recumbent folding and thrusting whose associated metamorphism reached at least to sillimanite grade in the west and to garnet-cordierite-perthite-sillimanite grade in the east by the end of the second event. The upright  $F_3$  and reclined  $F_4$  folds post-date the emplacement of the Croton Falls mafic rocks, and the associated metamorphism reached K-feldspar-sillimanite grade over all but the southwestern quarter of the map area.

The Acadian Orogeny is represented by the Croton Falls granites and the  $D_5$  deformation and its associated muscovite-kyanite-staurolite grade metamorphism. The deformation of this age includes large-scale upright folds and zones of ductile shearing.

The Acadian Orogeny is followed by ductile shearing and associated intrusion and deformation of the two-mica granites of Mississippian age, which are followed in turn by younger compressional ductile shearing.

FIGURE 3. AGE CONSTRAINTS ON THE DEFORMATIONAL EVENTS IN THE CROTON FALLS AREA



## ROADLOG

Miles from start	Miles from previous point	Remarks
0	0	Kline Geological Laboratory, New Haven
2.3	2.3	Through New Haven following road signs onto route 34
10.0	7.7	Pass through Shelton
22.9	12.9	Turn West on I-684 (towards Danbury & New York)
39.7	16.8	Connecticut-New York Border
41.6	1.9	Outcrop on right of Manhattan Schist with D <sub>5</sub> ductile shear
42.2	0.6	Exit 20. Turn south onto I-684 (towards White Plains & New York)
45.2	3.0	Croton Falls Complex makes up the ridge to the right
	0	Inwood Marble crops out in roadcut on right
46.9	1.7	Turn off on Exit 8. (Hardscrabble Road & Croton Falls)
47.1	0.2	Turn left from exit ramp onto Hardscrabble Road
47.3	0.2	Turn left onto Route 22 south
48.0	0.7	Underpass; route 22 passes under I-684
48.1	0.1	STOP 1. Park beside road just to the east of the underpass.

Stop 1: Climb up on the east side of the I-684 overpass on the north side of route 22. Walk north parallel to I-684 behind the crash-barrier to the first place you can readily climb up on top of the outcrop. Most features we want to see can be adequately seen from the top, and the troopers (and we) are worried about the distracting effects on the drivers in the heavy traffic on the highway of crowds of people at ground level.

Stop 1 encompasses over half a mile of road cuts (see Figure 4) that contain most of the stratigraphic units of the area, and many of the structural relationships we wish to show you. We will be here for several hours.

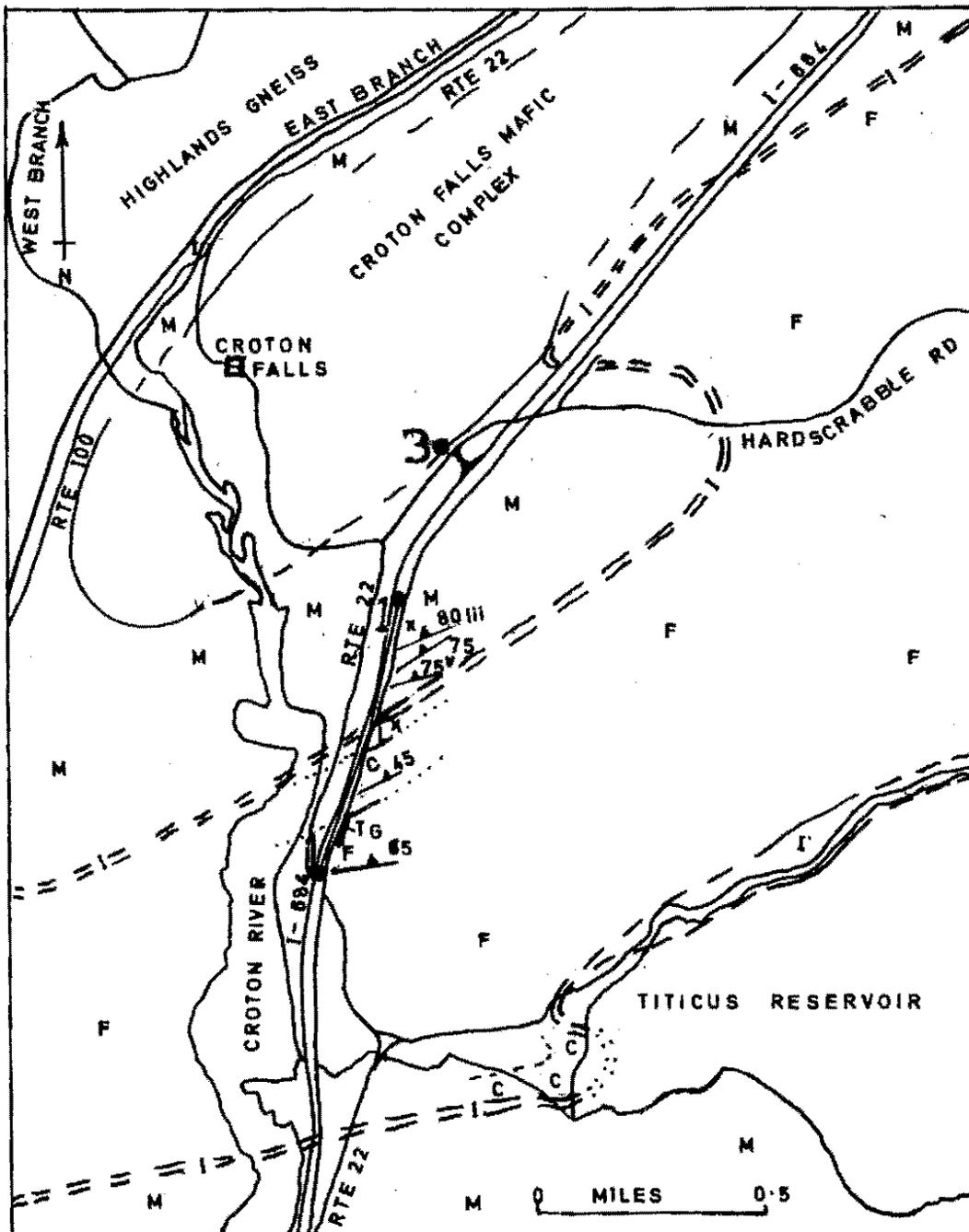
At the southern end of the road cut is a granite of uncertain age. To the north of this granite are layered Fordham Gneisses including various quartzo-feldspathic types, amphibolites, rusty-weathering pelites, calc-silicate assemblages, and pods of ultramafic rock. No granulite assemblages have been recognized here, but good examples are present in this belt of Fordham half a mile to the southwest of here. The quartzo-feldspathic rocks commonly contain hornblende as well as biotite. K-feldspar is present in many, and perthitic feldspar in a few. The amphibolites are generally made up mostly of hornblende and plagioclase, with or without some clinopyroxene relicts, and with various amounts of biotite,

Pelitic rocks in this belt of Fordham to the northeast and southwest of the road cut are coarse-grained and garnet-rich. To the northeast perthite (or K-feldspar) and sillimanite are present, whereas to the southwest, kyanite is present and K-feldspar is absent. Calc-silicate assemblages in this outcrop contain little carbonate, and consist largely of clinopyroxene, clino-amphibole, and some epidote.

FIGURE 4. Geologic map of the area surrounding Stops 1 and 3, modified from the maps of Prucha et al., 1968, and Sneider, 1969.

Explanation of Symbols    o = Stop location    x = Age determination location  
 TG = Two-mica granite + Tourmaline  
 M = Manhattan Schist  
 = = = I = = = Inwood Marble (Shown schematically: it is rarely more than a few tens of feet thick)  
 L = Lowerre Quartzite  
 C = Slabby-weathering unit underlying Lowerre Quartzite  
 F = Fordham Gneiss

Attitudes shown are of the dominant foliation (generally parallel to  $S_1$ ), except where they are accompanied by small Roman numerals indicating the deformational event represented.



Layering is steep, and the most conspicuous upright folds are presumably of Taconian age. (No direct evidence of their age has been found here.)

The northern end of this segment of the road cut is a large two-mica granite.

To the north of the granite, after 30 feet without exposure, is the start of the (?) late Precambrian slabby-weathering unit. It consists of interlayered quartzo-feldspathic gneisses, schistose gneisses, migmatized gneisses, and fine-grained, lineated amphibolites. In some of the fine-grained migmatitic layers the dominant mafic mineral is a blue-green hastingsitic amphibole. In a few places the quartzo-feldspathic rocks are finer-grained than normal, and resemble arkosic quartzites.

The contact between these rocks and the overlying Lowerre Quartzite is sharp; being marked by a change from migmatitic gneissose rock to a fine-grained, rusty-weathering arkosic quartzite. To the north the quartzite becomes more impure, and more schistose.

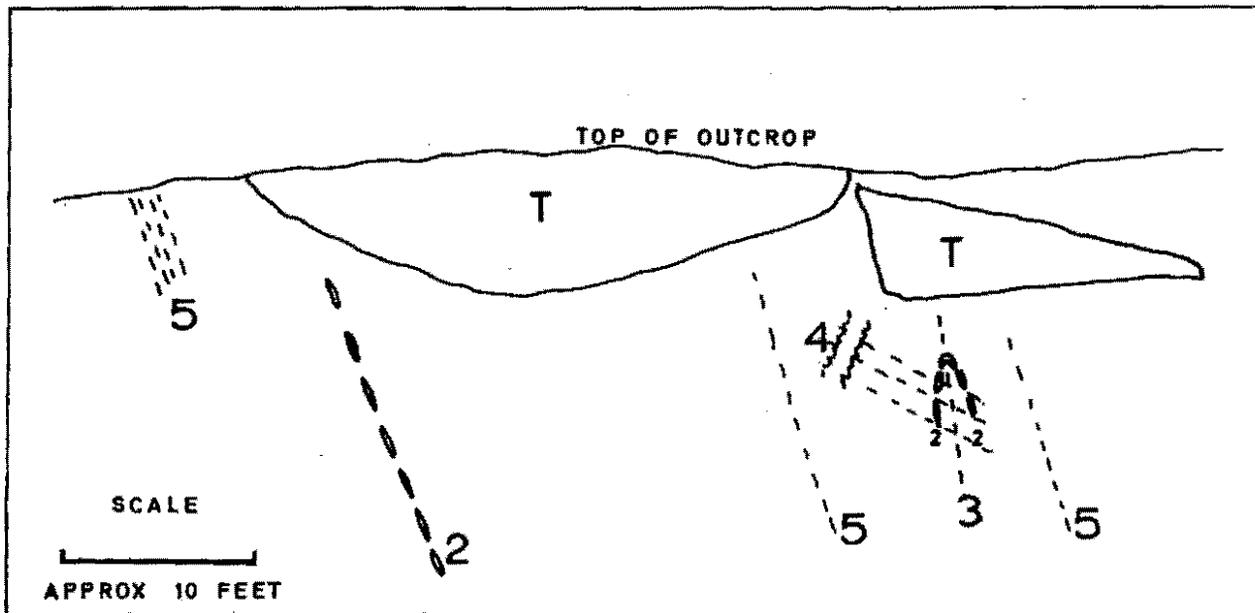
Hinge surfaces and the layering and bedding on the long limbs in the slabby-weathering unit and the Lowerre Quartzite dip moderately to the northwest over much of the outcrop, but considerable folding is present in some places (see Figure 6a).

To the north of the Lowerre, but separated by a two-mica granite (which gives an Rb-Sr isochron age of  $335 \pm 13$  m.y.), is a 10-foot outcrop (at road-level) of diopside-bearing Inwood Marble, followed immediately by Manhattan Schist.

The basal unit of the Manhattan Schist (Walloomsac Fm.) at this point consists essentially of quartz, plagioclase, K-feldspar, garnet, red biotite, and sillimanite, but a little retrograde muscovite is present in places. The high garnet content and the red biotite between them give the schist a distinct reddish hue. A few small layers of marble (up to 4 inches wide) are present near the contact. The dominant folding at this point is of  $F_3$  age, and some quartzo-feldspathic segregations can be seen to be folded while others are approximately parallel to the axial planes of the folds. The boundary with the upper unit is placed at the first amphibolite unit about 50 feet north of the Manhattan-Inwood contact. The basal unit re-appears near the northern end of the outcrop where calc-silicate rocks are present.

Approximately 100 feet north of the contact, and about two thirds of the way up the face of the road cut, is a small area where  $S_1$ ,  $D_2$ ,  $F_3$ ,  $F_4$ ,  $S_5$ , and the post- $D_5$  tourmalinized surfaces are all seen together (see Figure 5). A large tourmalinized surface makes up the top of the road cut, and below it, near its northern end, a zone of  $D_2$  flasers runs down the face of the cut. The zone is 1 to 2 feet wide, and is bounded along part of its south side by a younger quartz-K-feldspar segregation. The flasers consist mostly of schist, but some are quartzo-feldspathic and some consist largely of porphyroblastic plagioclase. A few consist of an intergrowth of coarse-grained quartz and plagioclase that is taken to be an earlier ( $S_1$ ) segregation. Twenty feet to the south of this  $D_2$  flasered zone is another

FIGURE 5. Schematic, oblique view of the face of the roadcut at stop 1 showing the geographic relationships between the  $D_2$  flasered zone, the folded  $D_2$  flasers, and the Tourmalinized joints.



Explanation of Symbols

- 2 =  $D_2$  flasered zone
- 3 =  $F_3$  fold deforming  $D_2$  flasers.  $S_3$  quartz-K-feldspar segregations aligned approximately parallel to the  $F_3$  hinge surface.
- 4 =  $F_4$  chevron folds.  $S_4$  cleavage cuts across the  $F_3$  hinge surface.
- 5 =  $S_5$  muscovite cleavage.
- 6 = fracture surface with tourmaline rosettes up to 18 inches in diameter.

similar one that is isoclinally folded by an  $F_3$  fold. The flasers are clearly visible where they bend around the  $F_3$  hinge surface. K-feldspar-rich segregations cut through the fold in approximately the  $F_3$  axial-planar orientation.  $S_4$  cleavage is faintly visible here cutting across the  $F_3$  hinge surface, and is more clearly visible a few feet to the north where  $F_4$  crinkle folds have formed in the more biotite-rich layers.  $D_2$  sillimanite is bent and broken by  $F_3$  folding;  $F_3$  sillimanite co-exists with K-feldspar; and  $S_4$  sillimanite cuts across K-feldspar of the  $D_3$  segregations. No primary muscovite is present in these rocks, but young muscovite is present a) on zones of  $S_5$  granulation, b) on and adjacent to the tourmalinized surfaces, and c) in some randomly distributed spots where later circulating fluids were able to penetrate.

The  $S_5$  cleavage is oriented approximately parallel to the  $S_3$  (and  $S_1$ ) foliations and as a result is difficult to detect in many places. One clear zone is present at the top of the roadcut about 15 feet north of the northern end of the big tourmalinized surface, and another (containing staurolite) is present 50 feet further to the north. Less conspicuous zones are present near the folded  $D_2$  flasers. Here, individual  $S_5$  surfaces containing granulated quartz and feldspar and fine-grained muscovite, cut through coarser (ungranulated) schists that contain K-feldspar and sillimanite and no muscovite.

The  $D_6$ -related tourmalinized surfaces cut across all the older structures. They are extensive, traceable fractures that maintain their orientation over much of the Croton Falls map sheet. On the large exposed surface at the top of the roadcut (Figure 5) tourmaline has grown into rosettes up to 18 inches in diameter that appear to be completely undeformed. Some tourmalinized surfaces have abundant muscovite on them; others (for example low on the roadcut about 70 feet north of the  $D_2$  flasered zone) have up to an inch of feldspar with minor quartz emplaced along them; and in one case (in the Croton Falls complex) tourmaline and kyanite have crystallized together on one of the surfaces. Two hundred feet to the northwest of here (Figure 5), in the roadcut on Route 22, two-mica granites are intruded along these surfaces.

From this point northwards, the Manhattan Schist is relatively unremarkable until, near the northern end of the roadcut, considerable segregation and granitization has taken place. Here the less-modified host rocks consist largely of quartz-K-feldspar-plagioclase-sillimanite-biotite-garnet schists with minor interlayered calc-silicate rocks and amphibolites. A large, irregularly-shaped segregation cuts across the layering and across the  $F_3$  hinge surfaces. The segregation varies considerably in composition from one place to another. In some places it consists of quartz, K-feldspar, and plagioclase, with minor garnet, biotite, sillimanite and graphite; whereas in others the feldspar content is negligible, and the rock consists almost entirely of quartz, garnet, and sillimanite. Where this rock is cut by  $S_5$  cleavage, kyanite is seen overgrowing two ages of older sillimanite (see illustration in Brock and Mose, 1979). A suite of seven specimens from this segregation gave an Rb-Sr isochron age of 442 m.y., (or 462 my. if one sample with extremely low Sr content is discarded as unreliable) (Brock, Brueckner and Brock, 1984).

		Continue south on Route 22
48.4	0.3	Junction with Route 116. Continue south on Route 22
51.0	2.6	Turn left on Route 138.
51.1	0.1	Turn right into supermarket parking lot. Park near the southern end of the parking lot. Stop 2.

## STOP 2

The outcrop at this stop extends from the southern end of the parking lot, north to the northeastern end of the shopping center. The outcrop consists of fine-grained granofelsic quartzo-feldspathic rocks interlayered with thin schists and scattered thicker layers of fine-grained, lined amphibolites. All these rocks have been repeatedly refolded and then cross-cut by two-mica granites. This suite of rocks is distinctly different to the normal Manhattan Schist, and we interpret it to be part of the Hartland Formation emplaced along the hinge zone of a major  $F_2$  fold.

Most of the schists and granofelsic rocks are devoid of aluminosilicate minerals, and many have only limited amounts of garnet and muscovite. The absence of even relict aluminosilicate minerals leaves us uncertain about the maximum grade of metamorphism reached. Near the northern end of the main exposure, a small layer of quartz-plagioclase-muscovite-biotite-garnet-staurolite schist is present, but since this rock occurs in a zone that appears to have been significantly deformed and hydrated during  $D_6$  deformation, it only indicates the grade associated with the  $D_6$  rather than the maximum grade attained.

The granites range from relatively fine-grained unfoliated dikes with sharp boundaries, slightly chilled edges, and zoned plagioclase grains, to coarser-grained to pegmatitic rocks with or without swirling flow foliation and with or without a later tectonic mineral alignment imposed. These granites give an Rb-Sr isochron age of 345±9 m.y. (Brock *et al.*, 1985).

At the southern end of the rock-cut beside the parking lot, shallow-dipping two-mica pegmatitic granites cut across the host schists. The granites have been slightly deformed, and a rough fracture cleavage is locally developed. Near the center of the cut beside the parking lot, the granite is finer-grained and is locally distinctly foliated. Several small apophyses of this granite and others in this outcrop can be seen to be folded, and in these cases it can be seen that the mineral parallelism in the granite cuts across the contacts and is parallel to the axial planes of the folds. Pre- $D_6$  structures in this outcrop are anomalously oriented and appear to show varying degrees of rotation due to  $D_6$  deformation. Although shear-related fabrics are only sporadically developed, rotation of up to 60° appears to be present in the host rocks. Where a set of folds of the same age could be identified and the variation in attitudes due to heterogeneous  $D_6$  shearing recorded, a moderate northwest-plunging  $D_6$  slip line has been derived (see Figure 6b). The sense of rotation indicates normal movement. The slip line is parallel to the aligned tourmaline commonly found in  $D_6$  zones throughout the area. The granites in this outcrop are little deformed, but similar (Mississippian) granites elsewhere sometimes show well-developed c-s fabrics reflecting normal shear sense.

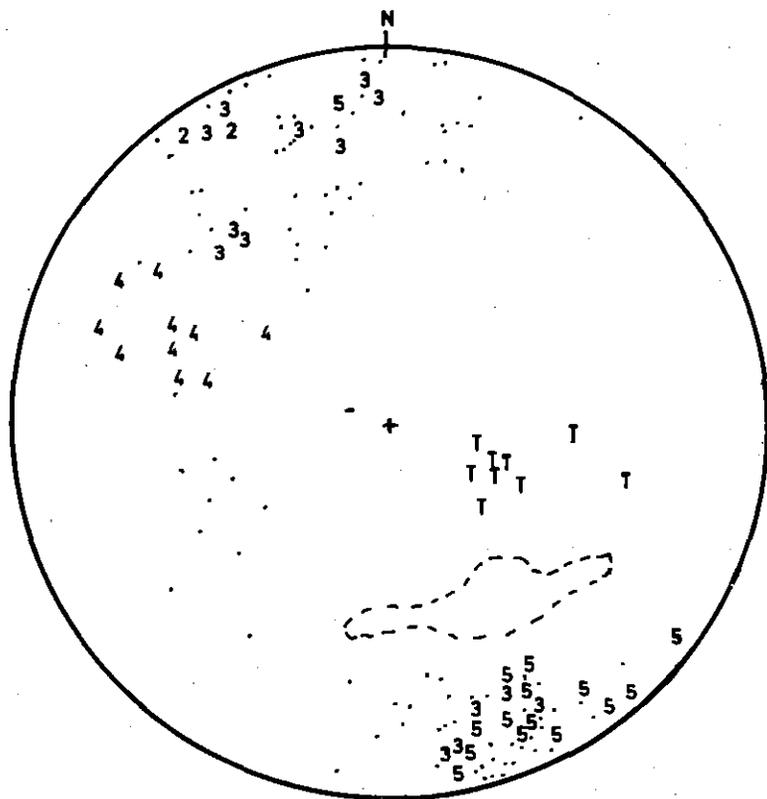


FIGURE 6a. (left) Equal area net showing orientations of S-surfaces in the post-Fordham units at Stop # 1.

Explanation of symbols

i. S-surfaces in the Manhattan Schist

• = Pole to  $S_1$ .      2 = Pole to  $S_2$ .  
 3 = Pole to  $S_3$ .      4 = Pole to  $S_4$ .  
 5 = Pole to  $S_5$ .      T = Tourmalinized Fracture Surface.

ii. Area outlined with dashes encloses poles to  $S_1$  and  $S_3$  surfaces in the Lowerre Quartzite and in the underlying slabby-weathering unit.

iii. The S-surfaces in the Fordham Gneiss overlap with the dominant cluster of the Manhattan  $S_3$  and  $S_5$  cleavages and cannot be plotted without unduly cluttering the diagram.

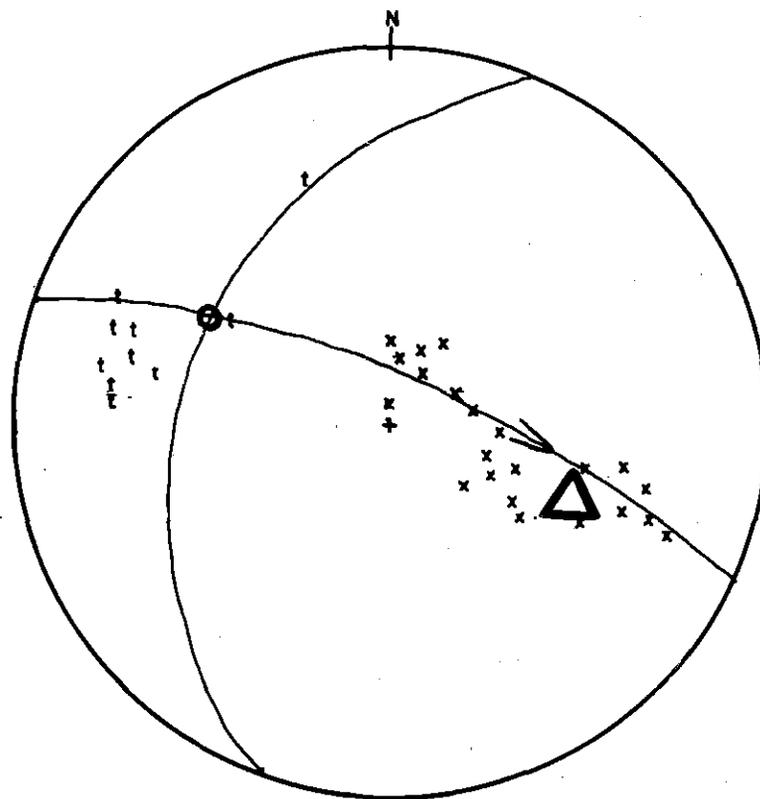


FIGURE 6b. (right) Equal area net showing the rotation of  $F_3$  hinge lines by  $D_6$  in the Goldens Bridge outcrop (Stop 2)

Explanation of Symbols

x = Hinge line to  $F_3$  fold

Triangle =  $S_6$  shear zone boundary derived from average of slip cleavages

○ =  $D_6$  slip line

t = Tourmaline lineations in this and other outcrops

Tourmaline is present within some of the granites, concentrated along the contacts of others, and as rosettes on some appropriately oriented fracture surfaces.

The identification of the fold events present in this outcrop is still uncertain for a number of reasons: the lack of evidence of difference of grades of metamorphism associated with the obvious sequence of fold events; the rotation of all axial surfaces to strike southeast making it difficult to connect with structures in surrounding areas; and the lack of older segregations or granitic markers all leave us without a strong basis for correlating from this outcrop to others.

		Return to Route 22 and turn right (north)
54.9	3.8	Bear right onto Hardscrabble Road
55.1	0.2	Outcrop on the left (opposite the entrance ramp to I-684) is stop 3. Drive on across the overpass over I-684 to find a safe parking space (for instance on the side road opposite the Getty Service station). Stop 3.

### STOP 3

Stop 3 is located on the contact of the Croton Falls mafic complex (see figure 4). The rocks exposed here include modified dunite, pyroxenite, gabbro and diorite of the complex, and down-graded emery-related rocks derived from blocks of Manhattan Schist that were incorporated into the mafic complex at or near its margin. Also present is a belt of foliated, leucocratic plagioclase-biotite-amphibole rock that cuts through and incorporates blocks of the mafic rocks, and stringers of more-normal quartzo-feldspathic vein material. The rocks in this contact zone have been subjected to considerable deformation and recrystallization during  $F_3$ ,  $D_5$ ,  $D_6$ , and possibly  $D_7$  times. The neighboring outcrops along the road in both directions are Manhattan Schist.

In the altered mafic rocks, the olivine has been partly replaced by phlogopite (and more rarely to anthophyllite), and the pyroxene by amphibole, although the augite in some of the more gabbroic rocks is still fresh and unaltered. The diorites are texturally the most modified rocks, having taken on a strong  $S_3$  foliation as they have throughout the complex. They consist of andesine and biotite with only minor amounts of original hornblende surviving, though a late-stage dark blue-green amphibole can be present around grain boundaries of biotite in places.

Unmodified emeries in the xenoliths deep within the Complex can consist entirely of corundum and magnetite, though usually some garnet is also present. In contrast, here at the margin of the Complex aluminosilicates and some plagioclase survived the contact metamorphism, though because of all the later modification, it is not always certain which minerals belong with which metamorphic event. One of the most striking rocks in this outcrop is the light-colored garnet-rich rock making up the steepest high cliff near the center of the roadcut. It contains sillimanite, staurolite, cordierite, almandine, gedrite, oligoclase, and

members of the magnetite-ulvospinel and ilmenite-hematite solid solution series', together with minor amounts of late kyanite, biotite and muscovite. Of these minerals the almandine and cordierite were clearly no longer stable together during the later metamorphism, and they are now always separated from one another by screens of Sillimanite, staurolite, and/or gedrite. In sheared chloritized zones emery relicts characteristically consist of magnetite, corundum, plagioclase, and biotite in which the original minerals are now largely replaced by sericite, chlorite, and epidote. Other common rocks include pelites containing plagioclase, sillimanite,  $\pm$  kyanite, corundum, biotite, garnet, magnesian hercynite, and black Fe-Ti oxides, and amphibolites that now have a grayish-green amphibole of the cummingtonite-grunerite family as the dominant mineral.

The leucocratic foliated plagioclase-biotite-hornblende rock is something of an enigma. In its light color, its foliation, layering and folding, it looks similar the host schists. However, in its mineralogy (andesine, hornblende and biotite) it very closely resembles the Croton Falls diorite except that it is much more leucocratic than any of the normal diorites. In addition, it has blocks of the Croton Falls mafic rocks (tectonically) incorporated into it. It is possible that it represents a xenolith of host schist that was incorporated into the diorite magma and that equilibrated chemically with the diorite (as opposed to the emery xenoliths that equilibrated with the ultramafic magma).

The rocks in this outcrop have been deformed and remetamorphosed a number of times, but the recrystallization has not affected all areas equally: some patches of dunite have survived almost unmodified; the gabbroic rock appears fresh and unaltered; and the cordierite-garnet rock has resisted recrystallization during most of the later, lower-grade events. In other areas, by contrast, the rocks have been severely recrystallized and/or reformed several times. Recrystallization during each event was largely controlled by access to water. Thus a zone of remetamorphism surrounds those shear zones and S-surfaces that penetrated the highly competent rocks of the complex, and partial preservation of mineral assemblages reflecting each metamorphic event is the rule. Based on relationships worked out within the complex, the lowest grade of metamorphism occurred early in the  $D_6$  cycle (Brock and Brock, 1981), although recent work elsewhere suggests a second metamorphic low with  $D_7$ . By this reasoning the chlorite-epidote-sericite alteration of the emery relicts could be assigned to either the  $D_6$  or  $D_7$ .

Within the Complex, emeries modified by  $D_6$ -related metamorphism develop assemblages containing kyanite, chlorite, and staurolite,  $\pm$  muscovite,  $\pm$  tourmaline, and old garnets often show extensive alteration to chlorite. Later prograde metamorphism is suggested by a number of textures, including the growth of biotite screens between chlorite and staurolite, and around young euhedral garnets that contain inclusions of chlorite. Young acicular sillimanite is present in these assemblages. It is uncertain whether this prograde metamorphism pertains to a late stage of  $D_6$  deformation or to  $D_7$ .

At this outcrop, the alteration of olivine to anthophyllite in dunite, and of pyroxene to hornblende in gabbro is attributed to  $D_3$ - or  $D_5$ -related metamorphism. The cordierite-garnet rock is texturally complex, and it is

FIGURE 7. Geologic Map of the area around stop 4, modified from Prucha et al., 1968, Fisher et al., 1970 and Rodgers, 1982

Explanation of Symbols    ⚡ = Stop location    x = Age Determination Location  
 .. .. . = Stream    + = Cordierite-Garnet Locality  
 M = Manhattan Schist    I = Inwood Marble

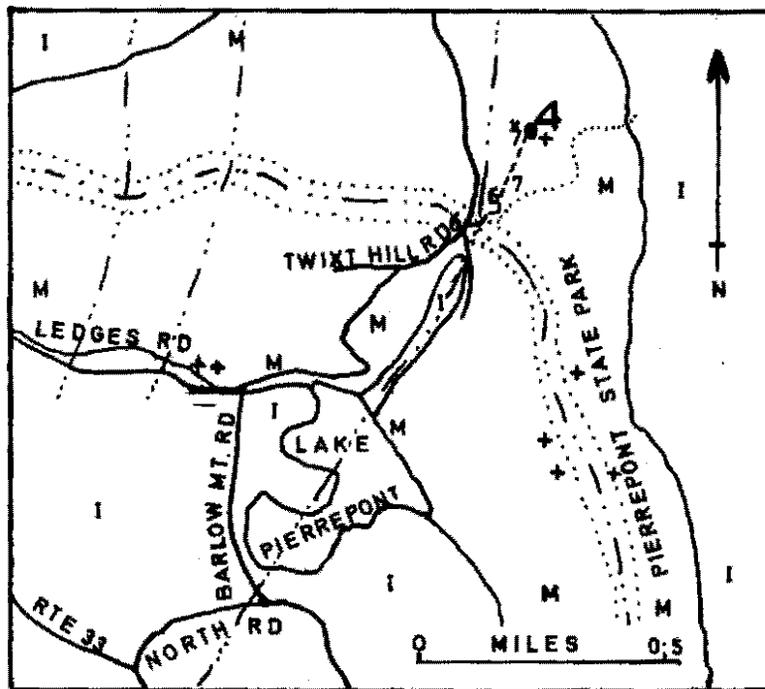
..... Schematic representation of Hartland Formation along the  $F_2$  axial trace that we think is responsible for its presence

— .. — .. —  $F_2$  Axial Trace

— ... — ... —  $F_3$  Axial Trace

6 = Zone of  $D_6$  ductile shearing

7 = Zone of  $D_7$  thrusting



difficult to relate its middle of assemblages to specific deformational events, but within the Complex emeries locally alter to gedrite + staurolite+sillimanite along  $S_3$  cleavages.

Deformation that disturbs the foliation in the diorite is post  $D_3$ . Where such deformation shows high-grade character it is probably of  $D_5$  age. ( $F_4$  has not been recognized in the complex.) Deformation of  $D_3$ ,  $D_5$ ,  $D_6$ , and  $D_7$  age are all thought to be present at some point along the eastern contact of the Croton Falls Complex.

Retrace route back down Hardscrabble Road and onto Route 22 south

55.4	0.2	(Mileage counted from outcrop of stop 3, not where you parked)
56.7	1.3	Turn left (east) onto Route 116
60.3	3.6	Cross Route 124
60.8	0.5	Bear left onto combined Routes 116 & 121
61.5	0.7	Balanced rock on right
61.9	0.4	Turn right onto Route 116
62.1	0.2	Turn left onto dirt road (Mopus Bridge Road)
63.6	1.5	Bear right onto Spring Valley Road
63.8	0.2	Bear right onto George Washington Highway
64.0	0.2	At stop sign turn left onto Ledges Road
64.9	0.9	At stop sign turn left onto Barlow Mountain Road
65.2	0.3	Hairpin bend on Pierrepont Drive
65.5	0.3	Turn right onto Twixt Hill Road
65.6	0.1	Turn right onto deadend road and park, without blocking the road. Stop 4.

STOP 4 is in the Pierrepont State Park. No hammering.

It is located off the trails in fairly thick bush, and is not easy to find. If you get separated from the group, walk west till you reach the road, and then follow the road south until you find the starting point (see Figure 7).

Enter the park through the chained gate across the dirt road. Bear left on the larger track and follow it down until it is about to cross a small north-flowing stream. Just before the stream turn left onto a very old track that goes off to the left. It is barely recognizable at this point, but becomes a little more obvious as one follows it. The track goes approximately north, passing between the main height of land to the west and a smaller ridge to the east. We will visit this ridge if time permits after the main stop. Follow the track for about a quarter of a mile, at which point it will once more become hard to follow. Bear right for a few tens of feet until the ground drops away between some quite large outcrops and you find yourself in a small amphitheater surrounded by outcrops.

The rocks at this stop are Manhattan pelites that were metamorphosed to perthite-cordierite-garnet-sillimanite grade during the  $D_2$  deformation, were somewhat down-graded (to K-feldspar-sillimanite grade) during the  $D_3$ , and then locally slightly modified during the  $D_6$  and  $D_7$  deformational events. The main purpose of this stop is to show you what the Manhattan

looks like when it has been up to such high metamorphic grade.

Most of the pelites consist of coarse-grained perthite and garnet, separated by trains of finer-grained quartz, biotite, sillimanite and magnetite that have been folded by  $F_3$  folds. The pelites at this locality have a higher percentage of potassic feldspar in them than is normal in the Manhattan Schist. A few grains of kyanite are present in a few places, and small amounts of sericite and chlorite are visible in most thinsections. Coarser-grained secondary muscovite is also present in a few places. Cordierite is only present in restricted areas; generally in only small amounts (less than 15%). (Specimens from nearby localities where it constitutes nearly 50% of the rock will be available for your inspection on the bus.) Cordierite, perthite and garnet  $\pm$  sillimanite co-existed stably during  $D_2$  times, but during  $F_3$  times the garnet-cordierite pair was unstable, and screens of biotite and sillimanite are generally present along the grain boundaries. In a few places cordierite is partly replaced by gedrite, and in others it is replaced by green biotite and/or a yellow ?amorphous material. In one locality along Washington Ledges the rock has been flasered, and the flasers of garnet-perthite-cordierite are separated by trains of almost pure sillimanite. In many of the cordierite-bearing rocks both the cordierite and the plagioclase have grown into large sieve crystals that enclose smaller grains of most other minerals. Garnet porphyroblasts routinely enclose biotite, sillimanite, quartz, plagioclase and opaques, but only rarely enclose perthite, and then only in the outermost zone.

The pelites have been intruded by perthite-bearing granites that have been deformed during  $F_3$  deformation. The contacts of these granites are now rather diffuse in many places. Trains of sillimanite cut across the perthite porphyroblasts .

Most of the rocks in this "amphitheater" area show only limited signs of late, lower-grade metamorphism, except in a zone of  $D_7$  shearing along the western edge of the amphitheater. In the shear zone grain-size is reduced and the rock is made up of fine-grained red biotite, quartz, and feldspars, with scattered relict sheared and/or rotated porphyroblasts of garnet and perthite, and with lesser amounts of sericite and chlorite (+a little surviving sillimanite). In places the muscovite and chlorite are concentrated in strain shadows of some of the porphyroblasts. A few small euhedral garnets grew towards the end of the deformation. The granites near the shear zone show more signs of recrystallization: some feldspars are completely sericitized and some garnets are completely chloritized.

Biotite from this  $D_7$  zone has given a K-Ar age of  $294 \pm 13$  m.y., and muscovite from another zone two miles north of here gives an age of  $309 \pm 8$  m.y.

The outcrop on the small mound to the south of the amphitheater (that we passed as we walked in) shows more evidence of lower grade remetamorphism. Here  $D_6$  deformation has produced muscovite-rich shear zones that cut across older structures, including  $F_3$  and possibly  $F_5$  hinge surfaces. The introduction of water necessary to make the muscovite is characteristic of the  $D_6$ , and where the  $D_7$  shearing follows  $D_6$ , it too has more muscovite associated with it.

- Turn around and turn left onto Twixt Hill Road
- 65.7 0.1 Turn left onto Pierrepont Drive
- 66.6 0.9 At Stop sign, continue straight ahead (south, on Barlow Mt. Road)
- 67.0 0.4 Turn right at "T" junction onto North St
- 67.3 0.3 Turn left onto Route 33
- 69.4 2.1 At Stop sign turn right onto combined Routes 33 and 35
- 70.0 0.6 Center of Ridgefield

## END OF TRIP

From here there are several possible routes back to New Haven. The simplest is probably to follow route 33 (Not 35) south to the Merritt Parkway (Route 15), or to the New England Throughway

45.5

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THE HOPE VALLEY SHEAR ZONE -  
A MAJOR LATE PALEOZOIC DUCTILE SHEAR ZONE  
IN SOUTHEASTERN NEW ENGLAND

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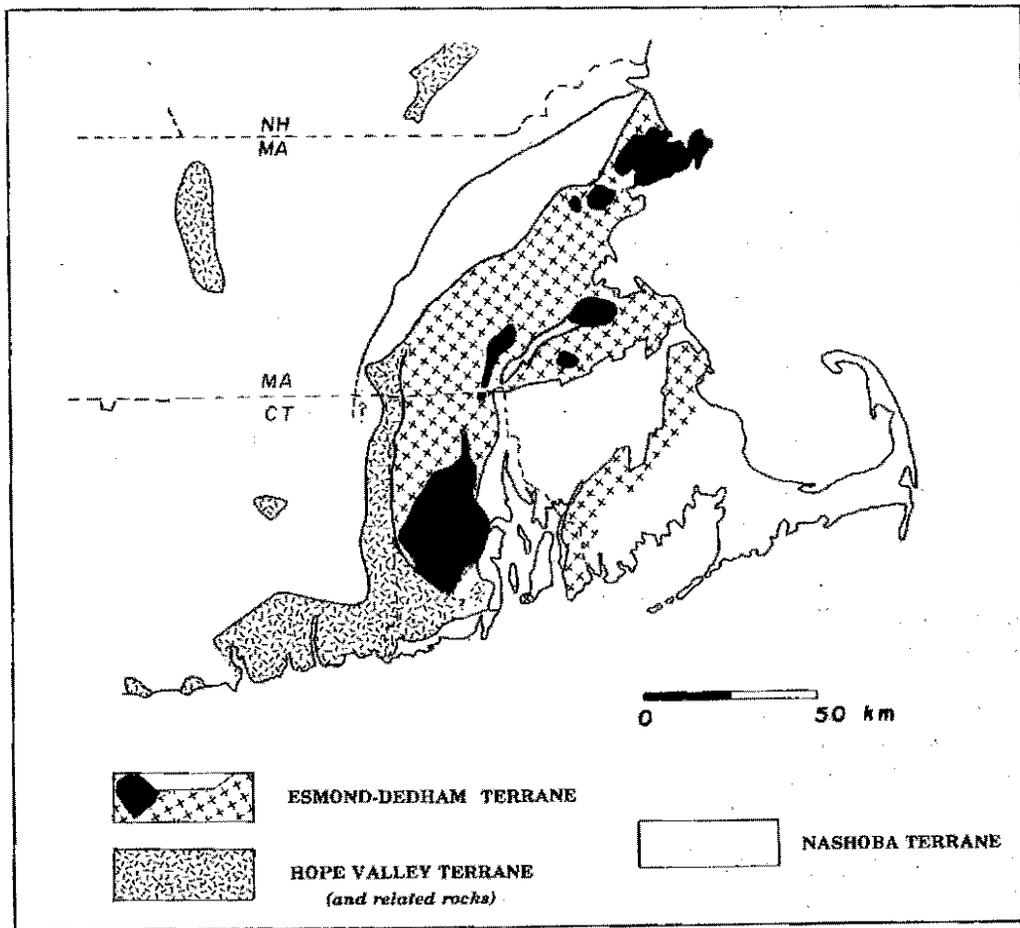
INTRODUCTION

A variety of late Precambrian rocks are exposed to the south and east of the Honey Hill, Lake Char and Bloody Bluff faults in southeastern New England. Granitic plutons and granitic gneisses predominate, with lesser amounts of volcanic and sedimentary rocks of several associations. These rocks generally have been treated as an integral tectonic unit and considered part of the "eastern basement" of the Appalachians. They have been grouped with other similar rocks along the eastern margin of the orogen, forming an extensive lithostratigraphic belt commonly referred to as the Avalon zone (e.g., Rast et al., 1976).

Recent work centered in Rhode Island provides some important new insights into the development of the eastern basement in southeastern New England. This work has determined that this segment of the Avalon zone is composed of two distinct groups of late Precambrian rocks (Hermes and Gromet, 1988; Gromet and O'Hara, 1984; Hermes and Zartman, 1985) separated by a major but previously unrecognized Alleghanian shear zone (O'Hara and Gromet, 1984, 1985). The differing Paleozoic histories of the two groups of rocks, as revealed by contrasts in the timing and nature of Paleozoic magmatic and deformational events, has led to their treatment as terranes: an eastern Esmond-Dedham terrane and a western Hope Valley terrane (Fig. 1).

The boundary between the Esmond-Dedham and Hope Valley terranes, which is partly cryptic, has been located and characterized by a combination of field, petrographic and geochronological methods. The boundary is zone of highly strained, ductilely deformed gneisses derived from rocks of both terranes. Structural elements of these rocks show a close spatial and geometrical relationship to the boundary and suggest they formed within a major right-lateral shear zone.

The purpose of this trip is to examine some of the gneissic rocks that constitute the boundary between the two terranes, termed the Hope Valley shear zone. The Hope Valley shear zone is named for its proximity to the town of Hope Valley in southwestern Rhode Island. The Hope Valley shear zone extends from south-central Rhode Island northward along the Rhode Island-Connecticut border into adjacent Massachusetts (Fig. 1). The trip begins along the southern segment of the shear zone in southwestern Rhode Island, where Hope Valley alaskite and associated biotitic gneisses (Stops 1 and 2) are found against Devonian Scituate granite gneisses (Stops 3-5). At Stop 5, the Precambrian Ponaganset



**Figure 1.** Terrane map of southeastern New England showing the distribution of the Esmond-Dedham, Hope Valley and Nashoba terranes. The Esmond-Dedham terrane is intruded by anorogenic Paleozoic plutons (dark shading), overlain by the Narragansett and Norfolk basin sediments (unshaded), and bounded to the west by the Bloody Bluff fault and the Hope Valley Shear Zone. The precise location of where the HVSZ joins the Bloody Bluff has not been determined. The Nashoba terrane may continue further south along strike. Late Precambrian granite gneisses in the Massabesic anticlinorium, NH (top center of figure) may correlate with the Hope Valley rocks and underlie much of the Merrimack synclinorium. After O'Hara and Gromet (1985).

gneiss also is observed. Later stops will be along the northern segment of the shear zone where the Ponaganset gneiss (Stops 6,8 and 9) of the Esmond-Dedham terrane is found against Hope Valley alaskite (Stops 7 and 10). Trip stops are shown on Figure 2.

### THE ESMOND-DEDHAM AND HOPE VALLEY TERRANES

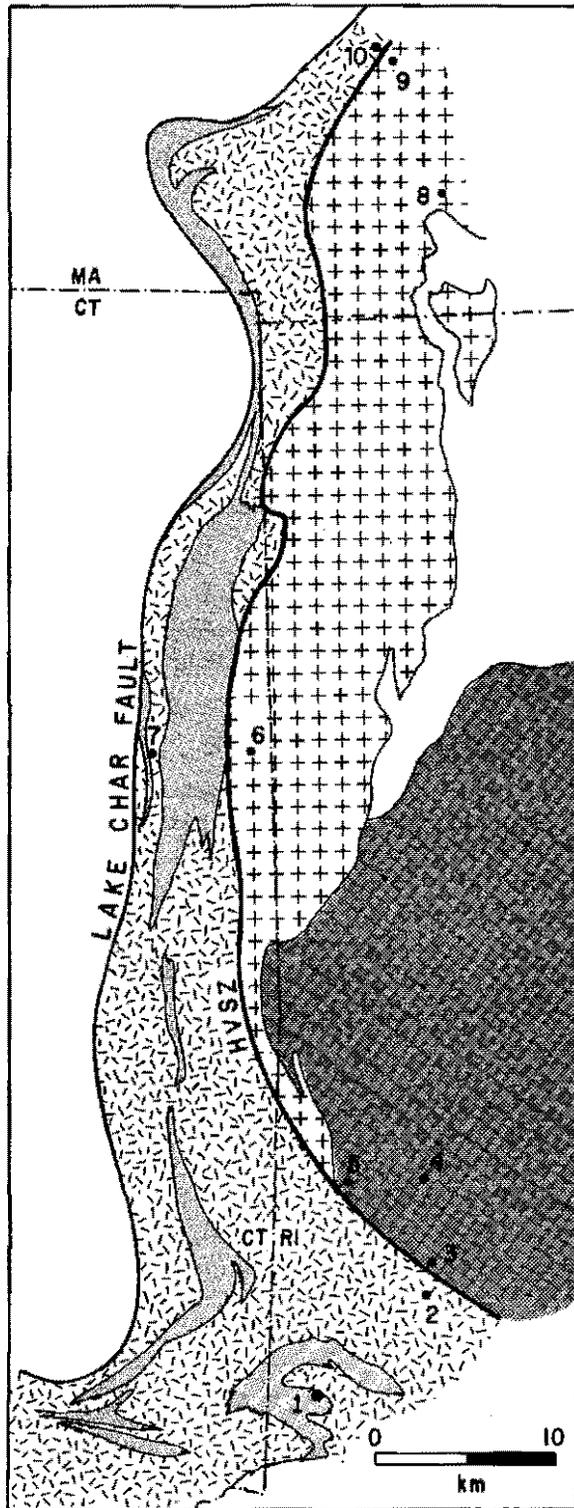
The Esmond-Dedham and Hope Valley terranes were recognized on the basis of observations made on a regional scale. The rocks and relationships seen on this excursion can provide only a partial appreciation of the distinctive nature of the two terranes. A brief summary of some of the more important characteristics of the two terranes is provided below. A more complete account is given in O'Hara and Gromet (1985).

The Esmond-Dedham terrane underlies most of Rhode Island and that part of eastern Massachusetts lying east of the Bloody Bluff fault. Prominent late Precambrian intrusive members include the Esmond group, the Dedham and Milford granitic rocks, and the Ponaganset gneiss. These rocks are a lithologically diverse group of ca 620 Ma granitic rocks ranging in composition from diorite to granite. They intrude older rocks such as those of the Blackstone series and are locally overlain by latest Precambrian to early Paleozoic marine sediments containing trilobites of Acado-Baltic affinities (Skehan et al., 1978). Collectively, these rocks are intruded by numerous alaskite (locally strongly peralkaline) and subalkaline granites of Orodovician to Devonian age. The 370 Ma Scituate granite in central Rhode Island is now known to be one of the largest members of this group (Hermes et al., 1981; Hermes and Zartman, 1985). Non-marine Pennsylvanian sediments fill the Narragansett and Norfolk Basins, which developed as pull-apart basins (Mosher, 1983) within the Esmond-Dedham terrane.

The late Precambrian rocks of the Hope Valley terrane differ considerably from those of the Esmond-Dedham terrane. The Hope Valley alaskite, Potter Hill granite and Ten Rod granite gneisses are the principal units and they are composed almost exclusively of highly leucocratic granite gneisses. Mafic to intermediate compositions are rare, indicating these rocks are not simply metamorphosed equivalents of the Esmond-Dedham rocks. The granite gneisses intrude small bodies of mafic to intermediate schist and gneiss usually assigned to the Plainfield formation or Blackstone series (undifferentiated). Quartzite and calc-silicate rocks assigned to the Plainfield are also present. Cross-cutting and relatively undeformed pegmatitic to aplitic dikes related to the Permian Narragansett Pier and Westerly granites occur extensively.

Large variations in deformational features and metamorphic grade occur across the region. In central and northern Rhode Island and adjacent Massachusetts, the rocks of the Esmond-Dedham terrane are weakly to moderately deformed and display low-grade metamorphic assemblages. Primary igneous textures are partially to well preserved in the plutonic units outside of local zones of shearing. To the south and west toward the boundary with the Hope Valley rocks, Esmond-Dedham rocks of all ages become increasingly more deformed and recrystallized. Rocks near the boundary such as the Ponaganset and Scituate are ductilely deformed gneisses possessing a strong penetrative lineation and/or foliation. The Ponaganset gneiss in particular becomes an intensely lineated augen gneiss containing 5-20% feldspar porphyroclasts in a highly recrystallized matrix. A probable tectonic origin for the Ponaganset-Hope Valley contact was first recognized by O'Hara (1983). This contact constitutes the northern segment of the Hope Valley shear zone.

The late Precambrian Hope Valley rocks are everywhere highly deformed and have a strongly developed penetrative fabric. Extensive recrystallization has left little if any of



**ESMOND - DEDHAM TERRANE**

-  PONAGANSET GNEISS
-  SCITUATE GRANITE

**HOPE VALLEY TERRANE**

-  HOPE VALLEY ALASKITE  
and related rocks
-  PLAINFIELD FORMATION

-  Stop locality
-  Geological contact
-  Major structural boundary

*Figure 2.* Locations of trip stops along the Hope Valley shear zone in western Rhode Island, eastern Connecticut and adjacent Massachusetts. Distribution of the larger masses of the Plainfield Formation within the Hope Valley terrane is shown.

the original igneous mineralogy. Many of these rocks closely resemble deformed Devonian Scituate granite. In our experience, these rocks cannot be confidently distinguished from one another on field or petrographic criteria. This led earlier workers (who had no way of knowing that the Scituate is Devonian in age) to consider the Hope Valley alaskite and the Scituate to be gradational. Consequently, small bodies of Scituate were mapped within the Hope Valley rocks, and small bodies of Hope Valley alaskite and Ten Rod granite were mapped within the Scituate.

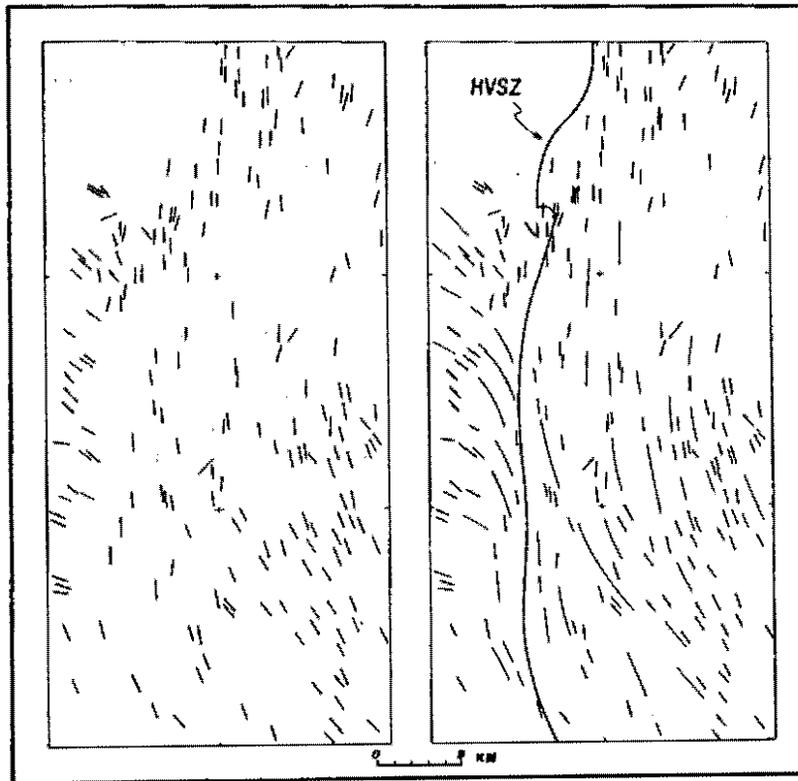
Recent geochronological and field studies (Gromet and O'Hara, 1984; O'Hara and Gromet, 1985) has clarified the extent of the Scituate granite, showing that it extends considerably further south in Rhode Island than previously mapped, and that the small bodies of "Scituate" mapped within the Hope Valley rocks of southwestern Rhode Island and eastern Connecticut are late Precambrian in age. The newly determined distribution is shown in Figs 1 and 2. It is now clear that the southwestern margin of the Devonian Scituate granite is a highly deformed granite gneiss, and that Ordovician to Devonian anorogenic plutons do not extend into the Hope Valley terrane. The SE-trending contact between Scituate granite and Hope Valley alaskite is tectonic contact and constitutes the southern segment of the Hope Valley shear zone.

### THE HOPE VALLEY SHEAR ZONE (HVSZ)

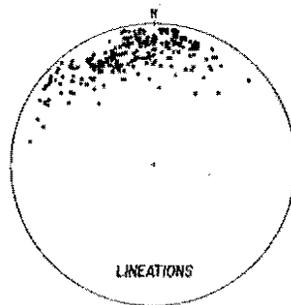
The northern and southern segments of the Hope Valley shear zone have been identified in somewhat different ways. The northern segment is a contact between the Hope Valley and Ponaganset gneisses located principally by field relations and deformational features (O'Hara, 1983). The southern segment is a lithologically cryptic contact between Scituate granite and Hope Valley gneisses located largely on the basis of geochronological methods (Gromet and O'Hara, 1984). However, the two segments together define a smooth and continuous boundary (Figs 1 and 2) that share a coherent set of structural characteristics.

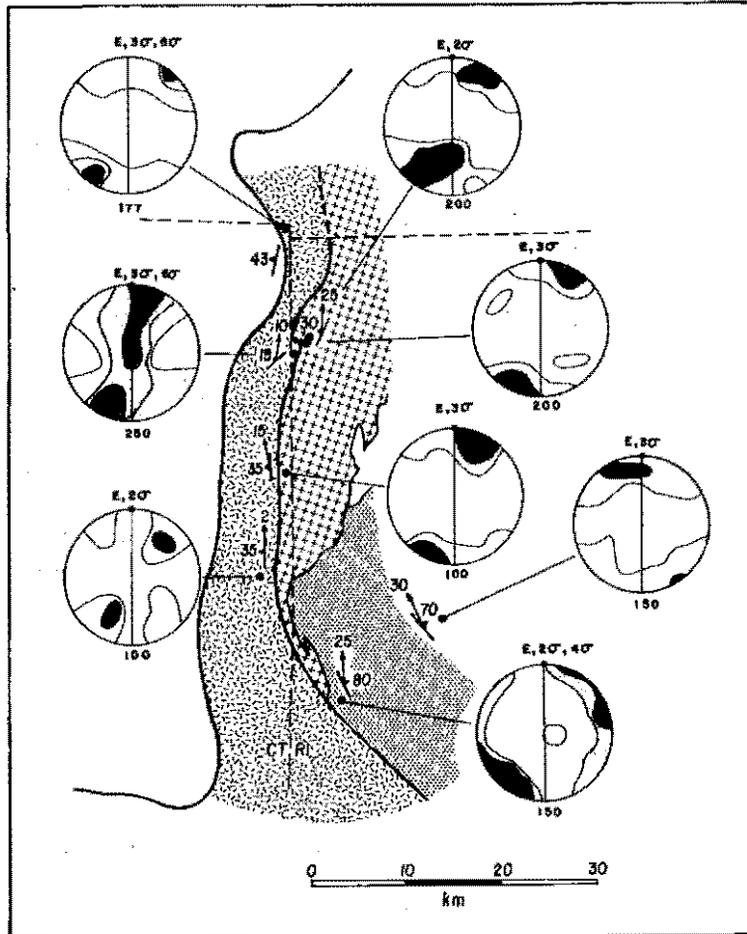
Lineations and foliations associated with the Ponaganset-Hope Valley contact are seen to pass smoothly and continuously along strike into the Scituate and Hope Valley gneisses along the southern segment of the HVSZ. The attitudes and relative intensities of the structural features vary in a systematic manner along the HVSZ. Along most of the Ponaganset-Hope Valley segment, a strongly developed lineation (approximately N-trend, 0-30°N plunge) is accompanied by a weak to moderate west-dipping foliation (10-40°) (e.g., Stops 6, 7 and 8). As the HVSZ is followed south, the foliation steepens and intensifies (Stop 5) into a transition region where the boundary deflects to the southeast. Continuing along the SE-striking southern segment, the foliation in the Scituate and Hope Valley gneisses overturns and dips northeast (Stops 2 and 3) and the gneisses are more highly foliated than lineated.

These variations in structural features provide some important insights into the formation of the HVSZ. The dominant structural feature of the Ponaganset-Hope Valley segment is the strongly developed north-trending lineation. Three lines of evidence occurring on a range of scales indicate that the lineation developed as a stretching lineation within a north-trending right-lateral shear zone. On a regional scale, the lineations display an overall NW-trending pattern away from the HVSZ, but are deflected into parallelism with the HVSZ on approach from either side (Fig. 3). Similar patterns are observed in ductile shear zones on a variety of scales (Ramsay and Graham, 1970), where the elongation direction (represented here by the lineation) is deflected toward the shear plane (the Ponaganset-Hope Valley contact) at high shear strains. The sense of deflection in Figure 3 indicates a right-lateral shear sense on the shear plane.



**Figure 3.** Left: lineation trends in gneisses from eastern Connecticut and western Rhode Island based on six quadrangle reports (Dixon, 1974; Frost, 1950; Harwood and Goldsmith, 1971; Moore, 1963, 1983; Quinn, 1967). The plunge of the lineations is shallow to the north (See below). Right: same as left with the Ponaganset-Hope Valley segment of the HVSZ indicated. Flow lines have been added to emphasize the regional lineation pattern. Note the deflection of the lineation towards the boundary from either side. Below: equal area projections of lineations from the area, showing shallow northerly plunge. From O'Hara and Gromet (1985).





**Figure 4.** Equal-area projections (lower hemisphere) of quartz c-axis preferred orientations for samples of Ponaganset gneiss, Hope Valley alaskite and Devonian Scituate granite close to the boundary between the Esmond-Dedham and Hope Valley terranes. The patterns are plotted so that the foliation (line) is oriented vertically and the lineation (dot) occurs at the top of the pattern. Their geographical orientation can be obtained by reference to structure symbols at the sample localities. Note the similar asymmetry of the patterns along the length of the boundary. One sample, away from the boundary, indicates the opposite sense of shear. Contour intervals are indicated at the top of each pattern where E is the expected number of points within the counting circle for a randomly distributed population and sigma is the first standard deviation from E. The number of grains measured is indicated for each pattern. From O'Hara and Gromet (1985).

On the outcrop scale, feldspar augen with asymmetric deformation tails are commonly developed in the Ponaganset gneiss (e.g., Stop 6). Asymmetric tails are best developed parallel to rather than normal to the lineation. The tails are reasonably inferred to have formed during shear parallel to the lineation. The asymmetric tails give predominantly right-lateral shear senses. On a microscopic scale, asymmetric quartz c-axis fabrics (Fig. 4) also are indicative of non-coaxial shear. The fabrics can be explained by a process involving dislocation creep on dominantly prism [c] slip systems in quartz, with the asymmetry about the lineation (extension) direction indicating a right-lateral shear sense (O'Hara and Gromet, 1985).

A major component of right-lateral shear along the Ponaganset-Hope Valley contact implies compression along the SE-trending southern segment of the HVSZ. Several observations on a variety of scales support this inference, including attitude changes along the boundary, foliation development, and orientation of lineations relative to the boundary. The attitude of the HVSZ can be inferred from the foliations in the gneisses, which change considerably along the transition from the northern to southern segments. The HVSZ is inferred to dip west along the northern segment, steepen through the transition region, then overturn and dip northeast along the SE-striking southern segment. A block diagram (Fig. 5) illustrates these changes. It is notable that the gneisses associated with the southern segment are more highly foliated than lineated, indicating a stronger component of flattening, and that the north to northwest trend of the lineations is at a higher angle to the southeast strike of this part of the boundary. Collectively, these features argue that the Scituate granite gneiss has overthrust the Hope Valley alaskite gneiss along the southern segment of the HVSZ.

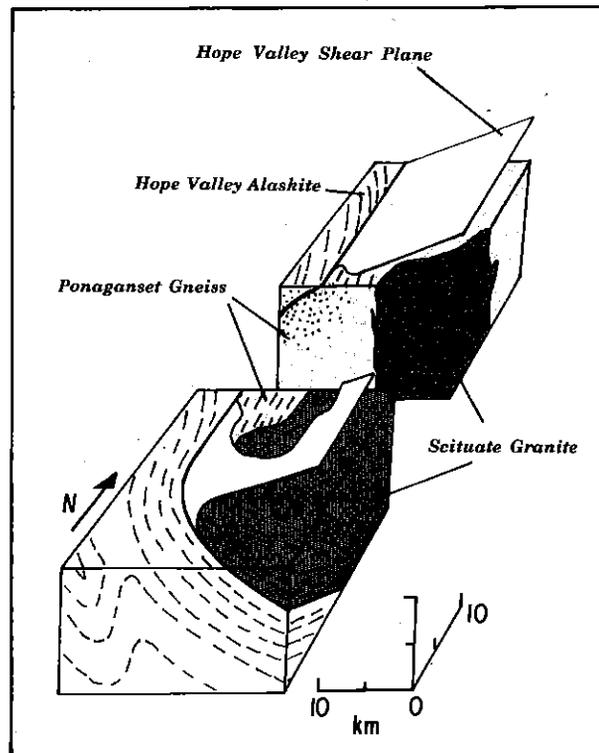
To summarize, the different structural features of rocks along the HVSZ indicate that it originated as a high-grade ductile shear zone with major components of dextral shear along the Rhode Island-Connecticut border and south-directed overthrusting in southern Rhode Island. Deformation associated with the HVSZ produced a thick rind of gneissic rocks along the western and southern margins of the Esmond-Dedham terrane, leaving a less deformed, lower grade interior. Differences in the deformational character of plutonic rocks in southern New England were recognized by Goldsmith (1978), who identified an eastern brittlely deformed terrane and a western ductilely deformed terrane in this region. Goldsmith's deformation terranes correspond to the core of the Esmond-Dedham terrane (brittle), and the outer margins of the Esmond-Dedham terrane and the Hope Valley terrane (ductile).

The age of the Hope Valley shear zone appears to be Alleghanian. An upper age limit is given by the 370 Ma Scituate granite, which has been truncated and deformed by the HVSZ. A lower limit is provided by the Permian Narragansett Pier granite and its associated pegmatites (Stops 1, 2 and 3), which intrude rocks of both terranes and are largely undeformed. An Alleghanian age for the HVSZ is implied by an overall parallelsim of metamorphic isograds in the Pennsylvanian sediments of the Narragansett basin (e.g., Murray and Skehan, 1979; Mosher, 1983) to the southern segment of the HVSZ. The highest metamorphic grades are found in the extreme southwest corner of the basin, where it most closely approaches the HVSZ.

The relationship of the HVSZ to other major fault zones in southeastern New England is the subject of further study. The Bloody Bluff fault and the HVSZ are similar in that together they constitute a western tectonic boundary to Esmond-Dedham rocks. The HVSZ continues northward toward the Bloody Bluff, although the details of how they join remain to be worked out. It seems likely, however, that the HVSZ and the Bloody Bluff are a single structure, with the along-strike variation in deformational characteristics (from brittle along the northern Bloody Bluff to increasingly more ductile southward on the

HVSZ) being indicative of progressively deeper structural levels exposed to the south. This feature appears to be the result of overthrusting and thickening along the southern segment of the HVSZ as it changed from predominantly strike-slip to south-directed thrust motion.

The Lake Char and Honey Hill fault zones are fundamentally different from the HVSZ and Bloody Bluff and probably have no direct relationship to them. The Lake Char and Honey Hill are generally low angle structures separating quartzo-feldspathic basement gneisses from metamorphosed volcanic and sedimentary rocks. Rocks on both the upper and lower plates share an extended strain history that appears to include Alleghanian, Acadian and possibly older deformations (Dixon and Lundgren, 1968; O'Hara and Gromet, 1983). In the Permian, the Lake Char-Honey Hill fault zone appears to have been a basement-cover boundary that was warped and truncated by the HVSZ-Bloody Bluff fault.



**Figure 5.** Schematic block diagram illustrating the inferred geometry of the HVSZ (white plane) in west-central Rhode Island and adjacent Connecticut. The boundary is inferred to dip west along the north-trending segment. Further south it steepens, passes through vertical then dips to the northeast along the southeast-trending segment. From O'Hara and Gromet (1985).

## REGIONAL SIGNIFICANCE

The HVSZ represents both a major regional geologic break and a locus of deformation within the plutonic "basement" rocks of southeastern New England. The distinctive character of the Esmond-Dedham terrane -- its diverse assemblage of late Precambrian plutonic rock types, latest Precambrian to early Paleozoic cover sequence with Acado-Baltic fauna, Ordovician to Devonian anorogenic magmatism, and Pennsylvanian non-marine basins -- indicates an evolutionary history quite distinct from that of the Hope Valley terrane. The absence of Ordovician to Devonian anorogenic granites in the Hope Valley terranes implies the Esmond-Dedham rocks were remote from Hope Valley rocks through this time period. On this basis, it appears unlikely that these two groups of rocks acquired their present relative positions prior to the Late Paleozoic.

It is significant that high-grade metamorphism within the Esmond-Dedham terrane is restricted to its southern and western margins and is Alleghanian in age. The Alleghanian orogeny appears to be the only major Paleozoic deformational event to have affected these rocks. Evidence for the involvement of Esmond-Dedham rocks in the major Taconic and Acadian orogenies in the New England Appalachians is lacking. In contrast, rocks of the Hope Valley terrane occupy the cores of domes located further west (e.g., the Willimantic and possibly Pelham domes in Connecticut and Massachusetts) and, along with rocks of the Merrimack synclinorium, are involved in Acadian structures.

The above observations on the southeastern New England segment of the "Avalon zone" or eastern basement rocks of the Appalachians strongly point to the existence of two different terranes (in the tectonostratigraphic sense) that were accreted to the North American continent in two discrete episodes. The first involved the Hope Valley terrane in or prior to the Devonian. Initial accretion during the Taconic orogeny is possible, with subsequent Acadian consolidation to the ancient North American continental margin. In either case, it appears that the Hope Valley terrane constituted the eastern basement block of the Acadian orogeny, and perhaps the motive force for Acadian convergence. The second accretionary event involved the Esmond-Dedham terrane, which collided obliquely with the Hope Valley terrane along the Hope Valley shear zone in Alleghanian time. This event appears to be one of the principal causes of the Alleghanian orogeny in southeastern New England. Kinematic features of the Hope Valley shear zone and the absence of oceanic rocks along it suggest it was principally a transcurrent fault transporting rocks along the axis of the Appalachians.

Recognition of the bipartite nature of the "Avalon zone" in southeastern New England allows for both the involvement of Avalonian rocks in early and mid-Paleozoic orogenies as well as the late Paleozoic accretion of an "exotic" Avalonian terrane. This appears to resolve some of the fundamental discrepancies among tectonic models of the New England Appalachians (Osberg, 1978; Robinson and Hall, 1980; Rodgers, 1981; Hall and Robinson, 1982; Zartman and Naylor, 1984).

## ACKNOWLEDGEMENT

Partial support for this study was provided by NSF grant EAR-8407664

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## ROAD LOG

## Mileage

Cum. Int.

0.0 0.0

0.3 0.3

2.2 1.9

Begin at park and ride lot at Exit 1, I-95 Hopkinton, RI (8:30 AM 10/5/85).

Turn left onto Route 3 South

Turn left onto entrance ramp, I-95

**STOP 1 BIOTITE GNEISS AND PEGMATITE DIKES**

The grey porphyroclastic biotite gneiss exposed here is fairly typical of biotitic gneisses mapped by previous workers as part of the Plainfield Formation. They occur as scattered masses in this part of the Hope Valley terrane. In nearby exposures, the gneiss is intruded by Hope Valley alaskite. Here, the gneiss is intruded by dikes of the Narragansett Pier granite (a massive, equigranular, medium-grained granite) and related pegmatites.

The biotite gneiss has a fairly strong linear fabric (ESE trend, shallow E plunge) and a weaker foliation (ESE strike, steep SW dip). Narrow, small-scale shear zones and incipient shears occur approximately parallel to and normal to the lineation. The latter shears warp or fold the lineation, a feature not commonly observed in lineated gneisses to the north and east.

Larger granitic to pegmatitic dikes were emplaced along variously oriented surfaces, and some develop a somewhat anastomosing form parallel to foliation planes in the gneiss. Thinner, generally sharp-walled dikes occur along shear planes in the gneiss. Incipient shears commonly have thin (<2 cm) and discontinuous pegmatoid veins along the shear plane, and some shears displace or warp earlier pegmatite veins. It appears that the shears were actively forming during pegmatite emplacement and that some of the pegmatites took advantage of the shears during injection. The late but not always post-tectonic emplacement of dikes related to the Narragansett Pier granite seen here is consistent with observations made further to the east in the Narragansett Basin (e.g., Murray and Skehan, 1979; Mosher, 1983).

7.6 5.4

**STOP 2 HOPE VALLEY ALASKITE AND BIOTITE GNEISS, CROSS-CUT BY PEGMATITE DIKES**

This road cut was previously described by Hermes et al. (1981 a,b). The Hope Valley alaskite is a coarse to medium-grained pink leucocratic gneiss containing no more than a few percent biotite, magnetite and a few other accessory phases. Flattened and elongated aggregates of quartz and feldspar define a strong foliation ( $135^{\circ}/30^{\circ}$ NE) and a weak to moderate lineation (approximately  $165^{\circ}$  trend and  $10-15^{\circ}$ NW plunge). The alaskite intrudes a foliated grey biotite gneiss locally containing conspicuous magnetite octahedra (intrusive relationships are more convincingly displayed in other nearby exposures). The biotite gneiss here was originally mapped as part of the Blackstone series (Moore, 1958). However, it is quite similar to the biotite gneiss observed at Stop 1 and perhaps both are better grouped with the Plainfield Formation. Undeformed pegmatites cross-cut the Hope Valley alaskite and the biotite gneiss.

In thin section, the alaskite and biotite gneiss display a highly equilibrated equigranular texture typical of rocks of the Hope Valley terrane. Quartz, microcline and lesser plagioclase share straight grain boundaries that meet at  $120^{\circ}$  junctions. There is little intragranular strain in these rocks despite the large strains associated with mesoscopic structures, indicating extensive recovery during or after ductile deformation.

The attitude and trend of foliation and lineation fabrics indicate that they are related to right lateral shear and south-directed overthrusting on the Hope Valley shear zone, which passes immediately to the north. The next outcrops seen on I-95 extending up the grade to the northeast are strongly deformed Devonian Scituate granite gneiss of the Esmond-Dedham terrane.

- 8.5 0.9 Take exit for Route 3.  
Road cuts in highly deformed Devonian Scituate granite, previously mapped as Hope Valley alaskite
- 8.9 0.4 Turn left onto Route 3 south

9.5 0.6 **STOP 3 DEFORMED SCITUATE GRANITE GNEISSES AND ASSOCIATED AMPHIBOLITE GNEISS**

The outcrop on the east side of the road contains a foliated, medium to coarse-grained augen gneiss (microcline, quartz, plagioclase, biotite, sphene and iron oxide) interlayered with dark amphibolite gneiss. Both the augen gneiss and the amphibolite gneiss were mapped as part of the Blackstone Series (Moore, 1958). However, the augen gneiss closely resembles porphyroclastic Scituate granite gneiss (see next stop) and the amphibolite gneiss has a much different composition and texture than the "Blackstone" of the previous stop.

The outcrop on the west side of the road is a leucocratic, medium-grained equigranular gneiss originally mapped as Hope Valley alaskite. It is indeed petrographically indistinguishable from common Hope Valley rocks, but two samples of identical rocks from adjacent road cuts on I-95 (immediately to the northeast) have given Devonian ages (O'Hara and Gromet, 1985). The observations here and in nearby areas underscore the difficulty in distinguishing Devonian Scituate rocks and late Precambrian Hope Valley alaskite. Mediumgrained, equigranular leucocratic gneisses are common to both.

The augen gneiss and equigranular gneiss are interpreted to be two different textural types of the Devonian Scituate granite that have suffered considerable post-emplacement deformation. Feldspar porphyroclasts in the augen gneiss have well developed deformation tails. Biotite streaks and feldspar deformation tails define a strong foliation ( $125^{\circ}/25^{\circ}\text{NE}$ ) and a weaker lineation ( $140^{\circ}/10^{\circ}\text{N}$ ). Near the amphibolite gneiss, the augen gneiss becomes finer grained and laminated. Layers of the amphibolite gneiss have been somewhat disrupted by shearing. Pegmatite dikes cut the older rocks, and aplitic dikes (some with discontinuous pegmatoid cores) cut the pegmatites and the older rocks.

- 10.1 0.6 Turn right onto K G ranch road

13.9 3.8

Junction Route 165. Park before intersection, walk west on Route 165 to roadcut on north side of road.

#### STOP 4 DEVONIAN SCITUATE GRANITE GNEISS

This road cut contains a strongly lineated coarse to medium-grained pink augen gneiss and finer grained leucocratic equigranular gneiss. Local contacts between these types range from sharp to gradational. Some of the finer grained gneiss with sharp-walled contacts have the appearance of aplitic dikes that intruded the augen gneiss and were deformed along with it.

A striking feature of these rocks is a strongly developed lineation ( $010^{\circ}/20-30^{\circ}N$ ) defined by streaks of biotite and rods of quartz and feldspar. There is only a weak foliation. The lineation is approximately normal to the face of the road cut and is best observed on undersides and side faces. Deformation is poorly expressed on surfaces normal to the lineation.

In thin section, the augen gneiss consists of quartz, microcline, plagioclase and biotite, with minor hornblende, epidote and sphene. Relict igneous textures are not observed. Feldspar augen are comprised of anhedral, generally equant recrystallized grains which show little intracrystalline strain. Coarse quartz grains make up linear aggregates and are highly undulose with good subgrain development. Finer quartz ( $<0.1$  mm) in the matrix are typically clear unstrained grains. Biotite (with associated granular sphene) and lesser hornblende are aligned parallel to the linear quartz aggregates. The overall mineral textures indicate deformation was ductile and occurred under amphibolite or higher grade conditions.

The rocks exposed here were originally mapped as Scituate granite gneiss (Moore, 1958). Recent Rb-Sr whole rock (O'Hara and Gromet, 1985) and U-Pb zircon (Hermes and Zartman, 1985) dates on the augen gneiss of this outcrop give Devonian ages (370 Ma), indicating emplacement along with the much less deformed Scituate rocks to the north and east.

13.9 0.0

Turn left onto Route 165 west.

16.8 2.9

#### STOP 5 DEVONIAN SCITUATE GRANITE GNEISS AND LATE PRECAMBRIAN PONAGANSET GNEISS

The Scituate granite gneiss exposed here is variable in texture. Two principal types are present: a medium-grained pink leucocratic gneiss and a coarser pink biotite gneiss. Moore (1958) mapped the leucocratic gneiss as Hope Valley alaskite and the coarse biotite gneiss as Scituate granite gneiss, and both assignments are reasonable on a petrographic basis. However, Rb-Sr analysis of both lithologies lie on a Devonian isochron (370 Ma), indicating both are varieties of Scituate granite.

Also present here is a grey porphyroclastic biotite gneiss interspersed with amphibolite layers. The grey porphyroclastic gneiss is similar to the Ponaganset gneiss known extensively in the region to the north, and we group this gneiss with the Ponaganset. The Ponaganset gneiss here

appears to be intruded by the medium-grained Scituate gneiss, although the highly deformed nature of the rocks makes confident identification of contact relations difficult.

All the rocks here are strongly lineated ( $355^{\circ}/10-15^{\circ}\text{N}$ ) and moderately foliated with a steep dip ( $350^{\circ}/70^{\circ}\text{E}$ ). In thin section, the lineation in the coarse Scituate gneiss is defined by alternating elongate aggregates of recrystallized quartz and biotite. The biotite aggregates are thin and discontinuous, and have associated granular sphene and magnetite. Feldspars vary from equant, anhedral grains to somewhat elongate or tabular; microcline is somewhat undulose. The finer grained leucocratic gneiss has a highly equilibrated texture with equant quartz and feldspar grains with smooth, anhedral grain boundaries. Quartz c-axes show a strong preferred orientation pattern that is characteristic of gneisses in western Rhode Island and eastern Connecticut (Fig. 4).

The pervasive N-trending, shallowly plunging lineation seen here and at the previous stop is a dominant feature of the gneisses along the Rhode Island-Connecticut border (Fig. 3). The strong lineation is characteristic of the north-trending segment of the Hope Valley shear zone and appears to have formed as a stretching lineation related to a major component of right lateral shear on the shear zone (O'Hara and Gromet, 1985).

19.4	2.6	Beach Pond. Lunch stop. Resume trip eastbound on Route 165
26.4	7.0	Junction Route 3. Turn left onto Route 3 North
27.7	1.3	Junction Route 102. Turn left onto Route 102 North
35.3	7.6	Junction Route 117. Proceed straight on Route 102
38.7	3.4	Junction Route 114. Proceed straight on Route 102
41.7	3.0	Junction Route 94. Turn sharp left onto Route 94 North
45.7	4.0	Junction Route 6. Turn left onto Route 6 West
49.7	4.0	<b>STOP 6 LINEATED PONAGANSET GNEISS</b>

The Ponaganset gneiss is a large elongate late Precambrian plutonic body forming the western margin of the Esmond-Dedham terrane. The Ponaganset varies from a coarsely porphyroclastic biotite gneiss to fine grained leucocratic gneiss, both of which are represented in this outcrop. The coarse augen gneiss is highly lineated (N-trend,  $10-15^{\circ}\text{N}$  plunge) but weakly foliated. It is locally intruded by a leucocratic gneiss. Sharp contacts are observed, and the leucocratic gneiss shares the same strong lineation. Vein quartz and related pegmatoid bodies are present. A mullion-like structure appears to be weakly developed in the outcrop.

The large feldspar augen (up to 40 mm) commonly have asymmetric deformation tails when viewed on surfaces parallel to the lineation. The augen are circular, ovoid or tabular on surfaces normal to the lineation. Most asymmetric augen indicate right-lateral shear sense, although some with a opposite shear sense are present.

In addition to quartz, alkali feldspar, plagioclase and biotite, minor minerals in the coarse augen gneiss include hornblende, epidote, sphene, and rarer secondary muscovite. In thin section the lineation is defined by elongate aggregates of quartz and feldspar, and by trains of biotite and hornblende. Tails on feldspar augen are composed of finer recrystallized

grains. Myrmekite is commonly developed around the margins of the augen. The leucocratic gneiss contains locally abundant secondary muscovite.

This locality is approximately 2 km from the surface trace of the HVSZ, and the rocks here are among the most highly deformed found along this segment of the boundary. More moderately deformed, coarse porphyritic granitic rocks occur further to the east and represent a reasonable protolith for the gneiss here.

53.2 3.5

### STOP 7 "SCITUATE" GRANITE GNEISS

This is a medium-grained, pink biotite gneiss that is remarkably similar in appearance to the Devonian Scituate granite of central Rhode Island. Moore (1983) appropriately mapped this and other nearby granite gneisses as Scituate. Rb-Sr whole rock analysis, however, indicates a late Precambrian age. The similarity of this rock to typical Devonian Scituate granite again underscores the difficulty in distinguishing and separating late Precambrian and Devonian leucocratic gneisses in this region. We tentatively assign the rock here to the Hope Valley terrane on the basis of its occurrence to the west of quartzitic rocks of the Plainfield Formation, which we group with the Hope Valley terrane. On a lithologic basis, the rock could be equally well considered a somewhat leucocratic and non-porphroclastic phase of the Ponaganset.

The gneiss has a strong, shallow-dipping foliation (NNW strike, 15<sup>0</sup>W dip) and a subhorizontal NNW-trending lineation. Streaks of biotite help define the lineation and can be seen on foliation planes. A finer grained and more leucocratic gneiss is present low in the road cut on the south side of the road. It is slabby in appearance and grades upward into the coarse gneiss typical of the rest of the outcrops here. Greater shearing appears to have occurred in the finer grained gneiss, although it is unclear whether its fine grain size is the cause or the effect.

In thin section, quartz, perthitic alkali feldspar, plagioclase and biotite are accompanied by minor hornblende, allanite, and epidote. In contrast to other gneisses along the HVSZ, minerals in this gneiss have retained large amounts of intracrystalline strain. Quartz grains (0.1-2 mm) have high aspect ratios (5:1), sutured grain boundaries and intense undulose extinction. Recovery is very limited. Perthites have irregular shapes and only minor recrystallization about their margins. Tabular plagioclase have bent and microfaulted lamellae. These features all suggest lower grade conditions during deformation than at the other localities visited. It might be significant that this locality is somewhat west of the HVSZ (surface trace 4.5 km to the east) and quite close to the Lake Char fault (1 km to the west).

53.4 0.2  
71.3 17.9  
72.7 1.4  
79.5 6.8  
81.5 2.0  
84.9 3.4

Take entrance ramp to I-395 North (old Route 52)  
Lake Char on right  
Take Exit 2 to Route 16 East  
Town of Douglas, Massachusetts. Make left at intersection to stay on Route 16 East  
Town of East Douglas. Continue East on Route 16  
Junction Route 146. Enter Route 146 North.

85.6 0.7

**STOP 8 LINEATED PONAGANSET GNEISS**

This strongly lineated porphyroclastic biotite gneiss has a well developed rod or pencil structure. The lineation (NNE trend, 25-30°N plunge) is defined by rods of quartz and feldspar and streaks of biotite. A foliation (110°/30°N) is also present.

The linear fabric seen here is a regionally extensive characteristic of the N-striking segment of the HVSZ. This outcrop is 32 km north of Stop 6, and 55 km north of Stops 4 and 5 where we first observed strongly lineated Ponaganset and Scituate gneisses.

90.8 5.2

**STOP 9 GREY AND PINK PONAGANSET (?) GNEISSES  
SEPARATED BY AMPHIBOLITE LAYERS**

Two texturally distinct gneisses are present here: a fine grained (<2 mm) grey gneiss with coarse white to pink feldspar augen and a medium grained (1-10 mm) pale pink leucocratic gneiss. The grey gneiss consists of alkali feldspar, quartz, biotite, chlorite, hornblende, granular epidote and sphene, opaque and minor secondary muscovite. The pink gneiss has a similar mineralogy but lacks hornblende. The gneisses occur separately as large slabs several meters thick that are bounded above and below by highly stretched amphibolite layers. All these rocks have a strong foliation (95°/20°N) and a weak lineation (N/20°N). The gneisses typically become very fine grained and laminated within 10 to 20 cm of the amphibolite layers. The fine grained margins of the slabs are seen in thin section to be totally recrystallized into a fine grained (.05-.1 mm) equigranular mylonite that has the same mineral assemblage as the slab interiors. That is, the fine grained margins are more highly sheared equivalents of the interiors.

The outcrop relationships observed here provide convincing evidence for the tectonic juxtaposition of two texturally different gneisses. The juxtaposition appears to be lithologically controlled by the amphibolite layers, which are more competent than the quartzofeldspathic gneisses they separate. It is important to note that the strike of the foliation and the sheared amphibolite layers are approximately perpendicular to the NNE strike of the HVSZ. We suspect that the N-dipping sheared amphibolite layers mark zones of south-directed thrusting within the Esmond-Dedham terrane. Asymmetric deformation tails on the feldspar augen are consistent with this. The origin of the south-directed thrusts may be related to similar thrusts recognized along the southern segment of the HVSZ. That is, as the southern margin of the Esmond-Dedham terrane overrode the Hope Valley terrane, the Esmond-Dedham terrane came under approximately N-S oriented compression and developed internal thrusts sympathetic to the major thrust along the southern segment of the HVSZ. It is notable that other similarly oriented thrusts have been recognized within the Esmond-Dedham terrane (e.g., the thrust at Snake Hill, Rhode Island: R. Kemp and N. Rast, pers. comm.).

The two gneisses here are considered to be part of the Ponaganset gneiss. The grey porphyroclastic gneiss is quite similar to (although more sheared than) typical Ponaganset. In other outcrops nearby, the medium grained leucocratic gneiss is observed to grade without break into a porphyroclastic gneiss similar to typical Ponaganset. This feature and its

sharp contrast with Hope Valley alaskite observed immediately to the north (Stop 10), lead us to group it with the Ponaganset.

A comparison to the new Massachusetts state map (Zen et al., 1983) indicates that the grey gneiss with feldspar augen was mapped as Ponaganset gneiss, but that the pink leucocratic gneiss has been mapped as "Scituate" gneiss (quotations added). Radiometric dates of "Scituate" gneiss in this area give late Precambrian ages (Zartman and Naylor, 1984). It appears appropriate to abandon the use of the name Scituate for these rocks as it is now known that the type Scituate granite is Devonian. Several problems arise when considering how the leucocratic "Scituate" gneisses relate to the other rocks of the region, and whether they should be assigned to the Esmond-Dedham or Hope Valley terranes. On the Massachusetts map, "Scituate" was considered gradational with Hope Valley alaskite, but "Scituate" also contained some rocks previously mapped with the Northbridge gneiss (usage now abandoned, with most of the Northbridge assigned to the Ponaganset gneiss). In our opinion, the leucocratic gneisses grouped as "Scituate" on the Massachusetts map include some rocks that should be considered as part of the Ponaganset gneiss (such as those here) and other rocks that probably are associated genetically with the Hope Valley alaskite. Unfortunately, as was made clear in our work further south, the assignment of leucocratic granite gneisses to specific units on the basis of field and petrographic characteristics can be subjective and inaccurate.

91.6 0.8 Take Exit ramp.

91.7 0.1 **STOP 10 HOPE VALLEY ALASKITE**

This outcrop of Hope Valley alaskite is a medium grained, leucocratic buff-colored gneiss. It has a uniform, almost bedded appearance over the extent of the exposure. The bedded appearance is due to parting planes that are parallel to a strong foliation in the gneiss ( $080^{\circ}/24^{\circ}N$ ). Fine grained muscovite is common on the parting planes. In thin section, the gneiss consists of strain free quartz, microcline, lesser plagioclase, minor biotite, sphene, muscovite and opaques.

The distinctive planar parting seen in this exposure is not present in the metaplutonic rocks of the last stop, but well developed parting is seen in several other areas mapped as Hope Valley alaskite and related rocks (e.g., Hope Valley alaskite near Framingham and Westborough, Massachusetts, and the Potter Hill granite gneiss in the Ashaway quad (Feininger, 1965), Rhode Island and Connecticut). Other Hope Valley alaskite exposures have a more massive, unstratified appearance (e.g., those at Stop 2). This suggests that the planar parting reflects a primary feature, such as layering in an originally stratified unit. A reasonable protolith for the rocks with prominent parting would be for example, a sequence of rhyolite flow and/or pyroclastic to volcanoclastic layers, whereas more massive Hope Valley rocks probably had plutonic protoliths.



## THE MIDDLE HADDAM AREA, CONNECTICUT, REVISITED\*

by

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and

Gordon P. Eaton, Texas A & M

### Introduction

It has been 27 years since we led a field trip to the Middle Haddam area on the occasion of the 50th NEIGC in 1958. That area is largely contained within the U.S.G.S. Middle Haddam Quadrangle as shown in **Appendix A**, modified after Figure 16-2 of Dixon and Lundgren (1968). The field mapping, started in 1955, had been carried on part-time by both of us while we were teaching at Wesleyan University. Thus the results and interpretations presented at that time were preliminary at best. In 1957 Rosenfeld went to UCLA and left most of the continuation of the field mapping to Eaton with Rosenfeld's minor contributions during parts of some summers thereafter. After 1959, when Eaton left Wesleyan for stints at both UC Riverside and the U.S.G.S., the field work was carried out in a much more interrupted manner with Eaton, largely supported by the U.S.G.S., still carrying the major load until completion of the Geologic Map of the Middle Haddam Quadrangle. That map has been available, open-file, from the U.S.G.S. since 1972. We currently are completing the detail work preparatory for submission of the map for publication in the U.S.G.S. GQ series with Eaton as principal author. Condensed versions of both that map and its accompanying tentative cross sections, accompanied by reinterpretation of the sequence of units, are presented here as Figures 1, 2, and 3 respectively. Table 1 is the description of the map units keyed to unit symbols appearing in those figures. For brevity, after introductory reference to a unit, its symbol rather than name will be used in the text. The purpose of this field trip will be to explore some of the geological relationships so as to expose participants to both the general bedrock geology and some of the facts that have forced us to revise a number of the 1958 interpretations.

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\*Because a "dry run" is difficult to do from Los Angeles or College Station, the road log, with locality map, will be supplied to registrees at the NEIGC. The stops will be chosen to expose field evidence for major points discussed in the text. There will be about 10 stops. Participants should be prepared for some scrambling in the woods.

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## Major structures

The Middle Haddam area contains parts of three gneiss-cored anticlines mantled predominantly by schists. Within the southern part of the Bronson Hill Anticlinorium and like most of that major structure's subordinate anticlines, two of these anticlinal structures share many of the characteristics of "mantled gneiss domes" (Eskola, 1949). On the north is the southern end of the elongate Glastonbury Dome, which extends to the north through Connecticut and most of Massachusetts, where it is called the Pelham Dome. Coming into the area from the south is the narrow northern end of the more-or-less pear-shaped Killingworth Dome. The western limbs of these two domes are truncated against the eastern border fault of the Triassic-Jurassic Newark Series. East of the Glastonbury and Killingworth Domes is the relatively narrow Monson Anticline, which extends north almost into New Hampshire and south almost to Long Island Sound, where its gneissic core forms a spur off of the Killingworth Dome.

The Monson Anticline differs in character from the two domes. Since the work of Lundgren (1964) in the Essex quadrangle to the south, the Monson Anticline should be characterized more specifically as the Monson Nappe. That nappe shows its root in the core gneisses of the subsequently developed Killingworth Dome in Essex. It must have been originally displaced toward the west like many other Acadian nappes to the north, the first one of which was brought to the attention of geologists in 1954 by J. B. Thompson in an NEIGC guidebook (cited in Thompson *et al.*, 1968, p. 218). The "Colchester Nappe," (Dixon and Lundgren, 1968; but see also Lundgren, 1964, p. 24 - 29), a large recumbent fold lying in the Hopyard Basin to the east of the "Monson Anticline" and overturned to the east, is then the backfolded part of the Monson Nappe. Thus the name, "Colchester Nappe," should be abandoned. The fold in the Monson Nappe should be called the Dickinson Creek Backfold, named for its axial exposure where a north-south stretch of Dickinson Creek in Marlborough and Colchester (Lundgren *et al.*, 1971, Map) has eroded through the schist of East Hampton (Oeh) or "Brimfield Schist" of Dixon and Lundgren (1968; see **Appendix A** here) on the gently dipping now inverted, but formerly upper, limb of the Monson Nappe. The northeast end of Section C - C' in Figure 2 cuts across the west limb of the Dickinson Creek Backfold of the Monson Nappe. It should of course be recognized here that the validity of the interpretation of the Monson Nappe depends on the correlation of units across the gneissic core containing its axial surface, a subject treated briefly in the next section. We here emphasize backfolding to the east because we show below that there is, within the root zone to the west in the Middle Haddam area, internal evidence of both the kinematic process accompanying such folding and simultaneous progressive

metamorphism during the later stages of the Acadian orogeny. Furthermore the geology of the whole of southeastern Connecticut cannot be properly comprehended without explicitly taking into account the effects of that backfolding. Although our view of its genesis (cf. Hepburn *et al.*, 1984, p. 97 - 99) differs somewhat from that of Robinson and Tucker (1982) for the area on structural trend to the north in central Massachusetts, it will be seen below that our view of the timing, nature, and importance of backfolding as a phase of the Acadian orogeny independently parallels their view that takes into account historical evidence.

While Lundgren's work in the Essex quadrangle was published just about the time of the plate tectonic "explosion," a retrospective look at his map, coupled with the later recognition of Avalonian cratonal rocks to the east, suggests that a main Acadian suture, postulating its presence, must lie either east of the Selden Neck "Nappe" or under the nappes of the Bronson Hill Anticlinorium. There appears to be continuity of structures in between. The Honey Hill - Lake Char Fault Zone, though conspicuous because of its relative recency may well have had relatively small displacement, judging from Lundgren's geologic mapping (1964) at its western terminus. Figure 16-2 (**Appendix A**) in Dixon and Lundgren (1968) describes the regional structural relationships, uncomplicated by unnecessary detail. That figure provides a fascinating view of the nappes to the east in the Merrimack Synclinorium affecting strata continuous with those of the Bronson Hill Anticlinorium.

In 1958, because of lithologic similarity to much of Og, we thought the Maromas Gneiss (Pm) formed the recumbent core of a fourth gneissic anticline between the Glastonbury and Killingworth Domes. As will be seen below, we now think that interpretation is untenable for a number of reasons. The reinterpretation of the nature of Pm is part of a rather radical reinterpretation of the geology of the whole area that integrates a number of stratigraphic, geochemical, structural, and kinematic facts into a more coherent picture. The old interpretations will be alluded to henceforth only to the extent that they are relevant to our present interpretations.

The relatively tight synclines separating the anticlinal masses of gneiss show complications because of a history of multiple folding evident both within the area and to the north and south. We infer that important folding in the Middle Haddam Area took place in the Taconic, Acadian, and Alleghenian orogenies. In 1958 we recognized that the Collins Hill Formation (Oc) lay in a proto-Great Hill Syncline before deposition of the Silurian Cough Formation (Sc) and that Acadian folding of Sc, the Silurian Fitch Formation (Sf), and the Devonian Littleton Formation (Dl) into the NNE - trending Great Hill Syncline

between the Glastonbury dome and the Monson Nappe took place about an axis now slightly clockwise from, but almost coincident with, that of the earlier folding. The ancestral syncline and proto-Monson Anticline to the east must have served as guiding loci for placement of the later Monson Nappe, of which the original Great Hill Syncline must have been the pre-doming, west-opening complementary underfold. While the axial surface of the isoclinal Great Hill Syncline dips WNW today and its axis projects into the sky to the south of Great Hill, the dip of its axial surface must steepen downward, reverse its direction of dip to ENE, and come back to the surface within Oc in the Ivoryton Synform between the Killingworth Dome and the Monson Nappe (Lundgren, 1964, Geologic Map). Such reorientation at the earth's surface is a likely consequence of the fact that the downward extension of the axial surface of the Great Hill Syncline coincides with the east-northeast-dipping part of the west or lower limb of the gently north-plunging Dickinson Creek Backfold (cf. Appendix A) and/or the perturbation due to the presence of the Killingworth Dome. Also, if the axial surface did not bend in this way, it would be expressed to the south in folding of contacts between older units to either the east or west.

The other syncline of interest is the post-Acadian almost NNW-trending nearly isoclinal Maromas Syncline with its axial surface dipping NE except where it borders the SW and W flanks of the Glastonbury Dome. As outlined by the base of Pm that syncline appears to project into the sky between the NE flank of the Killingworth Dome and the Monson Anticline. Also its axial surface must lie in the Ivoryton Synform. In fact the folding of the axial surface of the early-stage Great Hill Syncline by development of both the (here) gently north-plunging late-Acadian Dickinson Creek Backfold and the Alleghenian Maromas Syncline would seem to have reoriented the axial surface of the Great Hill Syncline, as discussed above. Hence the major part of the difference in orientation between the Great Hill and Ivoryton Synclines. To a limited extent, earlier synforms and contrasts in basement rocks (as between the Glastonbury and Haddam or Monson gneisses) would seem to have served as zones of weakness for development of later synforms.

### **Rock units and their relationships**

Table 1 and Figure 3 concisely describe the units and their relationships as presently interpreted, and we rely on them to communicate that information.

Those familiar with our 1958 interpretations will notice significant changes. The geochronologic work of Brookins and Hurley (1965) and Brookins (1980) establishing the ages of certain units, particularly Pm (the Maromas Granite-Gneiss of Westgate, 1899) as Pennsylvanian - Permian and the

pegmatites of the area as both Devonian and Permian, has perhaps had the greatest influence in this regard. Those findings were reinforced by our subsequent field work showing that Pm truncated the cotichules of Bible Rock Brook (Obc) and the metavolcanic rocks of Bible Rock Brook (Oba) and that therefore Pm and the amphibolite of Reservoir Brook (Pr) are the youngest metamorphic rocks in the area, lying in the core of a syncline. Whether that contact is depositional or intrusive then remains to be decided. The discordance in turn agrees with the fact that the north-northeast-trending, Acadian, Great Hill Syncline, folding rocks as young as Devonian and, in plan oriented almost perpendicularly to the Maromas syncline, is not reflected where it projects into the latter structure. Consistent with a stratigraphic interpretation of Pm is the presence in Rhode Island not far to the east of protolithically similar volcanics of the right age, the Wamsutta Volcanics. Attractive to a volcanic interpretation is the association of Pm as volcanic protolith with the less metamorphosed slightly gneissoid large subangular clasts of volcanic-like quartz-porphyry observed in the early Jurassic Portland Formation (Jp). The Wamsutta Volcanics and these abundant large pebbles and boulders cry out for geochemical comparison by isotopic methods to test their compatibility with Pm. While we favor interpreting the discordant relationship of Pm to represent an angular unconformity, another credible possibility is that of an intrusive relationship, favored by Westgate (1899). If so, because of the gravitational evidence discussed below, the intrusive must be in the form of a folded thin tabular mass. We slant against the Westgate interpretation because of presence in Pm of mica schist septa, interpreted as metasedimentary layers and the above-mentioned evidence of the clasts. Perhaps more ambiguous is the relationship of Pr to Pm. It is possible that Pr could have been emplaced originally as a mafic sill, but there is still ambiguity in the field relations of Pr.

We retain our belief that Sc is unconformable upon the Glastonbury Gneiss (Og) (Fig. 3), based on absence of cross-cutting relationship over many miles of mutual contact, the absence of expansion of high-grade metamorphic zones into the Siluro-Devonian strata, and regional evidence of similar relationships to the north in Massachusetts and New Hampshire ( see for example: Leo *et al.*, 1984). Both on the basis of field relationships and Brookins' ambiguous isotopic data, contrary to our views in 1958, we, like Brookins (1980) and Leo *et al.* (1984), are uncertain of the age relationship between Og and Oc. That uncertainty about the age relationship at the base of Oc carries over to the character of its relationship to the Middletown Gneiss (Omi). Formerly we had interpreted discoid quartz-rich nodules in a band near its base (Ocg) as pebbles of a conglomerate, indicating, perhaps, a more important stratigraphic break than might have been the case. The fibrolitic character of

these "pebbles," possibly indicating a concretionary/ metasomatic origin, caused us to question that interpretation.

It will also be noted in Table 1 and Figure 3 that we have subdivided the "old" Collins Hill Formation into four units and that we have correlated the oldest, or "new" Collins Hill Formation (Oc) with Oeh to the east of the Monson Anticline. The youngest of the new units, the calc-silicates and schists of Bodkin Rock (Osb), has been correlated with the Hebron Formation (Oh) to the east. Both of these units may well be as young as Silurian if Robinson and Tucker (1982, p. 1738) are correct. After we had made Osb - Oh correlation, we asked ourselves: "What then happens to Obc and the metavolcanic rocks of Bible Rock Brook (Oba) to the east?" The outcrop along the contact between Oeh and Oh is not particularly good, and the contact barely enters the area on the east. A few years ago we decided to check more closely the most promising hilltop along that contact and, to our delight, found a very thin amphibolite and some loose blocks of coarse garnet-rich rock (coticule?) whose probable nearby source bears an uncertain stratigraphic position relative to the amphibolite. There is need for further field checking.

### Metamorphism

The metamorphism of the Middle Haddam Area bears both similarities and dissimilarities to other areas to the north. To the north the main observed metamorphic imprint is Acadian. In the Middle Haddam Area Pennsylvanian rocks are metamorphosed, probably into the sillimanite zone (while the schist opposite the entrance to Hurd State Park (Ph) may be of insensitive composition, amphibolites of Pr appear to be of the character commonly found in the sillimanite zone). W-side-up snowball garnets in Ph on the E limb of and approximately coaxial with the Maromas Syncline indicate the syntectonic character of that metamorphism. That Permian metamorphism and its accompanying tectonism cannot extend very far north of the Middle Haddam Area; for E-side-up snowball staurolites in DI on the E limb of and rotationally coaxial with both the Great Hill Syncline, where it intersects the New London Turnpike ( Route 2 ), and the Dickinson Creek Backfold indicate a relatively undisturbed, syntectonic, Acadian metamorphism at the latter locality. Thus at least two metamorphisms have affected the area. It is uncertain whether Taconic metamorphism has affected Ordovician and older rocks of the area.

Isograds are delineated on Figure 1 without regard to their time of development. It should be noted that, where a relatively high metamorphic grade is most recent, it will tend to mask an earlier lower grade. In such a case the polymetamorphism will be hard to determine. There are two well-defined

staurolite - kyanite zones, one along the west sides of the Glastonbury and Killingworth Domes and the other essentially coincidental with the Siluro-Devonian rocks [ Snyder (1970) has found sillimanite within these rocks on the east limb of the Great Hill Syncline just north of the area]. Determination of isogradic locations is probably strongly biased by the distribution of those rocks sensitive to their determination. Thus location of the sillimanite isograd along the east flank of the Glastonbury Dome is less than satisfactory because the core gneisses don't have appropriately sensitive compositions, and the bounding Siluro-Devonian rocks are clearly already in the staurolite - kyanite zone along their western side. East of the areas in the staurolite - kyanite zone rocks of sensitive composition commonly contain sillimanite. In the areas so delineated in Figure 1, it is not rare to have both kyanite and sillimanite in the same rocks, the latter in some cases obviously formed at the expense of the former. In 1958 we associated that reaction with a looping pressure - temperature - time path in which tectonism transported the rocks from the kyanite field of stability into that of sillimanite during a single metamorphic event. Now we do not rule out polymetamorphism as a causative factor, as it is easily possible that the Ordovician rocks of the area could have been affected by as many as three epochs of Paleozoic tectonometamorphism: the Taconic, the Acadian, and the Alleghenian. Just inside the east edge of the area is the orthoclase - sillimanite isograd, difficult to locate in many rocks of the right composition because of extensive retrograde metamorphism, evident in coarse, unbent, transverse muscovite flakes. Extensive retrograde metamorphism has also affected the rocks up to a few hundred meters east of the Eastern Border Fault (Figure 1).

Immediately west of the Eastern Border Fault of the Mesozoic, coarse relatively angular (therefore near-source) rock clasts in Jp of units recognizable east of that fault bear witness to the vertical distribution of metamorphic grade in the Eastern Highlands near the fault. This conclusion follows because of one noticeable contrast, apparently missed by Krynine (1950). Where the clasts have sensitive compositions, they are uniformly of lower metamorphic grade than is observed today in the same units in the Eastern Highlands. This observation, discussed in more detail in the 1958 field guide, obviously results from a combination of three geologic processes: upward decline in the grade of metamorphism in the Eastern Highlands at the end of metamorphism, downward displacement of sediments of the Newark Series along the Eastern Border Fault during their deposition, and the net erosion of the rocks on both sides of the fault to the present surface. The "sedimentary inversion" brought about by erosion of the eastern Highlands and deposition of the resulting detritus in the Newark Series, combined with the observed contrast, mandate such an interpretation.

It is perhaps worth mentioning that specimens of metamorphic rocks from this area have been studied by many laboratories and from many points of view in order to gain insight concerning metamorphic processes. Geochemical studies have included oxygen isotope fractionation (and study of other exchange reactions) with thermometric goals and the use of unstable isotopes with geochronological goals (cf. Brookins, 1980). Garlick and Epstein (1967) used a number of our specimens from the Middle Haddam area in their pioneering study of oxygen isotope fractionation among the minerals of metamorphic rocks as a function of metamorphic grade.

Petrographic observations on rocks from the area stimulated the extension of solid inclusion piezothermometry (interestingly, a field with its roots in some observations made by Sorby in the middle of the Nineteenth Century) to a common metamorphic combination, quartz - garnet (Rosenfeld, 1969, p. 318 - 320, 335 - 340). Adams *et al.* (1975, p. 593 - 596) used a specimen of Sc from the east limb of the Great Hill Syncline near the New London Turnpike as an important check on the internal consistency of observationally and experimentally based (for the first time) solid inclusion piezothermometry with experiments on the aluminum silicates, field observations of their occurrence, and oxygen isotopic information obtained for the same and nearby specimens. The combined data, with some redundancy, indicated a depth of metamorphism for that specimen of about 17 km near the temperature maximum of its pressure - temperature - time path.

### **Pegmatites**

The Middle Haddam Area is famous for its pegmatites. Minerals from some of those pegmatites have played a prominent role in the development of radiometric dating from its launching early in this century by Rutherford, Boltwood, and Holmes. In 1958 it could be said that this area "probably contains more pegmatites on which absolute age determinations have been made than any other similar area in the world" (Stugard, 1958, p. 650). As in the case of other geochemical/ petrological methods, the area has played the role of "Rosetta Stone" for comparison of many geochronological methods; and, indeed, one of the principal initiating motivations for our study of the geological relationships around those pegmatites was to put the quantitative data into a full-blown geological context and pose new questions to test the skills of the quantitative geochronologists. Brookins (1980) has established what had previously been inferred on geological grounds alone, namely that the abundant highly deformed pegmatites of the area are of Acadian age and that the relatively undeformed ones, famous among mineral collectors for their beautiful mineral specimens, are of Permian age.

## Utility of gravity measurements

Our measurements of gravitational acceleration throughout the area have placed useful restrictions on our structure sections because of significant density variations among units (cf.: Eaton and Rosenfeld, 1972). Thus, we found out that the low-density Maromas Gneiss could not extend very far down. This contributed to its interpretation as a thin and sheet-like mass that was folded into an almost isoclinal syncline. Data on rotated garnets, acquired later and discussed below, support that structural interpretation. Gravity measurements also helped us infer that pelites along the axis of the Great Hill Syncline and its Ordovician progenitor extend to a depth as much as 9,000 feet some distance north of the area and that its axial surface assumes a more nearly vertical attitude at depth, consistent with the backfolding interpretation above. Gravity measurements also helped support the interpretation that the structural basin (Hopyard Basin) containing the backfolded Monson Nappe is a shallow one with relatively dense schistose units extending no deeper than about 2,500 feet.

## Rotated porphyroblasts, fold kinematics, and tectonics

With discovery of the utility of snowball garnets in unscrambling tectonometamorphic sequence and kinematics in Vermont (cf.: Rosenfeld, 1968, 1970; Hepburn *et al.*, 1984, p. 93 - 101), in 1959 we tried out some of the procedures used in Vermont on unit D1 in the Great Hill Syncline where it crosses the New London Turnpike. While snowball garnets were not observed, large strike-parallel elongate staurolite porphyroblasts were found to show snowball microstructure. Further that microstructure indicates something startling about the kinematics within the Great Hill Syncline during growth of staurolite. *The sense of rotation does not reverse across the axial surface as might be expected during the flexural folding of a syncline.* Rather the sense appears to be east-side-up and invariant with position in the syncline. That is, the under side of the Great Hill Syncline had been displaced upward relatively. This perplexing relationship remained a provocative fact for many years, with provocation amplified when one of us made observations of the same kind in the Piora isoclinal syncline of the southern Gotthard region, Switzerland, in 1963. *Finally* it was realized that the peculiar tectonometamorphic phenomenon was a synmetamorphic kinematic record of the rotations accompanying backfolding, or *retrocharriage*, in the Alps (Rosenfeld, 1978; 1985, p. 445 - 449; figure 4) *after* the syncline had become isoclinal. There had been an inclination from the early part of this century to attribute the rotational senses to the overriding of the Alpine nappes, even though there was much evidence that peak metamorphism had *post-dated* those structures. Also the phenomenon

evidently had the same significance in eastern Connecticut that it had in the Alps.

The relationship to *retrocharriage* appears to be as follows (viewing the developing cross sections in both areas in the direction of counter-clockwise rotation for the ensuing snowball porphyroblasts, i.e. to the north in Connecticut and to the east in the Alps): First there is the over-riding of the nappes toward the left, somehow related to a rightward dipping suture at depth and presumably near the root zone of the highest nappe. Then there is the development of the domes/massifs due to some combination of regional compression and buoyant upthrusting. This is accompanied or followed by regional clockwise rotation of the substrate lithosphere as, to the left, the lithosphere starts to override a second right-dipping suture, causing the rightward gravity-induced synmetamorphic slump or *retrocharriage* of material at higher crustal levels atop the domes. It is here that we appear to differ with Robinson and Tucker (1982, p. 1739), who seem to accept a stage of leftward subduction ("westward underthrusting") to explain the backfolds after the main rightward subduction along a right-dipping suture accompanying nappe-formation. We credit the *retrocharriage* to a secondary gravity-sliding *without need for change of suture orientation, only a shift in suture location to the left*. The counter-clockwise rotation evident in the porphyroblasts within the isoclinally folded schist-filled synclines and/or nappe underfolds is caused by shear between the less easily deforming gneiss-cored anticlinal masses as the latter are rotated clockwise due to the *retrocharriage* that simultaneously causes the rightward rolling-back or backfolding of the preexisting nappes. The same rotation accounts for the clockwise rotation and overturning of the axial surfaces of the intervening isoclinal synclines. The rightward shift of thermally blanketing material at shallower levels helps explain in both cases the positioning of isogradic surfaces for their subsequent intersection with the earth's surface during uplift and denudation. Thermal relaxation would cause isograds during development (approximately isotherms for devolatilization reactions) to rotate counter-clockwise relative to the buried rocks, with higher grade rocks on the right- or under- sides of the subsequently left-dipping isograds.

During the period of perplexity concerning rotational senses of porphyroblasts in the Great Hill Syncline, Dixon and Lundgren (1968) had discovered to the east the Dickinson Creek Backfold mentioned earlier. That fold bears the same spatial and kinematic relationship to the Great Hill Syncline that the Chiera and Alp Campolungo Backfolds bear to the Piora Syncline. Thus *retrocharriage*, noted at the same time as the Dickinson Creek Backfold, but offset west in Vermont (Rosenfeld, 1968, p. 200), is apparently

an important aspect of the later stages of the Acadian Orogeny in parts of New England (for more on a relationship to plate tectonics in Vermont, see: Hepburn *et al.*, 1984, p. 93 - 101) as it is of the Alpine Orogeny in Europe.

We also looked at the Maromas Syncline from the standpoint of rotated porphyroblasts. That fold shows evidence of having behaved kinematically like a "normal" flexure-slip syncline with rotations of opposite senses on its two limbs during the Alleghenian tectonometamorphism. Thus, in Ph on the east limb near the entrance to Hurd State Park, snowball garnets show west-up rotation like the asymmetric chevron folds with which they are associated. In an amphibolite of the Collins Hill Formation, Oca, on the southwest limb, rotated garnets show about the same amount of rotation with the opposite sense, again like the asymmetric chevron folds with which they also are associated.

There is much more to be done with rotated porphyroblasts in the Middle Haddam area, as we have not done a thorough study of their behavior throughout the area. There may yet be surprises.

### Summary

Figure 3 concisely summarizes the geological history of the area. Within the Paleozoic there are three major breaks, each consisting of major tectonism followed by substantial erosion. There is also strong evidence that, at least, the latest two tectonisms were accompanied by medium- to high- grade metamorphism. The earliest tectonism, the Taconic, not well defined within the area, was apparently fairly closely coincident in orientation to that of the later Acadian. Late-stage backfolding was an important phase of the Acadian Orogeny in eastern Connecticut. The latest Paleozoic orogeny, the Alleghenian, was strongly discordant; and the boundary of its effects is not well defined. Given that boundary's presence west of New Haven, its fairly well-defined location to the east in Rhode Island, and the evidence mentioned earlier along the New London Turnpike, it seems probable that the Middle Haddam area is very near that boundary. It would seem that there remain many opportunities for isotopic geochemists in further testing some of the, what may appear to be, rather speculative interpretations that have been put forward above. There's probably even some opportunity for more field study, especially taking advantage of rotated porphyroblasts.

## Acknowledgments

Most of our acknowledgments are of those who helped us almost three decades ago, especially our then contemporaries working in nearby areas and areas to the north: Janet Aitken, Bobbie Dixon, Dick Goldsmith, Larry Lundgren, John Lyons, John Sanders, George Snyder, and Jim Thompson. In particular, it is important to recognize the significance of the field work of Lundgren and Goldsmith, who took advantage of Nature's gift in the form of the northward-plunging structures in southernmost Connecticut to provide some of the most important data yet for geological relationships east of the Bronson Hill Anticlinorium. Since the earlier time we have benefitted from extensive interaction with Dick Naylor, Pete Robinson, and, recently, Peter Bird. E-an Zen has been helpful with regional ideas. Since 1958 we have also benefitted from the input of a host of isotopic geochemists, cited in the references. In fact studies of this area reflect the continuing tradition of exemplary cooperation between both the field geologists and the isotopic geochemists working in eastern Connecticut that has resulted in many synergistic benefits to our understanding. Particularly we have interacted frequently, both in the field and in-house, with Doug Brookins; and he has been a principal factor in forcing us to reexamine some of our old ideas. We have also benefitted from advice and encouragement (including prodding) from Marland Billings, the late John Lucke, Joe Peoples, John Rodgers, and Jim Thompson. Also Howard Day, Ed Grew, and our friends over at Brown University, including the late Bill Chapple and Lonny Quinn, kept us sensitive to the fact that something went on during the Permian in this part of New England!

Sources of financial support for the study included The National Science Foundation, the Committees on Research of Wesleyan University and UCLA, and the U.S.G.S.

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**TABLE 1**  
**DESCRIPTION OF MAP UNITS**

- [ Jp ] PORTLAND ARKOSE – reddish-brown to pale-brown arkoses ranging from micaceous mudstones to coarse sandstones and polymict cobble and boulder conglomerates containing angular to rounded clasts of the rock units present east of the border fault (but of lower metamorphic grade), pegmatite, and, rarely, vesicular basalt. Conglomerates become more abundant and coarser, and clasts become more angular toward the border fault. Clasts show varying degrees of retrograde or diagenetic alteration. Clasts of quartzite, probably derived from the Clough Formation, are found only from the latitude of Duck Hill north.
- [ RJd ] DIABASE DIKE—dark-gray, tough, ophitic, hypersthene-bearing augite-plagioclase-magnetite diabase, locally containing biotite and hornblende; characterized by interstitial micrographic granite and microclusters of augite.
- STRONG ANGULAR TECTONOMETAMORPHIC UNCONFORMITY
- [ IPDp ] PEGMATITE, UNDIVIDED—pink to white granitic pegmatite possibly of more than one age; it contains, in addition to microcline, microperthite, and (or) albite: quartz, highly variable amounts of muscovite and biotite, and minor beryl, apatite, black tourmaline, sphalerite, and several uranium minerals. Includes foliated and nonfoliated, bodies, both concordant and discordant [not delineated on this map].
- [ Pr ] AMPHIBOLITE OF RESERVOIR BROOK—dark-gray to black, massive to flaggy, mainly schistose to granulose, biotite-hornblende amphibolite containing oligoclase or andesine; minor sphene, apatite, and ilmenite; oligoclase schist and gneiss and masses of granite augen gneiss, locally containing few dark minerals and resembling the Maromas Gneiss. Rocks show extensive alteration near the Triassic border fault.
- [ Ph ] SCHIST OPPOSITE HURD PARK ENTRANCE—very fissile, silvery muscovite schist containing rotated garnets.
- [ Pm ] MAROMAS GNEISS—very light gray to orangish-gray, massively to coarsely foliated microcline- and oligoclase-bearing biotite, biotite-hornblende, and biotite-muscovite granite gneiss, with subordinate granodiorite to quartz diorite gneiss containing minor sphene, zircon, and garnet. Some of the gneiss has large augen and flaser of orthoclase, easily observed in

curbstones in Middletown. Includes abundant lenses, strata and bands of feldspathic biotite-muscovite schist, plagioclase-quartz, diopside-garnet granofels, hornblende-biotite-quartz plagioclase gneiss, and garnetiferous hornblende-biotite schist, the largest of which may be culminations of infolded Collins Hill Formation.

#### STRONG ANGULAR TECTONOMETAMORPHIC UNCONFORMITY

- [ DI ] LITTLETON FORMATION—Mainly silvery-gray to lead-gray muscovite - biotite - staurolite -garnet-plagioclase schist, with subordinate interbedded light-gray "sugary" feldspar-quartz-mica-garnet granofels, and distinctly subordinate muscovite- biotite- garnet-albite schist and bytownite- hornblende-clinozoisite- garnet granofels. Lengths of staurolite crystals commonly exceed 2 inches; and, where elongate parallel to strike, staurolite shows "snowball" microstructure, indicating syntectonic growth. Lead-gray color due to finely disseminated graphite.
- [ Sf ] FITCH FORMATION—medium-gray, finely laminated, fine-grained calc-silicate schist and granofels containing highly variable proportions of plagioclase (commonly bytownite), quartz, biotite, muscovite, microcline, calcite, diopside, clinozoisite, tremolite-actinolite, garnet, sphene, and scapolite. Minor tourmaline, apatite, graphite, pyrrhotite, ilmenite, allanite, and zircon. Outcrops are characteristically fluted and pitted due to weathering of common calcitic layers and nodules and commonly display extremely tight folding.
- [ Sc ] CLOUGH FORMATION—white to very light, gray, muscovitic, locally garnet-bearing, quartzite and quartz pebble conglomerate, with subordinate laminae and stringers of silvery muscovite-biotite-garnet schist. Contains subordinate kyanite, staurolite, tourmaline, rutile, and zircon. Garnet in schist laminae commonly occurs as thin wafers having the thicknesses of including laminae.

#### ANGULAR TECTONOMETAMORPHIC UNCONFORMITY

- [ Og ] GLASTONBURY GNEISS [undifferentiated on map; may be older than Collins Hill Formation]

*Muscovitic phase*—light-gray, massive to moderately well foliated, prophyroblastic to augenoid, muscovite- and (or) sericite-bearing granite gneiss consisting primarily of microcline (and locally orthoclase),

oligoclase, and quartz (the aggregate total of which constitutes about 92 percent of the rock), and biotite, locally with minor sphene, garnet, zircon, allanite, and magnetite and sparse calcic scapolite. Potash feldspar commonly in the form of augen.

*Biotitic phase*—light- to medium-gray, locally orange-stained, massive to foliated, microcline-bearing biotite granite and granodiorite gneiss; distinguished from muscovitic phase by absence or near-absence of muscovite and the local presence of epidote, a trace of hornblende, and a little garnet. Locally contains minor coarse-grained allanite. Here and there are flattened clots and spindles of dark-gray hornblende gneiss of probably xenolithic origin.

*Hornblendic phase*—medium- to dark-gray, massive to subschistose hornblende-biotite granodiorite to quartz diorite gneiss; distinguished from biotitic phase by the presence of abundant hornblende, a much greater abundance of epidote and a lower proportion of microcline.

[ Oh ]HEBRON FORMATION—interbedded gray to brownish-gray quartz-biotite-plagioclase schist and pale-greenish-gray, quartz-plagioclase-biotite-microcline-actinolite or hornblende-diopside calc-silicate rock, containing subordinate sphene, graphite and, rarely, calcite; locally contains major amounts of scapolite (probably equivalent to Osb).

[ Osb]CALC-SILICATE ROCKS AND SCHISTS OF BODKIN ROCK (PROBABLY HEBRON EQUIVALENT)—Highly varied, well-bedded formational unit consisting largely of three major rock types (all showing considerable cataclasis and alteration near the Triassic-Jurassic border fault):

(1) Greenish-gray calc-silicate granofels and light-gray calcite marble containing highly variable proportions of calcite, diopside, calcic amphibole (mainly hornblende), calcic plagioclase, clinozoisite, garnet, quartz, sphene, and apatite. Although natural outcrops suggest that this rock type is subordinate to that of the schists, artificial cuts reveal that it is abundant throughout the section.

(2) Rusty-weathering muscovite-biotite-garnet schist commonly containing albite, coarse kyanite (with or without fibrolitic sillimanite), graphite (either finely divided or in coarse flakes), and staurolite; also subordinate, but nearly ubiquitous, are rutile, brown tourmaline, apatite, ilmenite, and pyrite

(3) Light-purplish-gray well-banded plagioclase-biotite gneiss and granofels commonly containing muscovite, scattered small garnets, and subordinate brown tourmaline, ilmenite, and graphite.

[ Oba]METAVOLCANIC ROCKS OF BIBLE ROCK BROOK

Dark-gray to very dark gray laminated feldspathic amphibolite and hornblende schist containing hornblende, plagioclase (An 35-An 70; not uncommonly in the form of larger insets), fine-grained garnet, and lesser amounts of biotite, diopside, epidote, ilmenite, and sphene. Locally some massive amphibolite and epidote-rich lenses are present.

[ Obc]COTICULES OF BIBLE ROCK BROOK

Pinkish-gray finely-laminated highly resistant rock (coticule) composed largely of very small garnets, quartz, and oligoclase and containing variable, but usually lesser, amounts of hornblende, biotite, cummingtonite (?), apatite, and locally coarse-grained euhedral magnetite. This rock is finely interlaminated with a subordinate gray feldspar-quartz-biotite schist and granofels. It is a prominent ledge and cliff former throughout its mapped length.

[ Oeh]SCHIST OF EAST HAMPTON (EQUIVALENT TO BRIMFIELD SCHIST of Dixon and Lundgren, 1968)—principally gray to rusty-weathering relatively coarse-grained muscovite-biotite schist commonly containing subordinate garnet, sodic plagioclase, coarse graphite, ilmenite, pyrrhotite, tourmaline, sillimanite (not uncommonly associated with orthoclase porphyroblasts), rutile, and zircon. Unit is locally characterized by the presence of pegmatitic stringers and a migmatitic appearance. Relatively poorly developed schistosity with coarse discordant muscovite due to widespread retrograde metamorphism. Quartz contains dispersed hairlike rutile needles. A less abundant rock type within this unit is bytownite-hornblende gneiss containing subordinate biotite and garnet. Probably correlative with the Collins Hill Formation. Supporting that interpretation and the correlation of Osb with Ohb is the predicted and recently detected presence of thin amphibolite and coticule along its eastern contact with the Hebron, probable easterly correlates of Obc and Oba.

[ Oc ]COLLINS HILL FORMATION (PROBABLY EQUIVALENT TO PARTRIDGE FORMATION TO NORTH) [undifferentiated on map]

*Main rock types* – Highly varied unit somewhat similar, except for the common presence of feldspathic amphibolites, to the calc-silicate rocks and schists of Bodkin Rock and consisting largely of four principal rock

types, listed in order of their decreasing abundance:

(1) Rusty-weathering muscovite-biotite-garnet schist, commonly containing sodic plagioclase, kyanite, sillimanite (commonly in fibrolitic clusters approximately equal in size to associated kyanite), staurolite, and graphite. Minor rutile, brown tourmaline, apatite, and ilmenite.

(2) Greenish-gray calc-silicate granofels containing highly variable amounts of calcite, diopside, calcic amphibole (mainly hornblende), plagioclase (mostly bytownite), scapolite, clinozoisite, biotite, garnet, quartz, sphene, apatite, and, rarely, zoisite and microcline.

(3) Gray-banded plagioclase-biotite gneiss and granofels, commonly containing muscovite, scattered small garnets, and minor rutile, graphite, ilmenite, and apatite.

[ Oca ] (4) Dark-gray to very dark-gray, massive to laminated amphibolite, hornblende gneiss, and hornblende schist, containing intermediate plagioclase (some as insets) and locally containing biotite, garnet, sphene, and ilmenite. Locally associated with laminar coticule. Some outcrops display euhedral porphyroblasts of dark-red garnet, as much as 10 mm across, weathered into sharp relief and commonly showing rotation [undifferentiated on map].

*Subordinate rock types:*

[ Occ ] (1) Beds, laminae, and contorted nodules consisting largely of pinkish-gray, fine-grained coticule (garnet-quartz granofels) with subordinate plagioclase, biotite, and, locally, cummingtonite. [undifferentiated on map]

[ Ocm ] (2) Mottled dark-gray and light-greenish rock consisting of interdigitating and irregular flat lenses, from 6 inches to several feet long, of hornblende-plagioclase rock (amphibolite) and diopside calc-silicate rock. [undifferentiated on map]

[ Ocq ] (3) Rusty-weathering muscovite quartzite, locally garnetiferous. [undifferentiated on map]

[ Ocq ] (4) Basal unit on north flank of Killington dome: very light gray oligoclase-quartz-biotite-muscovite gneiss, commonly containing considerable magnetite. Locally displays abundant small pebble-like

discoid lenses composed primarily of quartz and fibrolite, etched into relief by weathering. [undifferentiated on map]

[Omi] MIDDLETOWN GNEISS

[Omu] *Upper unit* – very light gray to rusty orange-stained layered sodic plagioclase gneiss and granofels characterized by variable amounts of the subcalcic amphiboles, cummingtonite, anthophyllite, and gedrite (commonly as large radiating clusters in the plane of the foliation). Also contains hornblende, chlorite, garnet, biotite, magnetite, and ilmenite. Subordinate amphibolite and plagioclase-quartz-biotite gneiss interlayered throughout. [undifferentiated on map]

[Omm] *Middle unit* – Light-gray to medium-gray oligoclase-quartz-biotite gneiss commonly containing hornblende, garnet, and magnetite and, less commonly, cummingtonite or orthoamphiboles. [undifferentiated on map]

[Oml] *Lower unit* – Predominantly dark-gray interlayered amphibolite and hornblende gneiss containing intermediate to calcic plagioclase, epidote, diopside, garnet, magnetite, and sphene; some lighter colored plagioclase-quartz-biotite gneiss containing cummingtonite or orthoamphibole is interlayered with the amphibolites. [undifferentiated on map]

[Om] MONSON GNEISS—light- and dark-gray interlayered sodic plagioclase-quartz-biotite-hornblende gneiss and granofels, biotite-hornblende gneiss, biotite gneiss and very dark gray amphibolite. Some of the gneisses contain microcline, garnet, and garnet, and epidote and minor allanite and sphene. Distinguished from the Middletown Gneiss principally by the total absence orthoamphiboles.

[Ohg] HADDAM GNEISS—generally light-gray thickly layered sodic plagioclase-quartz-biotite-hornblende gneiss and sodic plagioclase-quartz-hornblende gneiss commonly containing magnetite, garnet, sphene, epidote, and rarely, clinopyroxene. May be equivalent to the Monson gneiss, although it displays a very much lower proportion of interlayered amphibolites.

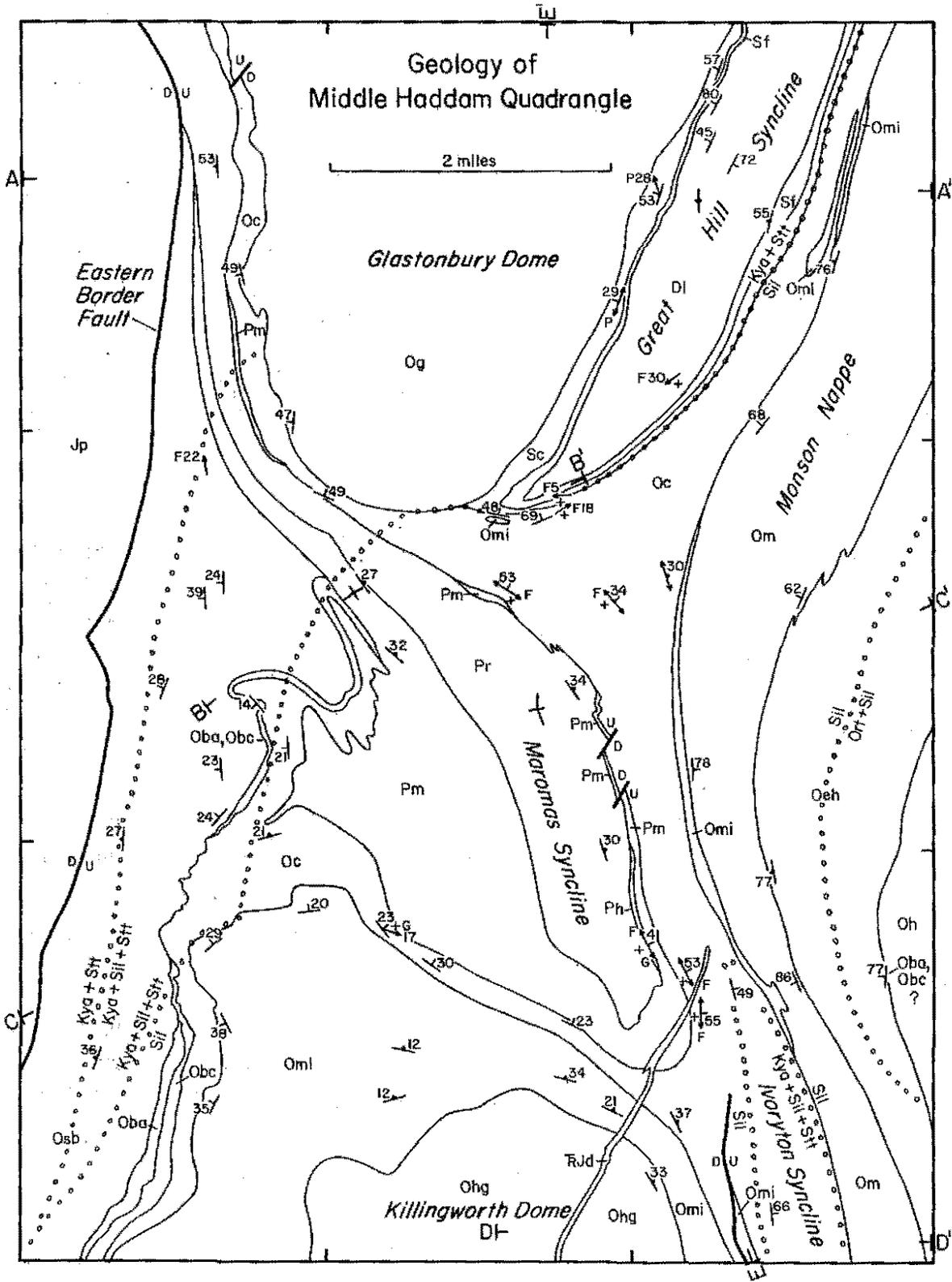


Figure 1

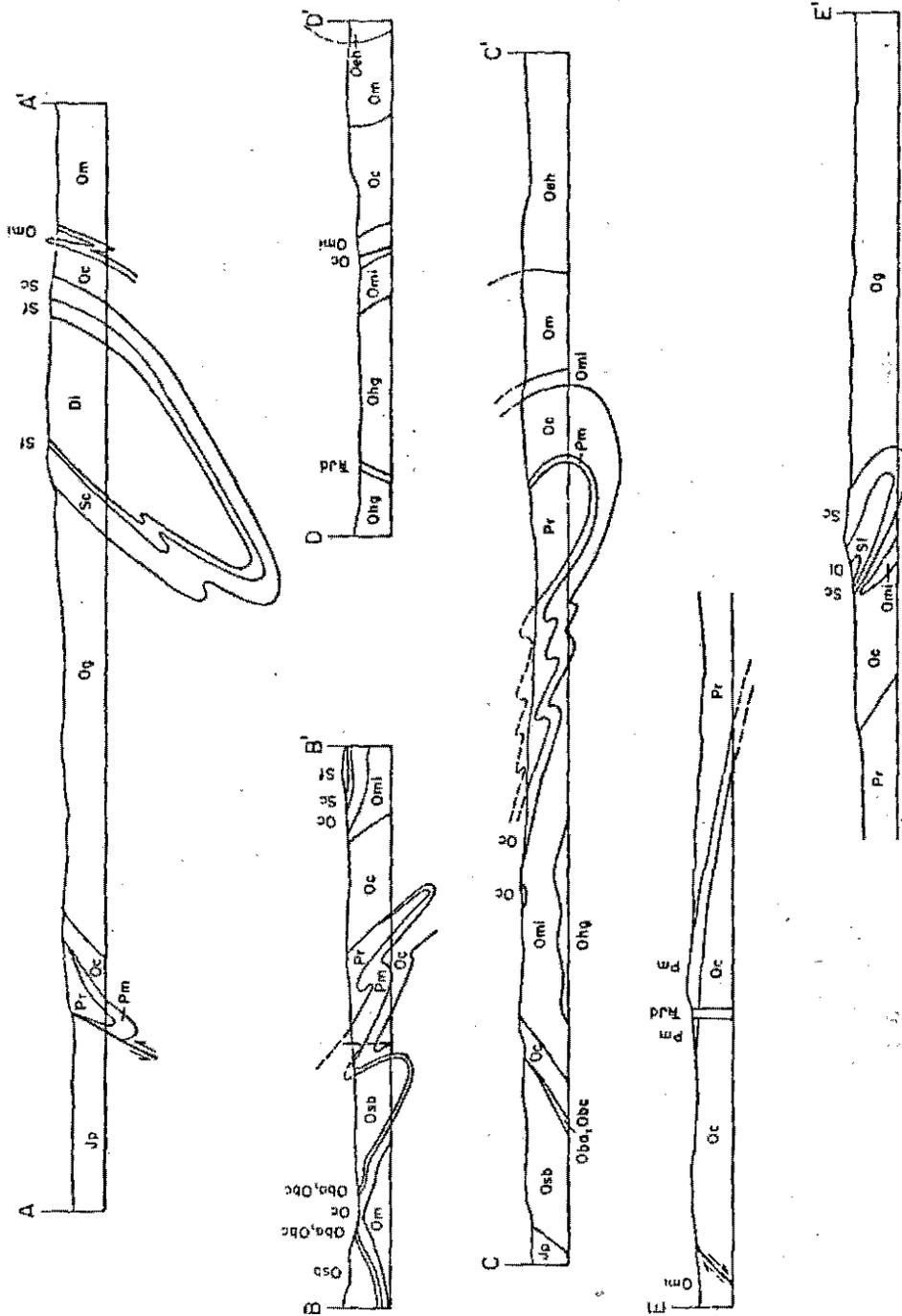


Figure 2







# DEGLACIATION OF THE MIDDLETOWN BASIN AND THE QUESTION OF THE MIDDLETOWN READVANCE

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## INTRODUCTION

The Middletown basin (fig. 1) lies along the eastern boundary of the Central Lowland physiographic province of Connecticut. The east side of the basin is traversed by the Eastern Border Fault (Lehmann, 1959), a major normal fault that separates the Mesozoic red beds of the Central Lowland from the chiefly Paleozoic schists and gneisses of the Eastern Highland. Most of the basin lies within the lowland province, which is characterized by a north-northeast topographic grain. This grain developed on a sequence of terrigenous clastic rocks and basalt flows that strikes north-northeast and dips east-southeast. The Middletown basin is defined by basalt ridges on the south, west, and north and by crystalline-rock uplands on the east. Its principal streams flow to the Mattabeset and Connecticut Rivers, which join in a large wetland north of downtown Middletown (fig. 1). The Mattabeset flows southeast across the bedrock grain, and divides the Middletown basin into two parts: a north-sloping region (south of the river), in which streams flow north to the Mattabeset and Connecticut Rivers; and a lowland region (north of the river), which comprises several buried bedrock basins, and in which streams flow generally south to the Mattabeset and Connecticut Rivers. The Connecticut River crosses the Eastern Border Fault on the east side of the Middletown basin, where it enters a channel through crystalline rocks to Long Island Sound. Just east of the fault, this channel is joined by another deep, buried channel (fig. 1) that is believed to be the preglacial course of the Connecticut River (Bissell, 1925).

Four major bodies of stratified drift are present within the Middletown basin: the Berlin clay, the Rocky Hill dam deposits, the buried channel deposits, and the Cuginchaug valley deposits (fig. 1). Age relationships among these units have been deliberated since the late 1800's, and different chronologies have been proposed by Loughlin (1905), Rice and Foye (1927), and Flint (1933, 1953). The central issue has been whether the Berlin clay was deformed by the last ice advance across New England, or whether it was deformed by a readvance during the last retreat. Ultimately, the readvance hypothesis was made convincing by R. F. Flint (1953, 1956), who related it to a late-glacial climatic reversal suggested by a pollen sequence from Durham Meadows (Leopold, 1956), located at the south end of the Middletown basin (fig. 1). This readvance, which Flint (1956, p. 276) called the Middletown readvance, has been correlated with events in the Midwest (Flint, 1956; Mayewski and others, 1981), New York, New Jersey, and Pennsylvania (Sirkin, 1967; Connally and Sirkin, 1973) and New England (Schafer and Hartshorn, 1965; Flint, 1976).

In Flint's chronology, the deposition of the Berlin clay was separated in time from the Rocky Hill dam deposits and the buried channel deposits by the Middletown readvance. In this scenario (Flint, 1953, p. 899), the Berlin clay was deposited in a glacial lake before the readvance, then overridden and deformed by it; after the readvance, the other deposits were laid down as one continuous

glaciofluvial unit adjacent to the Berlin clay. The recent discovery that all these other deposits are deltaic (Stone and others, 1982) calls into question the occurrence of a readvance, because it consequently separates contiguous lake-bottom deposits and deltas. These units are interpreted as the distal and proximal deposits of one major lake, glacial Lake Middletown, on forthcoming USGS surficial geologic maps of Connecticut (J. P. Schafer, J. R. Stone, E. H. London, and W. B. Thompson, unpublished data), the Middletown quadrangle (E. H. London and R. E. Deane, unpublished data), and the Middle Haddam quadrangles (E. H. London, unpublished data). The credibility of this new interpretation hinges on a refutation of the readvance. This trip presents the case against a readvance at Middletown, in terms of the evidence for systematic ice retreat from the Middletown basin (Part I), and the reinterpretation of features that have been attributed to the readvance (Part II). The reinterpretations are augmented by discussions at related stops. Paragraphs in each part are numbered for cross-reference in the text and stop descriptions.

Ultimately, the case for or against a readvance must account for the "till equivalent" (deformation and diamict: Simpson, 1959) in the surface zone of the Berlin clay. The study of this zone must be extended to other deposits of the Berlin clay before the Middletown readvance can be rejected or redefined. In the Berlin basin (fig. 1), for example, the evidence seems more compelling: "till" overlies varved clay in test borings (Bingham and others, 1975), ice-contact deposits overlie varved clay (Flint, 1933; Deane, 1967), and thrust faults and hook folds occur in the surface zone (from photographs XX-13, XX-14, XX-27, and XX-41 in the slide collection of the Department of Earth and Environmental Sciences, Wesleyan University). A local readvance may have occurred at Berlin, which then should be called the Berlin readvance. Such an event, however, would pose the same problem for correlation of lake-bottom deposits and deltas as at Middletown. The purpose of this trip is to stimulate interest in the question of the readvance and to motivate further study of the Berlin clay.

## PART I. DEGLACIATION OF THE MIDDLETOWN BASIN

### Depositional Settings

- (1) During deglaciation, melt water was ponded in the south part of the Middletown basin--trapped between the ice margin and terrain that sloped toward it. Sediments were released into an ice-contact lacustrine setting and deposited as deltas, subaqueous outwash, and lake-bottom deposits. Melt water escaped through the lowest outlet from each valley, and progressively lower outlets were uncovered as the ice receded. Correspondingly, stratified deposits were laid down at progressively lower altitudes.
- (2) The ice-contact heads of deposits decrease in altitude from south to north across the south part of the Middletown basin. These heads mark positions of the ice margin during deglaciation (Koteff and Pessl, 1981). Where separate heads of deposits relate to the same lake plane, they further define the shape of the ice margin during deglaciation. Relationships between deposits, lake planes, and spillways are established by adjusting their altitudes for postglacial crustal upwarp, which for the Connecticut valley has been calculated at a rate

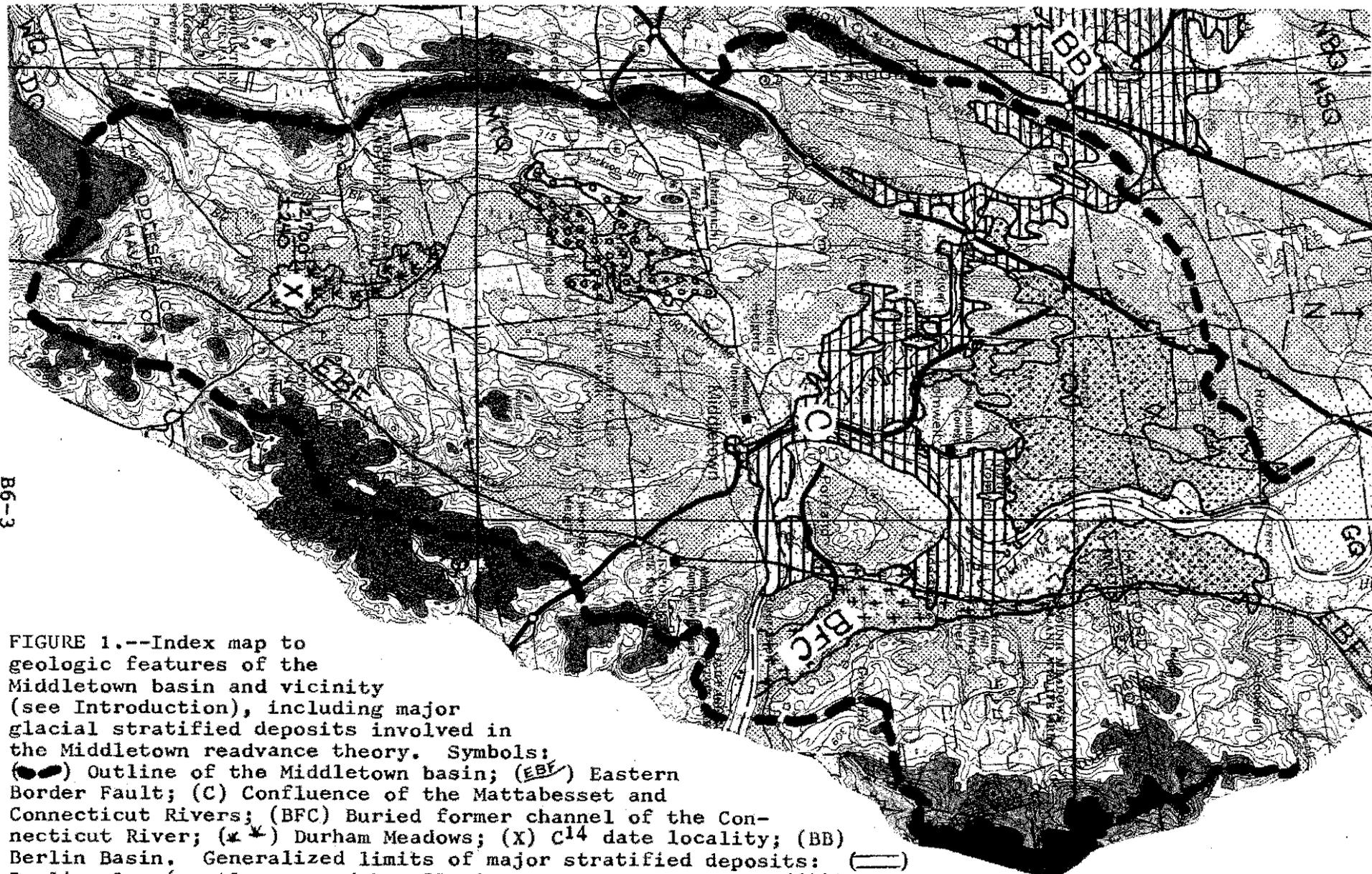


FIGURE 1.--Index map to geologic features of the Middletown basin and vicinity (see Introduction), including major glacial stratified deposits involved in the Middletown readvance theory. Symbols: (●) Outline of the Middletown basin; (EBF) Eastern Border Fault; (C) Confluence of the Mattabeset and Connecticut Rivers; (BFC) Buried former channel of the Connecticut River; (X) Durham Meadows; (X) C<sup>14</sup> date locality; (BB) Berlin Basin. Generalized limits of major stratified deposits: (—) Berlin clay (mostly covered by alluvium and swamp deposits; (••••) Rocky Hill dam deposits; (+ +) Buried channel deposits; (o o o) Coginchaug valley deposits. Quadrangle abbreviations: NBQ - New Britain; HSQ - Hartford South; GQ - Glastonbury; MQ - Meriden; MTQ - Middletown; MHQ - Middle Haddam; WQ - Wallingford; DQ - Durham; HQ - Haddam. Base map by USGS, 1979, scale 1:125,000, contour interval 50 ft. Shaded units show ranges in elevation above mean sea level:

0-100	100-250	250-500	500-750
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2  
1  
0  
2  
4  
6  
8  
10 MILES

B6-3

of 4.7 ft/mi (0.89 m/km) up toward N.20°W. (Koteff and Larsen, 1985). It is customary to interpret altitudes from 10-ft topographic contours in terms of the highest contour enclosing a landscape feature, plus 5 ft. All altitudes, thicknesses, and distances are expressed in terms of English units of measurement.

- (3) After the ice margin retreated from the north-sloping region, ponding continued in the Middletown basin because its lowest outlet, the Connecticut River channel, was blocked by glacial stratified deposits (Stone and others, 1982). Lacustrine deposition continued as a major glacial lake opened in the north part of the basin. The deglaciation history of this basin thus is representative of both north- and south-draining basins in Connecticut. Melt-water ponding was inevitable in north-drainage valleys, and it was caused by dams of stratified drift in most of the south-draining valleys in Connecticut (Stone and others, 1982).

#### Chronologic Interpretation of Glacial Melt-Water Deposits

- (4) As the ice margin retreated from the south end of the Middletown basin, glacial Lake Coginchaug opened in the lowlands of the Coginchaug valley at Durham. Lake-bottom deposits as much as 54 ft thick (Bingham and others, 1975) were laid down along the Coginchaug valley (fig. 2). South of these deposits, the lowest outlet from the valley was a 275-ft bedrock gap located at the south end of the Coginchaug valley. The stable lake controlled by this bedrock spillway is called the Durham stage of glacial Lake Coginchaug in this report.
- (5) The end of the Durham stage is recorded by an ice-contact delta in the southeast part of the Middlefield township (fig. 2; Stop 1). Its surface reaches an altitude of 315 ft. The 275-ft Durham spillway projects to a 301-ft lake level at this location. A pit at the southwest end of this landform exposes fore-set beds (without topsets) at or below an altitude of 305 ft. This delta provides significant information about the position of the ice margin at the end of the Durham stage. For this delta to have graded to the Durham lake plane, the ice margin must have abutted the land surface at no less than 301 ft; otherwise, melt water would have escaped northeast toward a lower outlet from the valley. An ice position conforming to the 300 to 310 ft topographic contours thus can be inferred for the end of the Durham stage. This position is correlated with the ice-contact head of deposits across the valley at Baileyville (fig. 2). These deposits reach an altitude of 315 ft at their south end. The Durham lake plane projects to an altitude of 308 ft at this location. These appear to be the last deposits graded to the Durham stage on the west side of the valley, and are correlated with the last delta on the east side for this reason. This correlation is shown by ice line 1 on Figure 2.
- (6) When the ice margin retreated from this position, melt water flowed northeast toward the Connecticut River—the lowest outlet from the Middletown basin. A series of spillways was carved at the intersection of the ice margin and a minor drainage divide just south of the river (fig. 2). Melt water evidently flowed away from these nickpoints through subglacial channels; a network of deep channels was carved into thick till and stratified deposits on the slope between this divide and the Connecticut River (Stop 3).

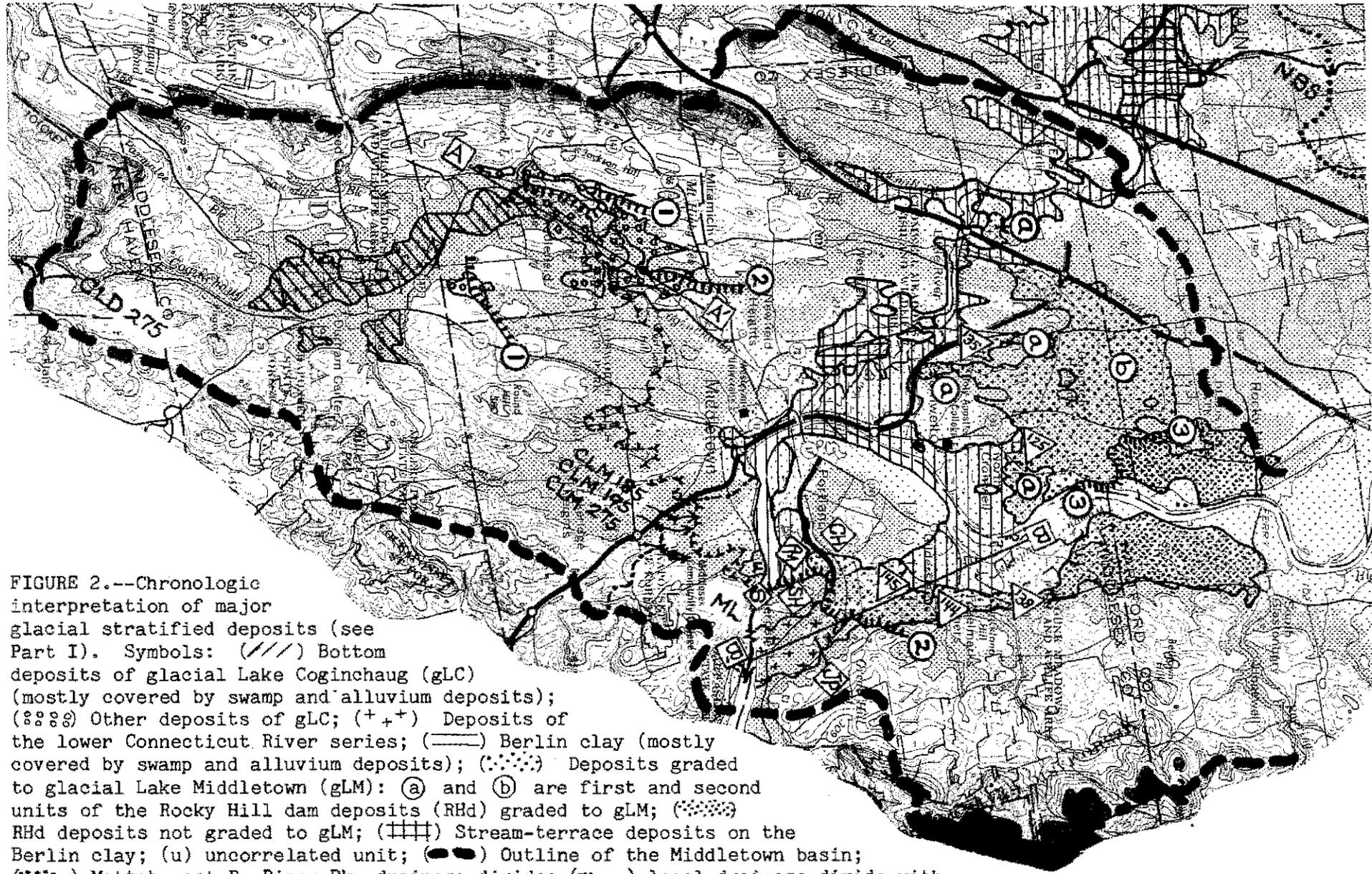


FIGURE 2.--Chronologic interpretation of major glacial stratified deposits (see Part I). Symbols: (//) Bottom deposits of glacial Lake Coginchaug (gLC) (mostly covered by swamp and alluvium deposits); (⊗) Other deposits of gLC; (+ + +) Deposits of the lower Connecticut River series; (—) Berlin clay (mostly covered by swamp and alluvium deposits); (⋯) Deposits graded to glacial Lake Middletown (gLM): (a) and (b) are first and second units of the Rocky Hill dam deposits (RHd) graded to gLM; (⋯) RHd deposits not graded to gLM; (||||) Stream-terrace deposits on the Berlin clay; (u) uncorrelated unit; (●●) Outline of the Middletown basin; (⋯) Matabbeset R.-Piper Bk. drainage divide; (---) local drainage divide with spillways for gLC; (---) glacial-lake spillway, altitude in ft (CLD - Durham stage of gLC; CLM - Middletown stage of gLC; ML - glacial Lake Middletown; NBS - New Britain spillway for glacial Lake Hitchcock); (---) Inferred ice-margin position, dashed where projected, and numbered sequentially; (△) topset-foreset contact in delta graded to gLM, altitude in ft; (S) The Straits of the Connecticut R.; (SH) Straits Hill; (CH) Crow Hill; (PM) Pecausett Meadows; (JP) Jobs Pond. Base map same as Figure 1.

- (7) The lower stage of glacial Lake Cuginchaug was controlled by these spillways south of the Connecticut River. It is called the Middletown stage in this report. This lake was unstable; it migrated downslope along with the ice margin. It comprised many "finger lakes" in separate valleys, and water levels in separate valleys may have differed at times. This possibility is suggested by the traces of melt-water channels across local divides. They may have been carved by melt water as it flowed from valley to valley toward the Connecticut River. The surface altitudes of many deposits within these valleys match up with those of the spillways south of the Connecticut River, however, suggesting that the lake was one continuous water body at times, too.
- (8) Deposits of the Middletown stage are mostly collapsed. Projected profiles of deposits in the Cuginchaug valley form an array of short segments that decrease in altitude from south to north (fig. 3). This reflects the incomplete development of outwash gradients during rapid retreat and/or the disruption of gradients by melting of subjacent ice. Pits in these deposits expose a variety of sedimentary facies cut by chiefly normal faults (Stop 2). These faults commonly extend through the surface beds, indicating that deposition took place on buried stagnant ice. The prevalence of massive to crudely bedded, matrix-supported gravel (some of which has been interpreted as flowtill; par. 29) signifies that deposition took place very close to the ice margin. The absence of large-scale foreset beds from most deposits suggests that they did not form as deltas graded to lake planes, but may have formed as chiefly subaqueous outwash (Rust and Romanelli, 1975).
- (9) After the ice margin retreated from the glacial Lake Cuginchaug spillways, glacial Lake Middletown expanded into the north part of the Middletown basin. This lake first opened in The Straits of the Connecticut River (fig. 2), located on the east side of the Middletown basin. It was dammed by glacial stratified deposits that filled the channel between Riverdale (fig. 2) and Long Island Sound. These deposits are deltaic (Stone and others, 1982). Large-scale foreset beds have been exposed at several locations along the river, especially in the Middle Haddam quadrangle. The deltas were deposited in successive, small lakes that opened in the lower Connecticut River channel. Each lake was dammed by slightly older stratified deposits to the south. This process was repeated many times between Long Island Sound and Riverdale, and continued in the buried channel in Portland. The deposits formed in this manner are called the lower Connecticut River series in this report.
- (10) A sudden shift in base-level control is recorded by the altitudes of deposits in the buried channel. Surface gradients rise and steepen northward from Riverdale to the north end of Straits Hill (fig. 2), then drop 30 ft in altitude (fig. 4). Correspondingly, topset-foreset contacts rise from 121 ft (Stop 9) to somewhere between 163 and 175 ft (Stop 10), then drop to 145 ft (Stop 4 in Stone and others, 1982). This shift marks the entry of glacial Lake Middletown into the former channel, via the gap between Crow and Straits Hills (fig. 2). This expansion of the lake was possible only after the ice margin had pulled back from the north side of Straits Hill. Before this occurred, melt water was forced to flow south across deposits in the buried channel. Depositional gradients were elevated and steepened in order to transport sediments across these deposits. As the altitudes of successive heads of deposits increased, the levels of successive lakes dammed behind them increased as well, as did the topset-foreset contacts of deltas graded to them. Topset beds in these deposits (Stops 9 and 10) are relatively coarser grained and thicker than those of subsequent deltas graded to glacial Lake Middletown (Stops 8 and 11).

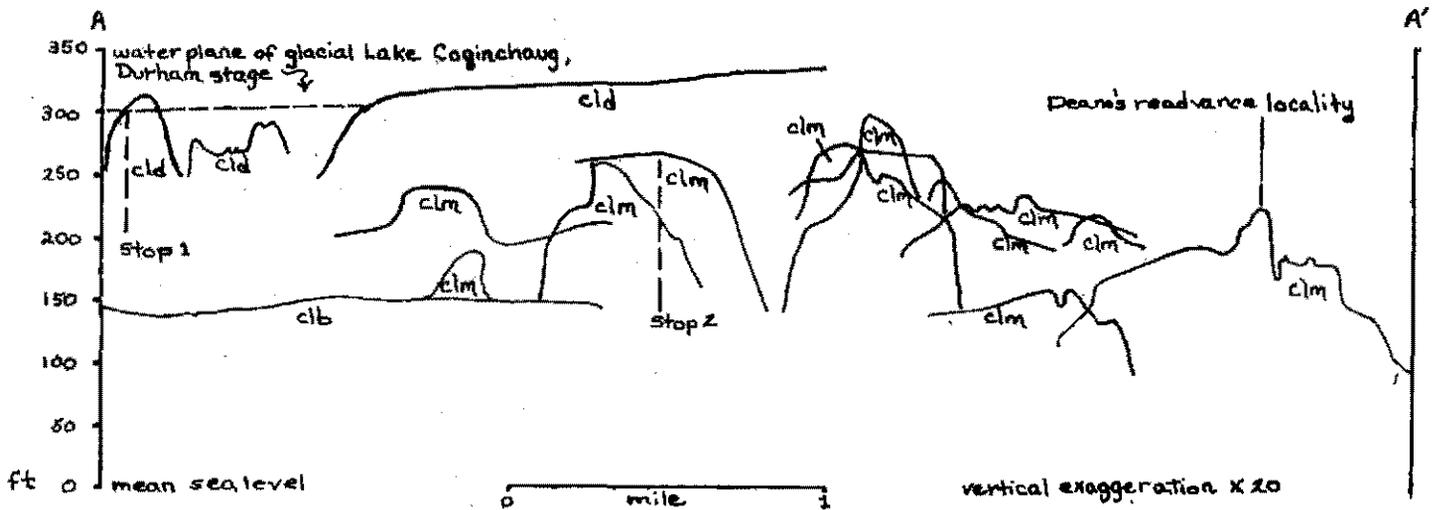


FIGURE 3.--Projected profiles of glacial stratified deposits in the Coginchaug River valley (see location, fig. 2). Maximum altitudes of deposits on either side of valley projected to an approximate longitudinal centerline. Symbols for deposits of glacial Lake Coginchaug: (cld) Durham stage; (clm) Middletown stage; (clb) bottom deposits.

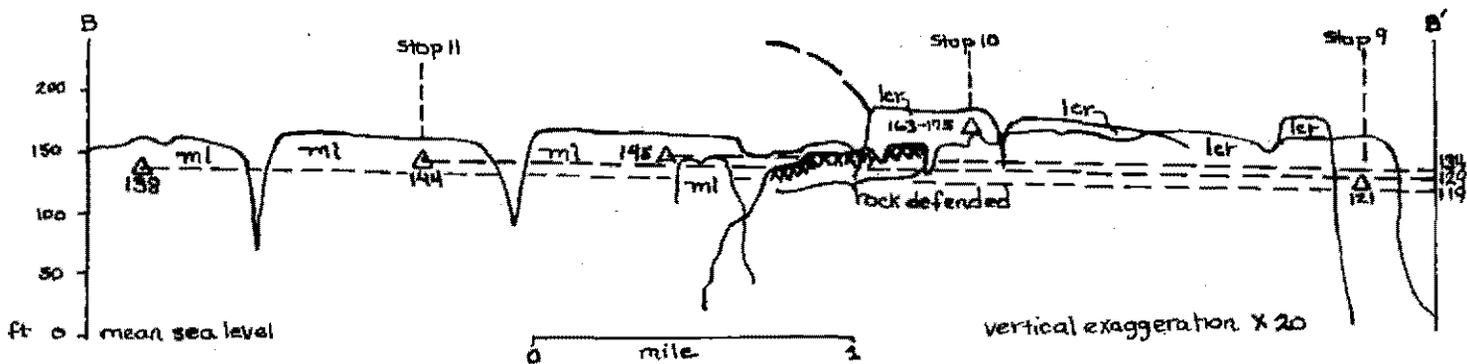


FIGURE 4.--Projected profiles of glacial stratified deposits in the buried former channel of the Connecticut River (see location, fig. 2). Maximum altitudes of deposits on either side projected to an approximate longitudinal centerline, oriented approximately N.20°W. (the direction of postglacial crustal upwarp). Symbols: (lcr) deposits of the lower Connecticut River series; (ml) deposits of glacial Lake Middletown; ( $\Delta$  121) location and altitude (or estimated range of altitude) of delta topset-foreset contact in ft; (---) inferred position of the ice margin during final deposition of unit lcr; (---) glacial lake plane, projected from delta topset-foreset contact, and adjusted for postglacial rebound.

- (11) This shift in base-level control is the basis for correlation of the last deposits of the lower Connecticut River series with the last deposits of glacial Lake Coginchaug. The last deposits of both units were laid down just prior to the entry of glacial Lake Middletown into their respective settings. The head of the lower Connecticut River series (Stop 10) reaches an altitude of 185 ft, indicating that the ice margin wrapped around the west side of Straits Hill at or above this altitude at the time of deposition. Assuming that the edge of the stagnant ice margin sloped very gently, this ice position projects to a slightly lower altitude across the Connecticut River, i.e. at or slightly below the lowest spillway for glacial Lake Coginchaug. It thus appears that the transition from glacial Lake Coginchaug to glacial Lake Middletown in the south part of the Middletown basin took place as the last deltas of the lower Connecticut River series were being deposited in the buried channel in Portland. This correlation is shown by ice line 2 on Figure 2.
- (12) As deglaciation continued, glacial Lake Middletown expanded along the Connecticut and Mattabesset River channels and along the ancestral channel of the Connecticut River in Portland. The earliest deltas graded to the lake were those deposited in the buried channel north of Straits Hill (Stop 11; and Stop 4 in Stone and others, 1982). Topset-foreset contacts in these deltas indicate an initial lake level of 145 ft, which corresponds to an initial spillway altitude of about 134 ft across deposits at Riverdale (fig. 4). One possibly contemporaneous unit is located south of the Connecticut River, just west of The Straits (fig. 2; Stop 4). Its surface altitude of 135 ft and its location just west of The Straits suggest that it could have graded to glacial Lake Middletown just after the time indicated by ice line 2 (fig. 2). Alternatively, its location at the base of a north-sloping melt-water channel suggests that it formed when the spillways for low-level glacial Lake Coginchaug were still in use. The origin and significance of these deposits will be the topic of Stop 4.
- (13) As the lake expanded, fine-grained, reddish-brown rhythmites were deposited along the Connecticut, Mattabesset, and ancestral Connecticut River channels. These deposits belong to the Berlin clay of Deane (1967), which is recognized as the lake-bottom unit of glacial Lake Middletown on forthcoming surficial geologic maps (mentioned in the Introduction). This unit is time-transgressive with respect to deltas graded to this lake. Lake-bottom sedimentation probably began in The Straits and Pecauset Meadows area (fig. 2) before the Portland deltas were deposited and continued as the Rocky Hill dam deltas were deposited. It may have diminished greatly after the ice margin retreated from the second unit of the Rocky Hill dam (par. 15).
- (14) The Rocky Hill dam deposits are an amalgamation of ice-contact deltas at Cromwell, Rocky Hill, Portland, and South Glastonbury. The deltaic origin of these deposits was not recognized until the 1960's, when they were studied in conjunction with glacial Lake Hitchcock (Hartshorn and Colton, 1967; Hartshorn and Koteff, 1968). They were named the Rocky Hill dam at this time, in reference to the role they played as the dam for glacial Lake Hitchcock.
- (15) The oldest unit of the Rocky Hill dam is a group of unconnected, ice-contact deltas built from ice-margin positions just north of the Mattabesset valley (unit a of deposits graded to glacial Lake Middletown, fig. 2). They were deposited at the jagged interface of ice and water. The altitude of a topset-foreset contact in one such delta, 135 ft (fig. 2), indicates that the Riverdale

spillway had lowered to approximately 114 ft by this time. The second unit is a massive delta that issued from an ice position in Rocky Hill and Cromwell (unit b and ice line 3, fig. 2). This delta filled a bedrock basin beneath the Goodrich Heights area of Cromwell (Bingham, 1976) and prograded southwest toward the Mattabeset valley. Access to the valley was blocked in places by drumlins perched on the south rim of this basin, so that the delta splayed into valleys leading to the Mattabeset. The level of the lake at this time is indicated by a surveyed topset-foreset contact of 125 ft in North Cromwell (fig. 2 and Stop 10). This projects to a level of approximately 103 ft at the Riverdale spillway. The third unit comprises deltas that did not grade to glacial Lake Middletown, but rather to a series of lakes controlled by spillways across the older units (e.g. Stop 6 in Stone and others, 1982).

(16) Glacial Lake Middletown ultimately extended as far north as Windsorville, Conn., about 19 mi north of the Middletown quadrangle (Stone and others, 1982). As the lake expanded, it gradually lowered, as deposits in the channel below Riverdale were trenched. After the lake reached Windsorville, it finally dropped below the level of the Mattabeset River-Piper Brook drainage divide in Newington (fig. 2). Two lakes then existed in the central lowland—glacial Lake Hitchcock to the north and glacial Lake Middletown to the south of the divide (Stone and others, 1982). The level of Hitchcock was controlled by the New Britain spillway (Hartshorn and Colton, 1967) across the divide. The altitude of this spillway evidently was controlled by the level of glacial Lake Middletown, which was the base level of erosion for the New Britain spillway.

(17) Glacial Lake Middletown drained when the Riverdale spillway was incised to the level of the lake floor near The Straits. As the lake bottom was exposed in the Middletown basin, it was traversed by streams that deposited reworked glacial sediments (Stop 7). Ultimately, the lake-bottom deposits at Middletown were trenched to a depth of 60 ft below mean sea level (Upson and Spencer, 1964), as the Connecticut River graded to a sea level below present. When the Rocky Hill dam was breached, the waters of glacial Lake Hitchcock rushed through a much deeper channel than at present. This accounts for the lack of stream terrace deposits along the Connecticut River south of the dam. Such deposits, if they exist, were buried by swamp and alluvium deposits during the Holocene sea-level transgression.

#### Summary of Findings

(18) (a) The ice-contact deposits in the Coginchaug valley were laid down in a lacustrine setting. This is deduced from the topographic setting (par. 1) and demonstrated by large-scale delta foresets in deposits at Stop 1. Deposits north of Stop 1 generally lack large-scale foresets, but their collapsed heads indicate that they were laid down when the ice margin protruded into this north-sloping valley and, hence, must be interpreted in terms of a lacustrine setting.

(19) (b) The buried channel deposits and the Rocky Hill dam deposits are deltaic in origin. Large-scale foreset beds have been exposed in pits throughout these deposits (Stops 9, 10, 11). Both units comprise numerous ice-contact deltas, deposited sequentially, and graded to different lake levels. These levels do not project to a single, stable lake plane. One major break can be discerned, however, between a series of deltas with increasing topset-foreset contact altitudes and a series graded to progressively lowering lake planes. This break occurs within the buried channel deposits. It reflects a major shift in base-

level control and demarcates a major unit boundary (between deposits of the lower Connecticut River series and of glacial Lake Middletown).

- (20) (c) Deltas in Portland and Cromwell were graded to progressively lower lake planes, as indicated by the altitudes of their topset-foreset contacts, adjusted for rebound (e.g. fig. 4). This trend reflects their sequential deposition into a single, slowly lowering lake, called glacial Lake Middletown (Stone and others, 1982). The existence of a major lake after southward drainage nominally was possible points to obstruction of the outlet from this basin. Melt-water deposits in the lower Connecticut River channel evidently choked the channel and acted as a dam. The initial altitude of the spillway across this dam can be inferred from topset-foreset contacts in the oldest deltas graded to glacial Lake Middletown (par. 12).
- (21) (d) The Berlin clay is a bottom deposit of glacial Lake Middletown. It lies within the boundaries of the lake and occupies the lowlands within these limits. It is surrounded and locally overlain by deltas graded to Lake Middletown. It thus appears to be the time-transgressive, distal facies of these deltas.

## PART II. THE QUESTION OF THE MIDDLETOWN READVANCE

### Background: Evolution of the Readvance Theory

- (22) Since the turn of the century, it has been believed that the varved clay deposits in the Mattabesset valley (the Berlin clay of Deane, 1967, p. 23) were overridden by glacier ice. Features cited were the "uneven, eroded upper surface of the clays and the local presence of kames and eskers upon them" (Shaler and others, 1896, p. 978); undulations and contortions of the clay layers, particularly at the surface (Loughlin, 1905, p. 26-27); and local beds of stony clay or crumpled clay with stones (thought to be till) on top of undisturbed varves (Antevs, 1928, p. 186-189). Prior to the 1940's, there was no consensus as to whether the clays had been overridden by the last ice advance across New England or whether it was deformed by a readvance during the last retreat (cf. Loughlin, 1905, p. 24; Rice and Foye, 1927, p. 37-38; and Flint, 1933, p. 969); but in 1953, R. F. Flint asserted that a "conspicuous readvance" had occurred in a 16-mi stretch of the Central lowland, reaching as far south as Middletown (Flint, 1953, p. 899). This conviction evidently evolved from field work conducted in the 1940's and 50's by Simpson (1959), Deane (1953, 1969), and Flint himself (1953). They found large upland areas where reddish brown, fine-grained lacustrine sediments (mostly "varved clay") were overlain by till. They assumed that the subsurface clay deposits were contemporaneous with the Berlin clay in lowlands and correlated the till above them with the deformed surface zone of the Berlin clay. Flint called the material in this zone "till equivalent, for although it is not, strictly speaking, an ice deposit, it was caused by the glacier" (Simpson, 1959, citing an oral communication from Flint in 1946). This term was used by extension for the pebbly silt and/or diamict that overlay the Berlin clay in pits at Middletown (Flint and Cushman, 1953). Flint, Simpson, and Deane thus envisioned a till sheet overlying late-glacial lake deposits on uplands and lowlands. This evidently is the "drift of the Middletown readvance" referred to by Flint (1976, Chart 3). Their reasons for inferring a readvance from these features, rather than a major glaciation, were not explained in their writings.

- (23) R. E. Deane extended the readvance till sheet south of the Mattabeset River. In both the Middletown (1953) and Hartford South (1967) quadrangles, he found two tills in upland areas, the upper one of which was clayey in the lee of lowland lake deposits. He concluded that the upper till was composed of redeposited late-glacial clay and therefore was correlative with the till equivalent in lowlands. Furthermore, he found till in outwash deposits in the Coginchaug valley (Deane, 1953). In some exposures, the till overlay outwash; in others, it was interbedded. He interpreted this as basal till separating two outwash units, one older and one younger than the readvance. He thus extended the readvance to the south part of the Middletown basin.
- (24) A seeming predilection for a readvance may have stemmed from an awareness of Midwestern glacial sequences, which featured several late-glacial readvances. These were believed to record climatic fluctuations, which by definition would have caused synchronous fluctuations of the border of the ice sheet. A search was on for evidence of climatic reversals that could be correlated to reconstruct these borders (Wright, 1971), and correlation was in fact the purpose of Flint's 1953 and 1956 papers, in which he promulgated the Middletown readvance. The pollen sequence from Durham, just south of the Coginchaug valley deposits, seemed to corroborate the readvance theory (Flint, 1956). It featured two late-glacial spruce maxima in pollen Zone A, the younger of which was interpreted as an indication of climatic reversal during late-glacial warming (Leopold, 1956). The C-14 date for the older maximum,  $12,700 \pm 280$  B.P. (W-46; Suess, 1954, p. 469), provided a maximum date for this readvance and permitted its correlation with the Valdres maximum (Leopold, 1956; Flint, 1956).

#### Reinterpretations

Many of the findings that underpin the readvance theory can be disputed in the light of recent field work and other studies published since the late 1950's.

- (25) (a) The "till equivalent" at Middletown was created by a variety of syn- and post-depositional processes. In two recently uncovered sections (Stop 7a, and Stop 5 in Stone and others, 1982), contorted bedding and massive, pebbly silt ("diamict") evidently were produced by penecontemporaneous deformation of stream-terrace sediments. Below an erosional unconformity, weakly contorted bedding in rythmites apparently resulted from load consolidation as well. From these exposures, it can be hypothesized that some of the deformation of the clay at Berlin was caused by loading. In this connection, it is important to note that many of the readvance localities mentioned at Berlin by Simpson (1959) and Deane (1967) were in areas where the clays are overlain by stream deposits (fig. 2), interpreted as an alluvial fan at the mouth of the New Britain spillway (Simpson, 1959; Deane, 1967). In another exposure (Stop 7b), surface beds were strongly deformed around a small pod of diamict. Elsewhere in this exposure, the redistribution of sediments apparently resulted from horst and graben faulting, which was caused by differential compaction. Warped beds and sub-vertical faults have been exposed at depth in the active clay pit; moreover, the lake-bottom surface at Middletown is pitted with small depressions (swamps) that also suggest differential subsidence. Although these depressions are suggestive of thermokarst, and some contorted beds (at Stop 7b) resemble frost involutions, the possibility that the Berlin clay was deformed by permafrost probably can be ruled out on the evidence that permafrost conditions did not exist in Connecticut during the last deglaciation (Black, 1983).

- (26) (b) The style of deformation in the surface of the Berlin clay at Middletown is not typical of glacier overriding (e.g. Stone and Koteff, 1979; Larsen, 1982; and Hicoek and Dreimanis, 1985). The larger faults have a vertical orientation, rather than the systematic subhorizontal orientation produced by thrusting. In addition, there is no surface till sheet on the Berlin clay at Middletown. Only one patch of diamict (Stop 7b) has been seen in the many recent shallow excavations in the Berlin clay.
- (27) (b) The correlation of subsurface lacustrine deposits on uplands with the Berlin clay in lowlands (par. 22) is erroneous. The Berlin clay deposits are relatively homogeneous; they possess the same general characteristics at the surface and at depth from basin to basin (see descriptions in Flint, 1933, p. 969; and Deane, 1967, p. 23). The surface of the Berlin clay slopes gently southeast from a general altitude of 55 ft at Berlin to 35 ft at Middletown. In contrast, the upland clays occur mainly as cores, wedges, and interbeds in drumlins (fig. 5). They also apparently occur in large patches beneath till on the uplands between Hartford and Middletown. Their composition is variable, ranging from pebbly laminites with diamict (Stop 6) to varved clay. More importantly, the upland clay deposits appear to be of different ages. Some are overlain by compact, fine-grained, jointed, iron-stained till, interpreted as lower till (fig. 5); others grade upward into compact, stony upper till (see discussion, Stop 6). Certainly, the clay beds in lower till are not the same age as the Berlin clay.
- (28) (c) Lacustrine beds may be a common component of thick till deposits in the central lowland. They have been described in drumlins in South Windsor (Colton, 1965) and East Hartford (J. P. Schafer, 1985, written communication) and seen by the author in a drumlin at Chicopee, Massachusetts (at Stop 5 in Larsen, 1982). Proponents of the Middletown readvance in the 1950's may have believed that the occurrence of clay beds in till was unique to central Connecticut and that it therefore called for a special explanation. This no longer seems to be the case.
- (29) (d) The Coginchaug valley deposits contain interbedded flowtill, which has been seen in recent exposures (now slumped). Flowtill is probably what Deane (1953) interpreted as basal till in these deposits (par. 23). The origin and significance of flowtill may not have been recognized generally at the time (Hartshorn, 1958). In one of Deane's unpublished photographs (XX-24; see collection reference in the Introduction), crudely stratified, gravelly till can be seen to overlie faulted and overturned beds of sand. The composition and bedding suggest that this is flow till. The fact that this pit was located in an isolated, high kame (fig. 3) indicates that these sediments probably were deposited in an ice hole where the deposition of remobilized till was likely (Hartshorn, 1958). In at least one location, the Coginchaug valley deposits are overlain by colluvium composed of till that moved downslope from an adjacent drumlin (J. H. Hartshorn, 1958, unpublished field review of "Surficial Geology of the Middletown Quadrangle, Conn.," by R. E. Deane, 1953). The two tills seen by Deane in the Hartford South and Middletown quadrangles were probably different lithologic or genetic facies of the upper till of Connecticut, which was not described systematically until 1968 (Pessl and Schafer, 1968).

- (30) (e) Zone A of the Durham pollen sequence no longer is interpreted as a record of late-glacial climatic fluctuations. The decline in hardwood tree pollen percentages in the upper part of Zone A is thought instead to reflect dilution by sharply increasing numbers of conifer pollen (Davis, 1967). Pollen sequences from New England now are interpreted as a record of continuous climatic warming through Zone B.

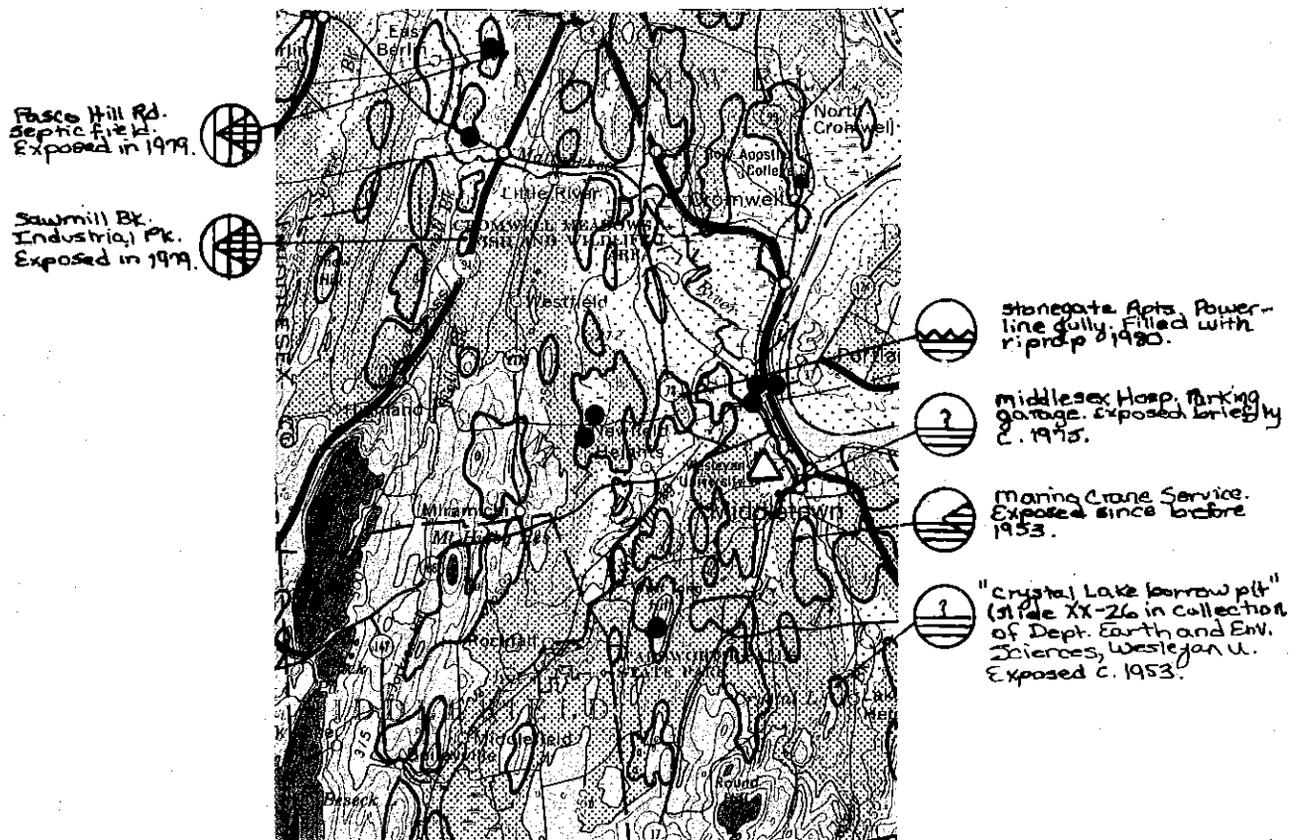


FIGURE 5.--Till/laminite localities in the Middletown quadrangle. Reported localities: interbedded laminite unit, unconformable contacts; laminite core, gradational upper contact; transported/rotated wedge of laminites; iron-stained till and laminites; details unknown. Selected borings with sub-till clay deposits: data from Bingham and others, 1975; data from unpublished reports (furnished by The Dept. of Public Works, City of Middletown). Thick till (areas where till is greater than 15 ft thick).

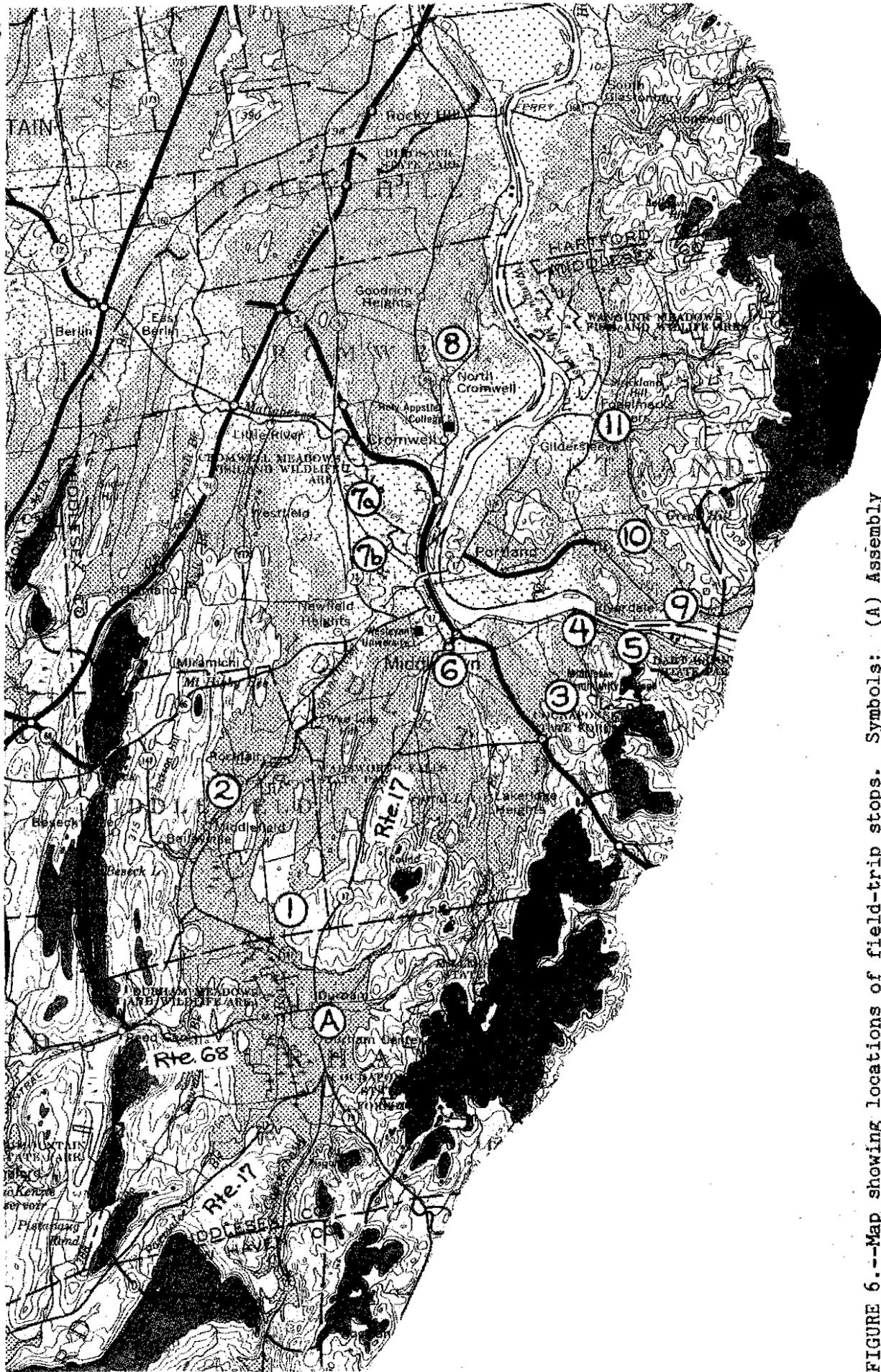


FIGURE 6.--Map showing locations of field-trip stops. Symbols: (A) Assembly point; (1) Stop number; (---) Outline of the Middletown basin.

## FIELD TRIP

Assembly Point: Coginchaug Regional High School, Durham, just south of the junction of Conn. Rtes. 68 and 17 (fig. 6). Time: 8:30 a.m.

From New Haven, take I-91 north to the Rte. 68 exit (in Wallingford). Follow Rte. 68 east to Durham. En route, enter the Middletown basin at Reed Gap (site of a large traprock quarry). Approximately 2.4 mi east of the gap, cross the Coginchaug River, the principal north-flowing stream of the Middletown basin. The lowlands north and south of the road are the Durham Meadows, lake-bottom plain of glacial Lake Coginchaug, the first major late-glacial lake that opened in the Middletown basin. From the river, travel 0.5 mi to the junction with Rte. 17; turn right. The High School is just south of the junction, on the left (east) side of the road.

We will leave extra cars in the High School parking lot. Food, gas, and rest rooms are located nearby along Rte. 17 (between Rtes. 147 to the north and 79 to the south). No special lunch stop is planned. A field-trip itinerary will be handed out at the assembly point. Stops are located in the Durham, Middletown, and Middle Haddam quadrangles (U.S. Geological Survey, scale 1:24,000, contour interval 10 ft). The general location of each stop is given in terms of quadrangle "ninths." These are formed by 2-1/2 minute intervals of latitude and longitude that divide each quadrangle into nine equal areas.

**STOP 1.** Strickland Farm pit, Town of Middlefield, Durham quadrangle (NC 9th). Access to pit is via unnamed industrial park road not on topo. (topographic map) E. of Cherry Hill Rd. (unnamed on topo.) 0.33 mi N. of the junction with Middlefield Rd. (Conn. Rte. 147). Follow industrial park road 0.25 mi to intersection with unnamed road on left; take this road N. 0.2 mi to pit (owned by Town of Middlefield).

This pit exposes delta foreset beds at the top of a "kame" at the north end of Durham Meadows (figs. 1 and 2). The Meadows are a lake-bottom plain (Simpson, 1968) near the south end of the Middletown basin. The lowest outlet from this basin, south of the Meadows, is the 275-ft gap at the head of the Coginchaug River. The altitude of this spillway projects to a level of 301 ft at the kame, which itself reaches 315 ft. This projection gives a minimum estimate for the altitude of the lake plane at this location; a column of water in the spillway probably raised the level a few feet. Much of this delta thus stands slightly below the level of this lake, which is called the Durham stage of glacial Lake Coginchaug (par. 4).

Only foreset beds have been exposed in this 10 to 12-ft deep, horseshoe-shaped cut. The foreset beds intersect the 305-ft land surface, which has been graded to a depth of 0.5-1.0 ft. Thin topsets may have been removed, but it appears that they did not overlie this part of the delta, indicating that this was the distal, submerged part. Foreset beds dip west to southwest at 9 to 36

degrees. The degree of dip increases toward the center of the kame. Along the northwest wall of the pit, the upper foresets can be traced into bottomset beds, which are overlain conformably by tabular cosets of cross-bedded fine pebble gravel. No faults have been exposed in this cut, indicating that the delta had a free front at this location, and hence probably graded to an open lake (rather than an ice-hole pond). North of this pit, the surface and flanks of this landform have hummocky topography, indicating that it formed in contact with ice. The significance of this ice-contact slope with respect to the stages of glacial Lake Coginchaug is discussed in paragraph 5.

**STOP 2.** Stowe St. pit, Town of Middlefield, Middletown quadrangle (SW-SC 9ths). Enter via "Private Drive" on N. side of Stowe St. (unnamed on topo.), 0.05 mi W. of intersection with Conn. Rte. 157 (Main St.). Owned by Town of Middlefield.

The sedimentary facies exposed in this pit are common throughout deposits of the Middletown stage of glacial Lake Coginchaug (par. 8). In this pit, two cuts expose pervasively faulted beds near the base and top of a small kame terrace. In the lower cut, beds of plane-laminated sand and massive, matrix-supported gravel are oversteepened by closely spaced normal faults. In the upper cut, beds of sand are displaced by normal step faults that extend to the top of the section (at approximately 255 ft). At different locations, the upper pit face has shown trough and tabular cosets of cross-bedded sand, cut locally by arcuate erosional scours; and interbedded massive sand and climbing ripple sequences. The dip direction of beds is generally south-southwest.

Together, these sections suggest a fining-upward sequence of fluvial deposits laid down on a mass of stagnant ice. Melting apparently took place during and after deposition. Facies at the base of the terrace are more "proximal" (coarse grained, poorly sorted) than beds above. This normally would be interpreted as a subaerial, regressive outwash sequence deposited as the ice margin retreated from this location; however, in this north-draining valley, it can be inferred that these sediments were deposited in a lacustrine setting (par. 1). Inasmuch as they lack large-scale foresets, they do not appear to have formed as a delta graded to the lake plane. They may have been deposited as subaerial outwash upon a mass of stagnant ice (that propped it above the surface of the lake), or as subaqueous outwash (Rust and Romanelli, 1975), or both. Correspondingly, the upward change in facies may reflect a change in source from the submerged, grounded base to the ice margin, initially, to supraglacial streams.

**STOP 3.** Joseph L. Carta property, 274 Bartholomew Rd., City of Middletown, Middle Haddam quadrangle (CW 9th). Enter property via dirt road on E. side of Bartholomew Rd., 0.25 mi N. of Reservoir Rd. (unnamed on topo.), or 0.5 mi. S. of Bow Ln. Follow dirt road E. past shed to fence surrounding gullies.

These two gullies are incised in thick till deposits down slope from a spillway for the Middletown stage of glacial Lake Coginchaug (fig. 2). The gullies form branching, meandering channels, generally 20 ft deep. No streams enter these gullies, and the trickle of water that flows through them presumably is fed by springs at their heads (which are covered with cut brush). The location of such underfit channels at the base of a spillway suggests that they were carved by melt water flowing downhill from this spillway to the Connecticut River, the lowest outlet from the Middletown basin. At the time when this

spillway was used, the ice margin must have abutted the landscape at or above its level; otherwise melt water would have escaped through a lower spillway (par. 1, 6, 7). These gullies originate below this level, indicating that melt water flowed beneath the ice. Farther downhill, these gullies join a larger ravine along Reservoir Brook, which flows north to the Connecticut River. Many other dry channels exist on slopes between the spillways for glacial Lake Cogenchaug and the Connecticut River. It thus appears that subglacial drainage took place in this area throughout the Middletown stage of glacial Lake Cogenchaug.

**STOP 4.** Roadside Parks pit, City of Middletown, Middle Haddam quadrangle (CW 9th). Pit is on S. side of River Rd., 0.7 mi E. of Silvermine Rd. Dirt road to pit is marked "State Property, No Trespassing."

Three major units have been exposed in discontinuous fresh cuts along the southern two-thirds of this north-northwest trending cut:

upper sand and gravel unit (2 to 6 ft thick): beds of medium to fine pebble gravel and medium to coarse sand in trough and tabular cosets; and/or medium to very coarse sand with fine pebble gravel, in massive to crudely stratified beds with many reactivation surfaces.

——conformable and erosional contacts——

middle fine sand unit (4 to 7 ft thick): interbedded massive very fine sand, plane-laminated medium to fine sand, and sinusoidal beds of fine to very fine sand. Some beds are intrastratally deformed into anticlines and synclines as much as 3 ft high. Axes of folds and sinusoidal ripples are vertical or slightly inclined to southeast and northwest.

——conformable contact——

lower sand unit (3 to 6 ft exposed above slump at base of section): variable fine to very coarse sand, granular gravel, and fine pebble gravel; in massive, planar, and graded beds, which occur in alternating coarse-fine sequences, with parallel contacts.

The origin of these deposits is uncertain. They are located at the mouth of a melt-water channel, as well as at the margin of glacial Lake Middletown (par. 12). Their interpretation as an north-building fan or delta (into glacial Lake Middletown or a tunnel beneath stagnant ice) or as a south-building ice-contact deposit awaits determination of the overall current directions in these deposits (which have not been obvious in sections). Information leading to an environmental and chronological interpretation will be sought at this stop.

**STOP 5.** White rocks "lower till" locality, River Rd., City of Middletown, Middle Haddam quadrangle (CW 9th). Road cut is on S. side of River Rd., 1.5 mi E. of Silvermine Rd. (and Town Farms Inn). Road cut is just past White Rocks quarry, on E. side of powerline crossing. Parking is very dangerous; shoulder is narrow and soft.

This is one of few well-exposed, lower till localities in Connecticut. The physical appearance of this till should be compared with that at Stop 6.

The lower till of Connecticut is a widespread but discontinuous subsurface unit (Stone and others, 1982). It is found primarily in drumlins and other thick till areas, and in pockets between bedrock knobs. It less commonly occurs as a thin plaster on valley walls--in this case, on the wall of the Straits of the Connecticut River (fig. 2).

The "red" facies of lower till is exposed in this 6- to 7-ft section, which is located 1.0 mi E. of the Eastern Border Fault. The cut is kept clean by a waterfall. The till is very compact, and the water has sculpted potholes and pinnacles into the outcrop. The color of the matrix, dark reddish brown, indicates that this till was derived from red beds of the Central Lowland, west of the fault. It is composed chiefly of very fine sand, with a 3 to 5% concentration of mostly crystalline clasts. Slabs of till can be pried loose along black-stained, subhorizontal joints. These also are exposed in the bed of the stream at the top of the section. The presence of these iron-oxide stained joints indicates that this is lower till (Pessl and Schafer, 1968). These joints are thought to represent the base of a deep weathering profile, truncated by glacial erosion during Late-Wisconsinan time.

**STOP 6.** S. G. Marino Crane Service cut, Mill St., City of Middletown, Middletown quadrangle (CE 9th). Mill St. runs between Ridge Rd. and the rotary on Conn. Rte. 17 (S. Main St.) at the entrance to the Conn. Rte. 9 connector. Marino's driveway is on the S. side of Mill St., 0.3 mi SE. of the rotary. Park at the NW. end of the parking lot and walk down driveway behind the building to the west and south cuts. The north cut is on Mill St.

This locality is an example of interbedded till and laminated fine-grained sediments. Laminites are fairly common in drumlins and upland till deposits in the central lowland. Six localities have been reported in the Middletown quadrangle (fig. 5), and more subsurface "varved clay" deposits are indicated by borings. Large deposits of subsurface clay have been mapped to the north in the New Britain and Hartford South quadrangles (Simpson, 1959; Deane, 1967).

The Marino locality features a body of laminites between two tills. The underlying till has been exposed along the base of the west wall, but at present is covered by landslide deposits and a retaining wall. This till is fine-grained and compact, with a relatively small percentage of clasts (chiefly subangular pebbles, cobbles, and boulders). The matrix contains sand lenses and streaks of silt and clay that may be deformed laminae. It is overlain unconformably by laminites, which are exposed discontinuously along the west and south walls. The locations of fresh sections change frequently as a result of rotational landslides. The laminites are composed chiefly of sand, silt, and clay. Two facies have been identified: a pebbly facies, with interbedded diamict, graded coarse sand to clay, and laminated very fine sand to clay, in apparently arhythmic sequences (pebbles are concentrated in the diamict beds but also pepper the other beds); and a nonpebbly facies, composed of mainly thin, well-sorted, apparently rhythmic laminations, with 1- to 2-ft thick intervals of contorted bedding. These two facies grade vertically and laterally into one another, but details of their relationships have not been established. Near the top of the south wall, pebbly laminites grade upward into thick, sandy couplets. The coarse members consist of non-compact, massive, poorly sorted sand with pebbles and some streaky laminations. These beds are 2 to 5 ft thick. The fine members consist of micro-laminated very fine sand, silt, and clay. These beds are 1 to 2 in thick, with sharp basal contacts and gradational upper contacts. At the top of the south wall, the stony sand is overlain by a moderately compact,

massive, stony sand with rounded pebbles and cobbles. Their long axes dip down to the west-southwest. The two units interpenetrate along the contact. The overlying till is exposed in the north wall on Mill St. It is compact, sandy, and stony, with a relatively large percentage of rounded pebbles and cobbles. It interpenetrates the laminites along faults and folds. Detached and rotated slabs of laminites appear to have been interthrust from the northwest with till. One large slab dips 12 to 25° to the north-northwest.

Evidently, an ice sheet overrode the deposits of a glacial lake at this location. The event was probably late Wisconsinan, inasmuch as the overlying till in the north cut appears to be a basal facies of upper till (cf. lower till, Stop 5). The laminites were deposited in an ice-contact lake. This is indicated by the abundance of stones and diamict beds within them. This lake may have been pro- or subglacial, or beneath an ice shelf; and the lake itself may have existed prior to the last glaciation. A more detailed interpretation must be based on a systematic study of lithofacies and contacts and comparison with developing models for "glaciogenic subaquatic deposits" (Gravenor and others, 1984).

STOPS 7a and 7b. Michael Kane Brick Co. pits, Conn. Rte. 72 (Newfield St.), City of Middletown, Middletown quadrangle (CC 9th). Gravel driveway to Kane is on E. side of Rte. 72, 0.8 mi N. of Westfield St. Driveway (through trees) is marked by small sign attached to tree.

These pits expose the Berlin clay (par. 13, 21, 22, 25, 26, 27), which is more than 50 ft thick beneath the Kane property. The Berlin clay is the lake-bottom unit of glacial Lake Middletown (par. 9, 10, 12, 15, 16, 20). Here at Middletown, the lake-bottom surface reaches a maximum altitude of 35 ft. Where the original surface is intact, it is riddled with small depressions (swamps). Around clay pits, the original surface has been obliterated by mining and by dumping of dredge spoil (and tons of discarded bricks). The mining of clay is preceded by the stripping of sediments on top of it; thus, even in active clay pits, exposures of the original land surface are rare. In recent years, three such exposures have been observed at Kane. One was described in the 1982 NEIGC Guidebook (Stone and others, 1982, Stop 5). That section is still accessible, as is the one at Stop 7a. Stop 7b now is buried, but is described because similar features may be exposed in future excavations at Kane.

STOP 7a. Take the dirt road at the N. end of the kiln buildings to the working pit. This road runs E. to the railroad tracks, then turns N. toward the pit. Park near the pit and walk to the abandoned pit to the north. Cuts a and b are in the northwest corner of this pit.

This section is similar to the one in the southeast corner of the active clay pit (Stone and others, 1982, Stop 5). At both localities, the erosional unconformities are interpreted as stream channel scours and the overlying sediments as fluvial deposits. These consist mainly of massive to crudely bedded, poorly sorted sand, with scattered pebbles, varved clay fragments, and rotated clasts of poorly sorted sand. These deposits resemble some descriptions of the alleged "till equivalent" (par. 22), but the presence of traceable laminations indicates that they are water-laid sediments. These streaky, weakly contorted, disrupted laminations appear to have foundered during deposition, probably as a result of loading. The foundering of pebble gravel beds may have caused their dispersal in the poorly sorted sand. The included varve fragments and rotated clasts of sand appear to have been eroded from stream banks (as in fig. 7, cut a).

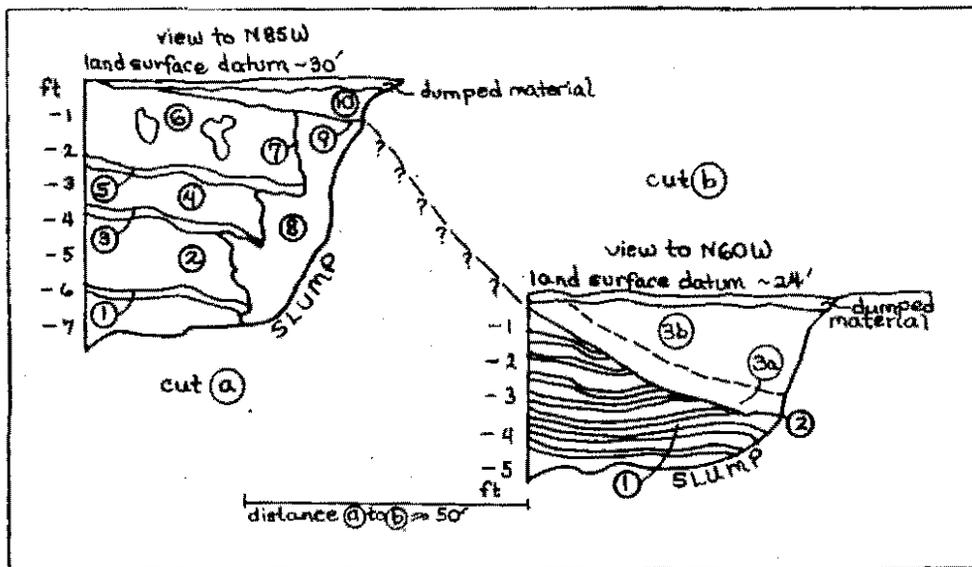


FIGURE 7.--Sketches of cuts a and b at Stop 7a.

cut a

- 10) Poorly sorted, crudely stratified to non-stratified fine pebble gravel and very coarse to medium sand
- 9) Undulating erosional contact
- 8) massive very fine sand with subangular to subrounded pebbles, inclusions of very fine sand and silt, and fragments of silt-clay beds
- 7) irregular, near vertical erosional contact, undercut beneath silt-clay beds
- 6) massive very fine sand with rotated inclusions of fine pebble gravel and very coarse to medium sand (inclusions look like material in unit 10)
- 5) silt-clay bed, pulled apart and penetrated by material of unit 6; gradational upper and lower contacts
- 4) massive very fine sand, with deformed pieces of unit 5; includes segmented stringers of medium to coarse sand and granular gravel
- 3) silt-clay bed; gradational contacts
- 2) massive very fine sand, with a bed of poorly sorted coarse sand at the contact with unit 3; material of the coarse lamination has subsided into the sand.
- 1) silt-clay bed

cut b

- 3b) fine to medium sand and very fine gravel with faint, disrupted laminations
- 3a) 10-in graded bed, fining upward from rounded pebble gravel and very coarse sand, through coarse sand, to medium to fine sand of unit 3b
- 2) erosional contact
- 1) rhythmically bedded very fine sand-silt and silt-clay; coarse members grade upward into fine; fine members have sharp upper contacts

These stream deposits become thicker, better stratified, and better sorted from northwest to southeast across the Kane property. In this direction, the surface altitude of the Berlin clay decreases from approximately 30 ft (at this stop), to approximately 19 ft in the southeast corner of the active pit, to 8 ft at the southeast end of the property. At this end, a 10-ft unit of trough and tabular cross-bedded pebbly sand was stripped from the top of the clay in 1979. The orientation of trough axes and sand-wave foresets was generally S.65-70°E. Based on this information, the materials above unconformities "9" and "2" in cuts a and b, respectively, are interpreted as the feather edge of stream terrace deposits on the floor of glacial Lake Middletown.

STOP 7b. The bulky waste landfill is located in the southwest corner of the Kane property, bounded on the W. by Rte. 72 and on the S. by a powerline. Access to the section was via a small shopping center parking lot on the south side of the powerline.

This important locality was exposed in the fall of 1982, as an old clay pit was prepared for a bulky waste landfill. The south wall of the pit was excavated to make a straight, vertical boundary for the landfill. This excavation encountered the original surface of the Berlin clay. The beds were contorted and faulted and were covered at one location by a patch of compact diamict. This cut was different from the other two localities at Kane, in that it lacked a stream terrace unit (to account for the deformation and diamict), and it more closely matched descriptions of "till equivalent" (par. 22). Moreover, it provided a relatively long, continuous exposure in which the style of deformation could be examined (par. 26).

The section measured 172 ft long by 5 to 8.5 ft high and faced northeast. A swamp was exposed in cross section at the west end of the cut. It had been filled with 3 ft of dirt spoil, which overlay a black, organic-rich layer, approximately 0.8 ft thick. This layer rested on a zone of contorted silt-clay laminae, approximately 2.5 ft thick, that had been churned by roots. Below this zone, laminations were contorted but traceable across the face. They were deformed chiefly by diversely oriented high- and low-angle normal and reverse faults with measureable slip surfaces of several inches and net displacements of 0.125 to 0.5 in. This section was cut by intersecting subvertical joints, along which black stains extended from the organic layer to the base of the cut.

Southeast of the swamp, the section consisted of chiefly fine sand laminae that were deformed into horsts and grabens (figs. 8 and 9). These structures were produced by many small-scale faults (figs. 10, 11, 12, 13); boudinage structures also were observed. Displacement patterns and directions were diverse and corresponded mostly to local movement of material downslope into grabens. The degree of deformation increased from the base to the top of the section.

In one graben, the uppermost laminae were deformed intensely into flame-like structures along closely spaced thrust faults (figs. 14 and 15). Along these faults, the deformed laminae intertongued with an irregularly shaped body of very compact, stony clay that occupied the center of the trough. This body measured 11.5 ft wide by a maximum of 4.0 ft thick. The clay was mottled with lighter streaks that appeared to be deformed laminations. The degree of interpenetration decreased from east to west along the base of the clay, and the western contact was a high-angle normal fault plane (just to the right of fig. 8).



FIGURE 8.--Center part of section at Stop 7b. Numbered intervals are meters. View to S.30°W.

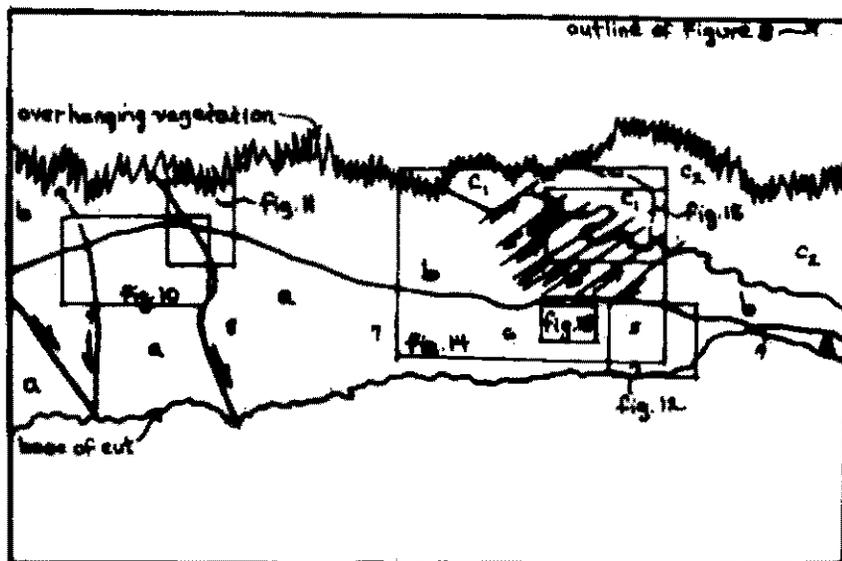


FIGURE 9.--Selected major features in Figure 8 (Stop 7b) and index to Figures 10-15. Units: a-microfaulted beds, traceable across section; b-contorted beds, sedimentary structures destroyed; c-diamict pod: c<sub>1</sub>-compact silt and sand with fine pebbles; c<sub>2</sub>-relatively stone-free, streaky silt and clay (deformed laminations?). Sediments of unit b may have slid along the surface of unit a. Unit c may have slid into a lake-bottom depression at the same time as unit b, causing them to interpenetrate along faults.

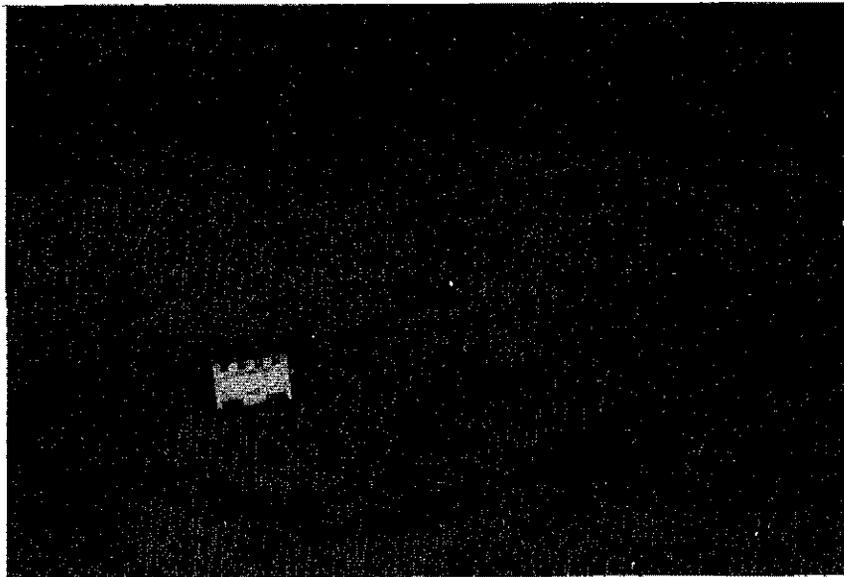


FIGURE 10.--Deformation of fine sand laminae at crest of horst (see location, Figure 9). On left, horst is bounded by a low-angle normal fault with antithetic reverse faults. Net displacement is extensional to left.

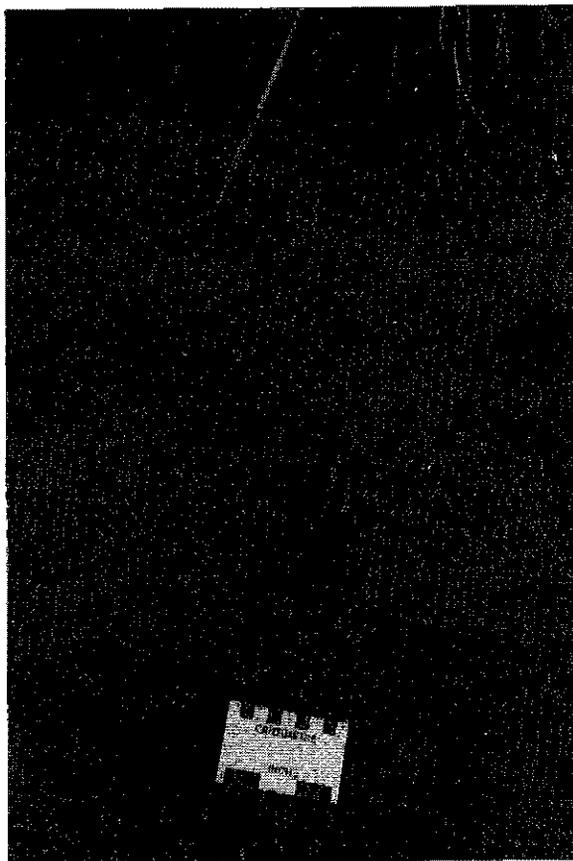


FIGURE 11.--Transition from horst to graben (see location, Figure 9). Light-colored bed is down-dropped to right by a series of step faults.

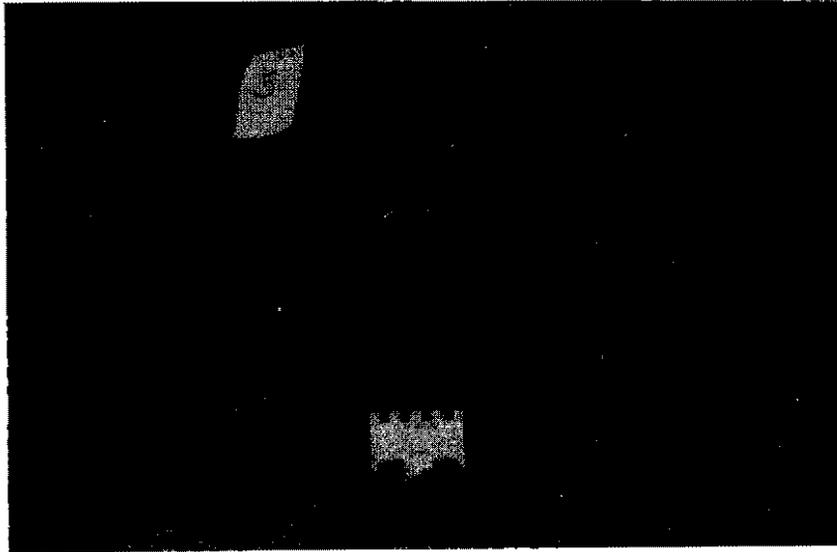


FIGURE 12.--Detail of small-scale graben (see location, Figure 9). Down-dropped, rotated block is bounded by a set of subvertical normal faults on right and by listric normal faults on lower left. Rotation of block is indicated by reverse fault on upper left.

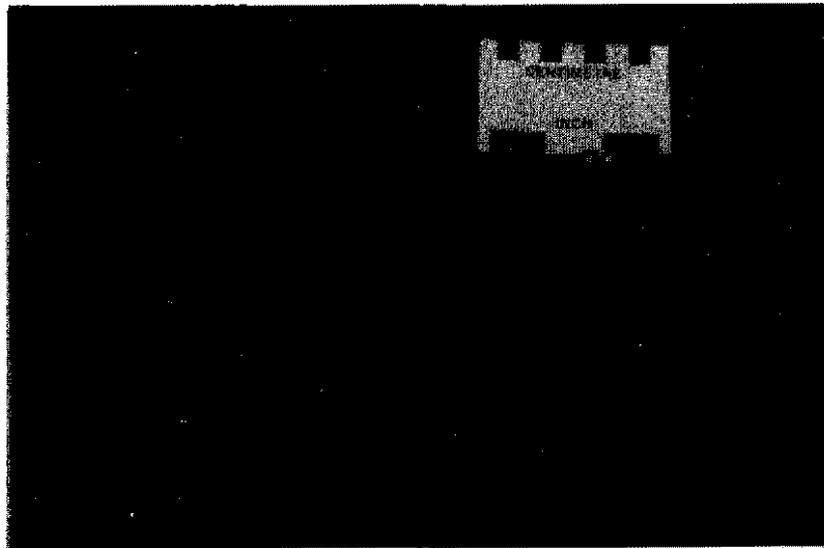


FIGURE 13.--Synthetic normal faults caused by extension and rotation to left (see location, Figure 9).

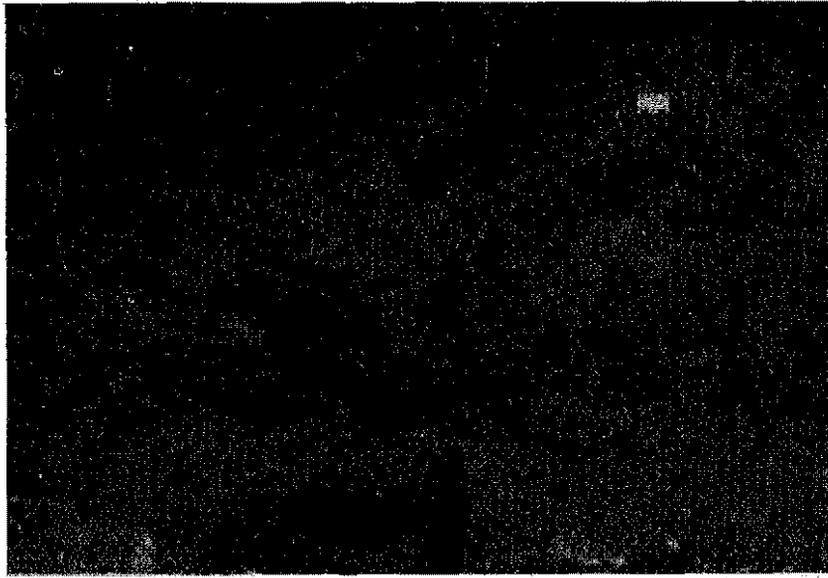


FIGURE 14.--Contact between intensely deformed, fine-sand laminae and diamict in graben (see location, Figure 9). The two units interpenetrate along parallel thrust faults.

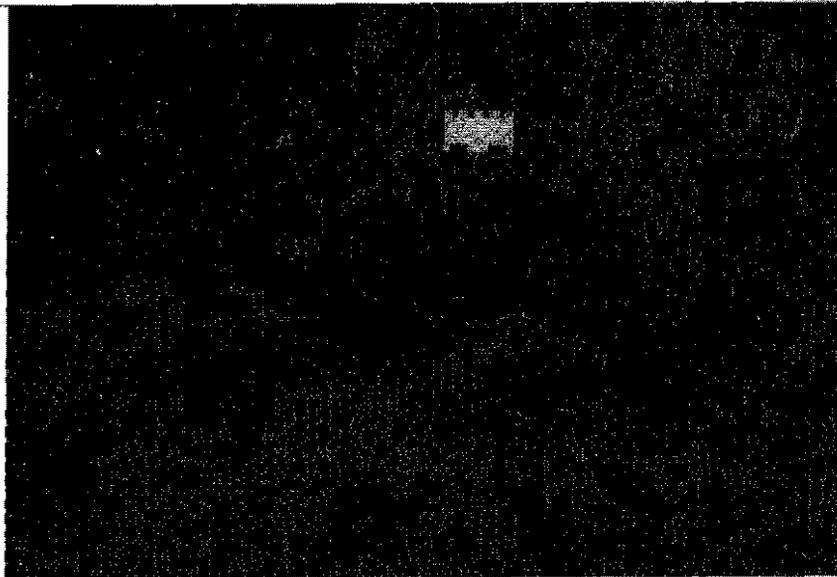


FIGURE 15.--Detail of contact in Figure 13 (see location, Figure 9).

Two episodes of deformation are discernible in this cut: one that affected all the laminae in the surface zone of the lake bed and one that affected just the sediments around the pebbly clay mass. Throughout the surface zone, laminae are cut by small-scale normal and reverse faults with diverse orientations, but with material transport into local depressions. This redistribution of sediments apparently was caused by horst and graben faulting, which most likely resulted from differential compaction of lake-bottom sediments overlying an irregular surface. In the active clay pit, beds are broadly warped and subvertically faulted at a depth of 20 ft. The numerous small swamps on the surface may indicate local subsidence. It thus appears that the Berlin clay at Middletown was affected at least locally by differential compaction. An alternative or additional cause of local deformation may have been mining activities: vibration, excavation, loading in proximity, and lowering of the water table all are known to cause the lateral yield of saturated clay toward the face of an excavation, as the bottom of the face heaves toward the cut, and the overlying clay settles (Terzhagi and Peck, 1948, p. 515-518; p. 527). The dumping of spoil on the surface (as in the bog at the west end of this cut) may have caused the peripheral deposits to bulge "isostatically."

The pebbly clay appears to have been embedded in the lake-bottom surface by movement from northwest to southeast. It may be a pod of iceberg till (Ovenshine, 1970) that slid into a local depression; or it may be a redeposited mass of lake-bottom sediments (perhaps dislodged by a grounding iceberg) that likewise slid downslope.

STOP 8. New Pond pit, Shadow Ln., Town of Cromwell, Middletown quadrangle (NE 9th). Shadow Ln. is on the E. side of Conn. Rte. 99, 1.75 mi N. of West St. (at the center of Cromwell). New Pond is approximately 0.5 mi down the lane. Road to pit crosses the dam at the S. end of New Pond.

This pit is located in the Rocky Hill dam deposits (par. 14, 15, 19, 20). The altitude of the surface here is 135 ft, and the surveyed altitude of the topset-foreset contact is 125 ft. The topset beds are composed of trough cross-bedded sand and pebbly sand, approximately 10 ft thick. They fine downward toward the foreset contact. The foreset beds are composed of chiefly sand and are 30 to 40 ft thick. The beds dip generally  $20^{\circ}$  toward S. $20^{\circ}$ W., which is also the orientation of the tongue-shaped body in which this pit is located. This landform is interpreted as a distributary lobe that prograded southwest into glacial Lake Middletown (par. 15). There is a sharp textural break between the foresets and the gently dipping bottomsets, which are composed chiefly of very fine sand and silt in planar laminations, massive beds, and climbing ripple sequences. These bottomset deposits are as much as 86 ft thick and fine downward to clay (Ryder and Weiss, 1971). Inasmuch as the bottomsets are relatively thick here and the topsets relatively fine-grained, this location is inferred to have been "distal" with respect to the source of this delta (ice line 3, fig. 2). By the time the delta front reached this location, the lake bottom surface had been raised from -21 ft (or lower) to +65 ft in altitude.

STOP 9. Riverdale pit, Conn. Rte. 66, Town of Portland, Middle Haddam quadrangle (CW-CC 9ths). Entrance to pit is S. of Rte. 66, 1.65 mi E. of junction with Conn. Rte. 17. Pit is owned by Joe Seiferman, proprietor of the Riverdale Motel (just W. of pit).

This pit is in glacial stratified deposits at the mouth of the buried former channel of the Connecticut River (fig. 1). The depth to bedrock here is more than 150 ft (Bingham and others, 1975), and these glacial stratified deposits may be more than 290 ft thick. The upper 25 to 30 ft are exposed in two cuts. The upper cut is oriented N.20°W. and exposes 15-18 ft of horizontally bedded, imbricate boulder-cobble gravel, fining downward to cobble-pebble gravel and sand. Clasts in the upper beds dip generally to the north-northwest. Some channels and bars are indicated in the lower gravel by troughs filled with graded cobble-pebble to pebble gravel and by wedges of sand in high-angle cross beds, overlain by plane-laminated sand. The lower cut, in the floor of the upper cut, exposes 10 ft of plane-bedded and trough cross-bedded sand on bedding planes that dip south-southeast. These appear to be delta foresets, and the topset-foreset contact, as measured in the floor of the upper cut, is 121 ft.

**STOP 10.** Hubert E. Butler Co. pit, Conn. Rte. 66, Town of Portland, Middle Haddam quadrangle (CW 9th). (Property owned by the Middletown YMCA and leased to the Butler Co.) Pit can be reached via the driveway to YMCA Camp Ingersoll Phelps, located on the N. side of Rte. 66, 0.45 mi E. of the junction with Conn. Rte. 17. Driveway is opposite the E. end of the median on Rte. 66. Permission to enter must be obtained from either the Butler Co. or the "Y." This pit is dangerous; stay away from the walls.

This pit is located in kettled melt-water deposits that fill the preglacial channel of the Connecticut River (fig. 1). Along the thalweg of this channel, deposits are more than 276 ft thick (Bingham and others, 1975). They comprise numerous small, ice-contact deltas graded to irregular lakes that formed between the ice margin, other deltas, and masses of stagnant ice. At the site of this pit, a delta prograded west-northwest into a lake at the north end of Jobs Pond (fig. 2), which was the site of a huge ice block. Stream flow was deflected west by this ice block and by the slightly older delta to the south. This is indicated by features seen in the pit: collapsed beds in the north and south flanks of this landform, topset gravels that appear to fine from east to west across the pit, and foreset beds that appear to dip west-northwest. The topset-foreset contact stands 10 to 12 ft below the land surface, which reaches 175 to 185 ft. The altitude of the contact is thus between 163 and 175 ft. This is at least 18 ft higher than in subsequent deltas (to the north) that graded to glacial Lake Middletown (par. 10). The elevated contact in this pit indicates that when this delta was deposited, melt water was forced to flow south across deposits between Jobs Pond and Riverdale, which have surface altitudes of 145 to 175 ft.

The topset beds in this pit are composed of boulder-cobble to pebble gravel and pebbly sand. Textures fine downward and bedding correspondingly changes from crudely horizontal at the top to cross-stratified below, in bar and channel deposits. The foreset beds are composed of pebble gravel and pebbly sand at the contact, fining downward to sand and fine pebble gravel. In a separate, 10-ft cut in the floor of this pit, the foresets have composite bedding, where trough and tabular subsets migrated down the foreset slope.

**STOP 11.** Sage Hollow Rd. pit, Town of Portland, Middle Haddam quadrangle (NE 9th). Sage Hollow Rd. runs E. from the junction of Conn. Rtes. 17 and 17A at Fogelmarks Corners. Gravel road to pit is 0.13 mi E. of this junction, on the S. side of Sage Hollow Rd. Pit is owned by the Town of Portland.

This pit exposes uncollapsed deltaic bedding in the flanks of a "kame terrace" on the east wall of the Connecticut River channel. The upper cut exposes approximately 11 ft of interbedded pebble gravel (in crudely horizontal beds and cross beds) and sand (in tabular cosets with high- and low-angle cross beds). The lower cut exposes at least 20 ft of plane-bedded sand in foresets that dip generally  $35^{\circ}$  west-southwest. The estimated altitude of the topset-foreset contact is 144 ft. The narrow terraces in Portland connect deposits of the Rocky Hill dam in South Glastonbury with deposits in the buried channel at Portland. They previously were believed to be fluvial kame terraces deposited alongside a mass of stagnant ice that filled the Connecticut River channel between Wangunk Meadows, Gildersleeve, and North Cromwell (Flint, 1953; Deane, 1953; see locations on base map, fig. 6). The absence of similar, connecting deposits across the river at Middletown prompted the inference that the channel at Middletown was completely blocked by this ice mass, which forced melt water to flow through the ancestral channel at Portland.

This pit shows that the terrace flanks are locally non-ice-contact and that they formed in places as free delta fronts. Each terrace may comprise several wedge-shaped deltas that prograded southwest along the ice margin into open water. In the case of this particular delta, a northeast-southwest ice margin is indicated by the trend of the terrace edge north of the pit and by cuts in this edge that show surface textures coarsening toward the northeast.

Deltaic bedding has been exposed throughout the narrow kame terraces in Portland. Adjusted for rebound (par. 2), the altitudes of topset-forest contacts show that these deltas graded to progressively lower water planes during their sequential emplacement (fig. 4). For this reason, they are interpreted as the deposits of one gradually lowering lake, glacial Lake Middletown (par. 10, 12, 19, 20).

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The sedimentology, stratigraphy, and paleontology of  
the Lower Jurassic Portland Formation, Hartford Basin,  
central Connecticut

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"It is too generally a habit to think that while  
extraordinary examples of geological structure  
may be found in remote parts of the world they  
are not to be looked for at home, where  
familiarity with the hills and valleys brings  
them to be regarded as commonplace."

- W.M. Davis, 1898, p. 25.

PART I.

Geologic Setting

In the Middle and Late Triassic, regional extensional stress associated with the incipient rifting of the North American and African plates created a series of half-graben and graben (the Newark Supergroup) along the eastern margin of the Appalachian orogen from Nova Scotia to South Carolina (Froelich and Olsen, 1984). The Hartford Basin, in central Connecticut and south-central Massachusetts, is approximately 140 km long and up to 30 km wide (Fig. 1). Basin subsidence began in mid-Norian and was primarily controlled by large, west-dipping, north-south-trending listric normal faults developed on the flanks of mantled gneiss domes of Devonian age (Olsen and others, 1982; Wise and Robinson, 1982). The greatest subsidence occurred on master faults located along the eastern edge of the basin. The western margin of the basin is delineated by minor normal faults, monoclinial flexures, and angular unconformities. Deposition in the resulting asymmetric trough was penecontemporaneous with subsidence, and for at least 25 million years the basin was filled with sediment derived primarily from the metamorphic and igneous terranes of the highlands to the east. Due to post-Jurassic uplift and erosion, the original extent and volume of the deposits in the Hartford Basin cannot be determined. As much as 7.5 km of Newark strata may remain in the north-central part of the basin (Burger and Ataman, 1984).

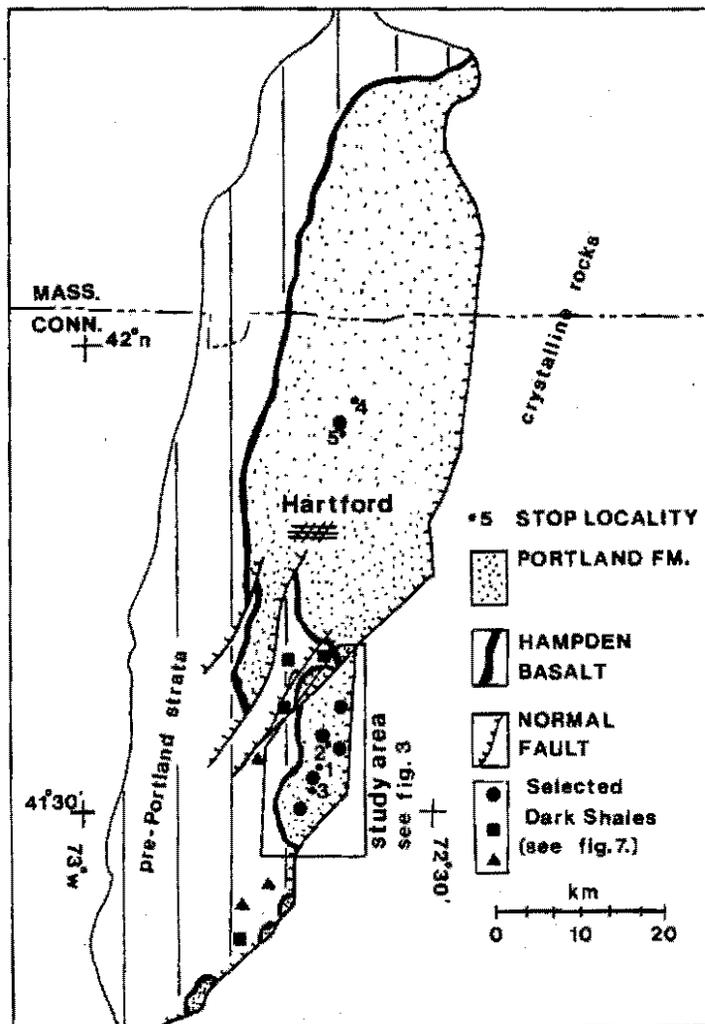


Figure 1. A simplified geologic map of the Hartford Basin.

Underlying magma sources were tapped intermittently during the infilling of the basin. Recently identified feeder dikes may have been the conduits through which tholeiitic lavas were extruded on to the basin floor during three separate intervals in the earliest Jurassic (Martello and others, 1984; Philpotts, 1985). The stratigraphy of the Hartford Basin consists of four sedimentary formations delineated by three interbedded, laterally-continuous basalt flow units (Table 1). This relatively simple stratigraphy is locally complicated by block faults, folds, and a lack of distinctive marker beds in the sedimentary rocks. All units in the Hartford Basin dip generally to the east and progressively younger rocks are exposed in a traverse from west to east across the basin.

<u>Stratigraphic Divisions</u>	<u>Estimated Thickness</u>
Portland Formation	450 - >2000m
Hampden Basalt	60 - 100m
East Berlin Formation	170m
Holyoke Basalt	100 - 150m
Shuttle Meadow Formation	100m
J Talcott Basalt	65m
<hr/>	
Tr New Haven Formation	>2200m

Table 1.

Data from Cornet (1977); Hubert and others (1978), Olsen and others (1982).

The sedimentary rocks of the basin have been dated and correlated with the standard European stages by palynofloral comparisons supplemented by megafossil plant and vertebrate evidence (Cornet and others, 1973; Cornet and Traverse, 1975; Cornet, 1977; Olsen and others, 1982). Deposition of Portland sediments began in the late Hettangian and continued through the Toarcian, a span of at least 15 million years (Cornet and Traverse, 1975; Palmer, 1983). The Portland Formation contains the youngest rocks in the Hartford Basin and the uppermost Portland strata are the youngest of all Newark Supergroup deposits (Olsen and others, 1982).

The Portland Formation is exposed in a belt that widens from 1/3 to fully 2/3 of the width of the basin from central Connecticut north to the Holyoke Range in Massachusetts; the Formation is largely faulted out in southern Connecticut. The Portland varies in thickness and ranges from as little as 450 m in southern Connecticut to perhaps as much as 4000 m in southern Massachusetts (Cornet, 1977).

The Portland Formation was named by Krynine (1950) for the spectacular exposures of medium to coarse sandstone in the old quarries at Portland, Connecticut, but these red-brown, planar- and cross-stratified sandstones comprise no more than half of the Formation as a whole. Extreme lateral and vertical variations are characteristic of the Portland and the rocks become progressively finer-grained from east to west across the basin. The coarsest rocks, including polymictic boulder and cobble conglomerate and pebbly sandstone, are found along the eastern margin of the basin. As much as 30% of the Formation consists of red, red-brown, grey, and black siltstone and shale (Krynine, 1950), primarily in the central and western portions of the

outcrop area. Asymmetrical and symmetrical dark shale cycles, similar to those described from the East Berlin Formation (Hubert and others, 1976) are common in the lower half of the Portland. These dark shale units are of paleontological importance and may prove to be of economic value as hydrocarbon source beds.

During the long interval of Portland sedimentation, southern New England was situated in the sub-tropics, 15°- 20° north of the paleoequator (Olsen, 1984). The lateral and vertical distribution and variable character of the rocks and fossils of the Portland Formation suggest that the Early Jurassic climate was seasonal and characterized by varied-length oscillations between humid and semi-arid conditions (Hubert and others, 1982; Thomson, 1983). Perennial and ephemeral streams flowed generally westward across the eastern and central parts of the basin. Along the eastern margin, alluvial fan complexes were deposited against active fault escarpments. The basin floor contained braided and meandering stream channels and broad, fine-grained floodplains. During dry intervals, the basin floor was occupied by ephemeral lakes and playas. However, at times of increased rainfall, large, perennial, stratified lakes became established in the basin and encroached on the eastern margin to form transgressive depositional sequences over the distal portions of alluvial fans. Several lacustrine shoreline deposits have been recognized within 0.75 km of the basin margin in central Connecticut.

#### Facies and Facies Distribution

The sedimentary rocks can be divided into four general facies and eight numbered sub-facies that define distinct depositional sequences. The facies divisions are based primarily on grain size, e.g., maximum clast size in the conglomerate units, or sandstone/siltstone ratios in the fine-grained units. Diagnostic fossils also assist in the recognition of facies, particularly in the fine-grained units. Facies are mappable units at a 1:24,000 scale. Sub-facies divisions are based on sedimentary structures, grain size, sorting, color, nature of bedding and stratification, and other characteristics. A summary table of facies and sub-facies defined within the Portland Formation follows. The colors of the rocks are various shades of red, red-brown, or brown unless otherwise specified.

FACIES	SUB-FACIES	DESCRIPTION	DIAGNOSTIC FOSSILS	DEPOSITIONAL ENVIRONMENT		
Conglomerate	1	matrix-supported, poorly-sorted boulder and cobble conglomerate	N/A	debris flow	Alluvial Fan	upper to mid fan
	2	poorly-stratified, clast-supported conglomerate and pebbly sandstone	N/A	shallow braided stream (ephemeral)		mid to lower fan
	3	cross-stratified conglomerate and pebbly sandstone	scarce reptile bones	braided stream (possibly perennial)		distal fan
Sandstone	4	planar-laminated and ripple cross-laminated silty fine sandstone	scarce invertebrate burrows	sheetflow	Basin Floor	along the eastern half of the basin
	5	planar- and cross-stratified medium to coarse sandstone with thin, interbedded siltstone	invertebrate burrows, plants, reptile tracks, scarce reptile bones	braided and meandering stream and floodplain		primarily in the central portions of the basin
Siltstone	6	thin-bedded siltstone with mudcracks and medium to coarse sandstone lenses and layers	reptile tracks, invertebrate trace fossils, ostracodes, conchostracans, plants	floodplain with shallow stream and crevasse-splay sandstone	Perennial Lake	lake margin and shoreline
Dark Shale	7	grey, well-sorted, ripple cross-laminated sandstone and siltstone	abundant plants, unionid mollusks, reptile tracks, conchostracans, ostracodes, pollen/spores, invertebrate burrows	shallow water at or near wave base and shoreline		deep lake water and lake bottom
	8	dark grey to black siltstone and finely-laminated shale	abundant fishes, coprolites, conchostracans, abundant plants	below wave base in quiet, oxygen-poor water		

Table 2. Summary of Portland Formation facies and inferred depositional environments.

The distribution of facies in the Portland Formation is shown in Figure 2. All units dip to the east and consequently both the lateral and vertical distribution of the facies is illustrated. The Conglomerate Facies of the Portland Formation is limited to the eastern margin of the basin. The Sandstone Facies of the Portland is generally restricted to the eastern half of the outcrop area, but also occurs in a narrow, north-south-oriented belt overlying the Hampden Basalt, particularly in northern Connecticut. The Siltstone Facies

occupies a broad belt along the central axis of the basin and limited areas along the eastern margin. The Dark Shale Facies is intercalated with the coarser Sandstone and Conglomerate Facies in central Connecticut, but in northern Connecticut and Massachusetts, the dark shale units are almost exclusively associated with the finer Sandstone and Siltstone Facies.

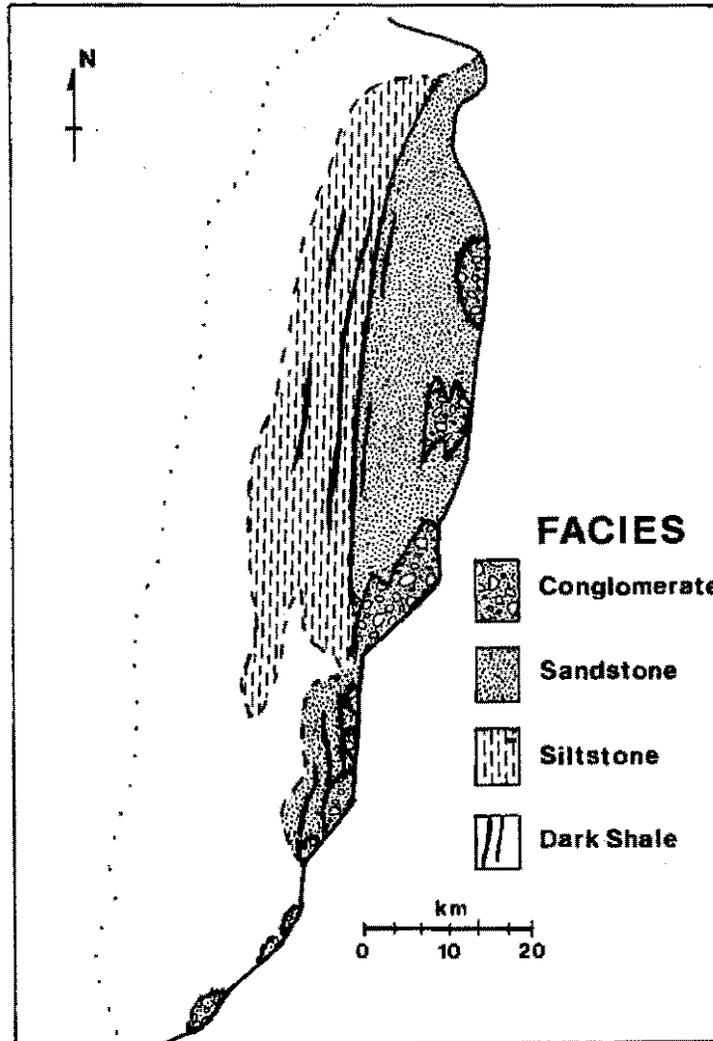


Figure 2. Distribution of Portland facies in the Hartford Basin. (Dark Shale not to scale and locations approximate).

Alluvial Fan Deposits: The Basin Margin

"...on the very margin of the valley, we find a coarse conglomerate in a few places, of quite peculiar character... The fragments are sometimes several feet in diameter, and the stratification of the rock is very obscure."

- E. Hitchcock, 1858, p. 11.

The best exposures of the coarse strata of the Conglomerate Facies are located along the basin margin in Durham, Middletown, and Portland, Connecticut. Over 700 m of measured section was compiled at Round Hill (Middletown) and Crow Hill (Portland) to understand the stratigraphic succession and depositional history of the coarse-grained units (Fig. 3). Preliminary results of investigations in this detailed study area have been previously discussed in LeTourneau and Horne (1984), LeTourneau and Smoot (1985), and LeTourneau (1985a, 1985b, 1985c).

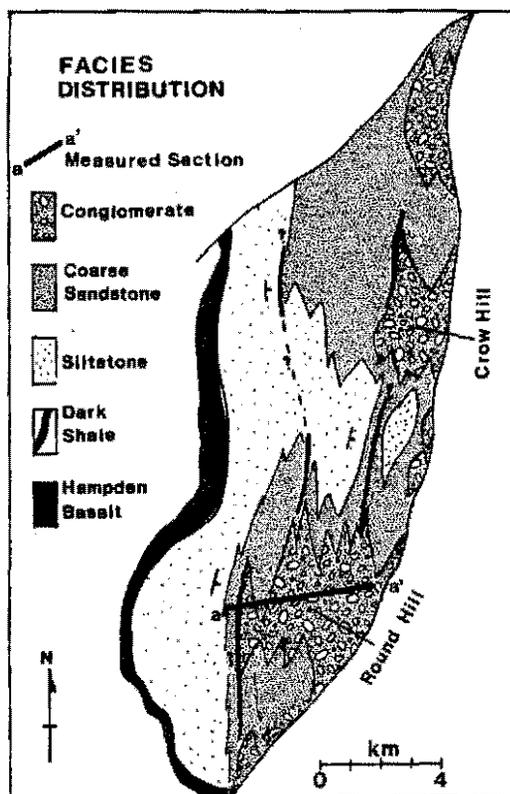


Figure 3. Facies distribution in the detailed study area, central Connecticut (see Fig. 1).

The texture, fabric, sedimentary structures, clast-size distribution, bedding geometry, and sediment dispersal patterns of the rocks of sub-facies 1 - 4 are characteristic of alluvial fan deposits. These coarse-grained sedimentary rocks form wedge- or prism-shaped bodies that thin and fine away from the basin margin (Fig. 3).

A map of the maximum clast-size distribution (Fig. 4a) demonstrates that the conglomerate grain size decreases rapidly over a distance of 2 - 3 km from the basin edge and that the coarsest units outline discrete lobate areas. The paleocurrent patterns at Round Hill and Crow Hill (Fig. 4b) radiate out from the central parts of the conglomerate lobes. The sediment dispersal patterns and the composition of the conglomerate clasts indicate that the source area for the alluvial fan sediments was in the Paleozoic crystallines east of the present basin margin.

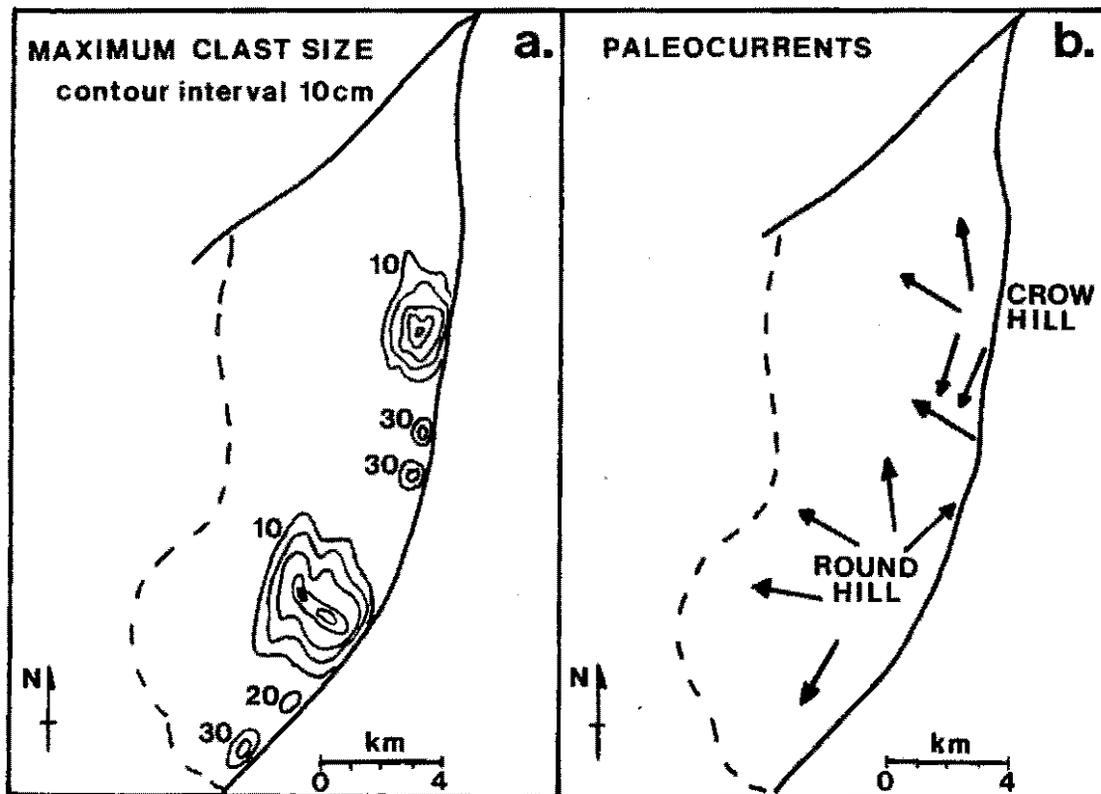


Figure 4. (a.) Maximum clast size distribution and (b.) paleocurrent patterns in the detailed study area.

Deposition on alluvial fans is by stream flow or debris flow processes which represent end members of a continuum of increasing flow viscosities beginning with clear water. Fluvial discharge across an alluvial fan forms shallow, temporary, braided streams that become more disperse in a down-fan direction. These braided streams shift back and forth on the fan surface as sediment accumulates in the stream bed. Very broad, shallow, essentially unconfined stream flows in the lower reaches of an alluvial fan are termed sheetflows.

Debris flows are highly viscous slurries of unsorted coarse sediment supported in a fine-grained matrix. Because of the high matrix strength and the non-turbulent internal flow, very large boulders and cobbles are commonly rafted on the surface or carried within the debris flow. Debris flows are the major process which transports large clasts to the mid-fan and lower-fan areas.

The poorly-sorted boulder and cobble beds of Sub-facies 1 compare most favorably with descriptions of modern and ancient debris flow deposits (Nilsen, 1982; Wasson, 1977). Diagnostic features include: random or chaotic orientation of clasts in a mud-rich matrix, concentration of the largest clasts near the upper and outer contacts of the deposit, hummocky and irregular upper contacts, planar and distinct lower contacts, and an absence of cross-bedding or stratification within the deposit. These debris flow units form distinct layers or lenses surrounded by pebbly sandstone and conglomerate of Sub-facies 2.

Sub-facies 2 consists of thin bedded, poorly-stratified conglomerate and pebbly sandstone in normal-graded and laterally-discontinuous lenses. These beds are interpreted as ephemeral braided stream deposits from the mid-fan area. The predominately fining-up depositional units suggest decelerating flow events terminating with the deposition of fine sand and silt during the waning stages of flow. Most of the fine drapes forming the upper contacts of individual units have dessication features (mudcracks and mudcurls), which strongly implies that flow events were punctuated by periods of non-deposition and subaerial exposure. Cross-bedding is generally rare in this sub-facies. Horizontal- and inclined-planar stratification is the most common internal structure in these units. The fine-grained upper contacts are planar-laminated to ripple cross-laminated.

Sub-facies 3 is characterized by trough and planar cross-stratified conglomerate and pebbly sandstone. Cross-bed sets average 0.2 - 0.3 m thick and reach a maximum of 1.5 m. Imbricate pebbles define the lower contacts and the foresets of the conglomerate and pebbly sandstone units. Typically, the upper and lower contacts are discontinuous and the lens-like, cross-stratified units interfinger within larger meter-scale beds. Pebbles and cobbles are dispersed in the coarse sand and

granule matrix or rarely occur in discrete lenses or layers.

The conglomerate and sandstone units of Sub-facies 3 are interpreted as deposits of moderately deep, possibly perennial, braided streams. This interpretation is based on the occurrence of sedimentary features indicative of persistent stream flow: abundant, well-developed, large-scale cross-stratification, moderate sorting, the absence of dessication features, and a fairly uniform style of stratification and bedding. The lack of large boulders (such as those found in Sub-facies 1 and 2) suggests that the streams were not competent enough to transport very large clasts or that debris flows were not associated with this sub-facies. When traced in a basinward direction, this sub-facies is laterally adjacent to, and in part correlative with, the Dark Shale Facies.

Sub-facies 4 is composed of thin, planar-laminated and ripple cross-laminated, poorly-sorted, micaceous, silty fine sandstone. This lithology characteristically contains a high percentage of carbonate cement and pore fillings. Granule interbeds are common.

Sub-facies 4 includes the finest-grained alluvial fan sediments which are, in part, transitional between the lower fan and basin floor depositional environments. These units can be traced for hundreds of meters or more and provide easily-identified, laterally-correlative horizons within the coarse sandstone and conglomerate units. Laterally, this sub-facies thins and pinches out against coarser units. In a basinward direction, it thickens and becomes gradational with thin-bedded siltstone.

The rocks of Sub-facies 4 are sheetflow deposits formed by shallow, unconfined flow on sandflats at the distal portions of alluvial fans. This unit is typically found downsection from the rocks of Sub-facies 1 and 2. This relationship is repeated through the vertical section at Round Hill; the predominately coarsening-up depositional sequences delineate successive progradational alluvial fan cycles. The fine-grained rocks of Sub-facies 4 provide convenient marker units that define the base of individual cycles.

Figure 5 is a paleoenvironmental reconstruction of the detailed study area in central Connecticut showing the location of alluvial fans and a perennial lake (see discussion below) in the Early Jurassic. Alluvial fan locations are based on the geometry and distribution of the coarse-grained units (Fig. 3), particularly Sub-facies 1 - 4, discussed above. The maximum clast-size distribution (Fig. 4a) and the radial paleocurrent patterns at Crow Hill and Round Hill (Fig. 4b) define discrete alluvial fans along the eastern margin of the basin. The location of the hypothesized lake is based on several well-exposed lacustrine cycles that are interbedded with the alluvial fan deposits, for example, at Laurel Brook in Middletown (Stop 3).

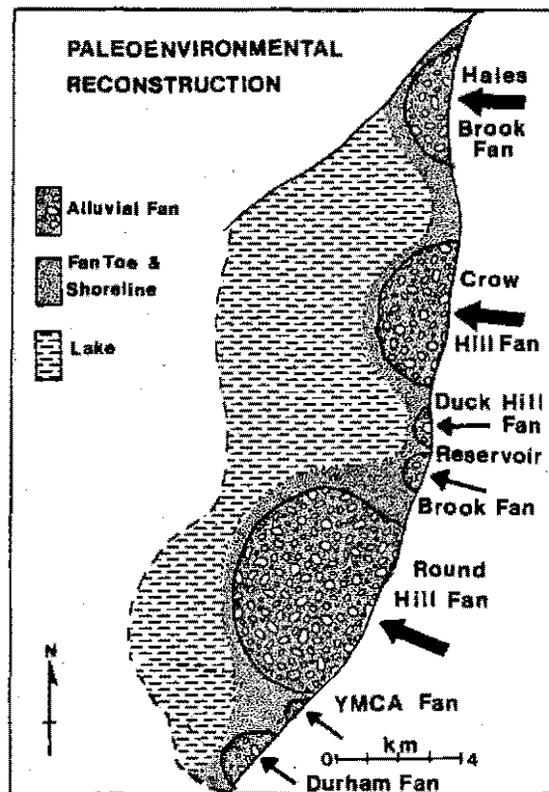


Figure 5. Paleoenvironmental reconstruction of the detailed study area during a period of maximum lake expansion.

#### Braided and Meandering Streams and Floodplains: The Basin Floor

The medium to coarse sandstone, siltstone, and minor shale of Sub-facies 5 and 6 dominate the sedimentary section along the central axis of the Hartford Basin and comprise well over half of the Portland Formation as a whole. Sub-facies 5 is primarily found east of the central axis of the basin. Sub-facies 6 is generally confined to the western side of the basin in central and northern Connecticut and southern Massachusetts (see Figure 2).

Sub-facies 5 consists of cross-stratified, coarse and pebbly sandstone with intercalated, laminated and thin-bedded, burrowed and mudcracked siltstone. The coarse sandstone forms the

thickest units, ranging from 0.2 to 1.0 m thick. Bedding consists of horizontal- and inclined-planar stratification and tabular and trough cross-stratification. Pebbles comprise less than 10% of the coarse fraction and rarely exceed 6 cm in diameter. The lower contacts are scoured into the underlying siltstone or sandstone and contain abundant mud rip-up clasts. The upper contacts are planar to undulatory and are delineated by sharp changes in grain size.

In Sub-facies 5, the coarse sandstone layers or lenticular beds are separated by micaceous, planar-laminated and ripple cross-laminated maroon, red, or red-brown siltstone. The siltstone beds range from 0.1 to 0.5 m thick and may be laterally continuous or pinch out over several meters. Distinctive characteristics of the siltstone units are the ubiquitous dessication cracks and abundant trace fossils, including invertebrate burrows and reptile tracks.

Sub-facies 6 is closely related to Sub-facies 5; these two divisions represent end members of varied sandstone/siltstone ratios for the fluvial basin floor deposits. Sub-facies 6 is composed largely of massive to thin-bedded red and brown siltstone with interbedded thin lenses and layers of medium to coarse sandstone. Internal structures in the siltstone include horizontal and wavy lamination, ripple and climbing-ripple cross-lamination, and minor oscillatory ripple cross-lamination. Mudcracks and soft sediment deformation features, including load casts and pseudonodules, are abundant. The suite of trace fossils found in the siltstone is similar to that described from the fine-grained units of Sub-facies 5. Orange-brown dolomitic concretions up to 15 cm in diameter are very common in some of the laminated siltstone units, particularly in exposures along the central axis of the basin. Evaporite mineral casts (halite and gypsum) have also been identified in this sub-facies.

The thin sandstone lenses and layers average 0.2 m thick and typically have scoured and irregular lower contacts and planar to undulatory upper contacts. Planar and trough cross-stratification is well developed and mud rip-up clasts are very abundant along the lower contacts and cross-bed foresets.

Sub-facies 5 and 6 formed in braided and meandering stream and floodplain depositional environments in the central portions of the Hartford Basin. The location of the sand-dominated units of Sub-facies 5 may indicate the relative positions of the stream channel networks at various stratigraphic levels. Channels became incised into the finer-grained deposits and as the streams scoured the substrate and migrated laterally across the surrounding broad floodplains, the cohesive silt and clay was incorporated into the cross-stratified stream channel sands as mud rip-up clasts. At the K-F Quarry (Stop 4), an illustration of the meandering nature of the broad, shallow stream channels in the central part of the basin is provided by abundant, well-exposed lateral accretion barforms or point bars that define

the edges of the 0.5 to 1.5 m thick sandstone lenses. Along the eastern side of the basin, trough cross-bedded, coarse and pebbly sandstone units are the typical braided stream deposits.

The siltstone units of Sub-facies 5 and 6 are fine-grained floodplain deposits. Sedimentary features in the siltstone are indicative of periods of rapid sedimentation alternating with prolonged episodes of subaerial exposure and dessication. When streams overtopped the shallow channels, sand and silt were deposited on the broad floodplains in climbing ripples and as planar-stratified sheets, and bioturbating organisms worked through the fine-grained, water-saturated substrate. Some of the thin sandstone lenses and sheets may be crevasse-splay deposits formed adjacent to the main stream channel networks during high discharge conditions. Shallow floodplain lakes and oxbow lakes formed as a result of channel meandering, by periodic inundation of the floodplains, and because of the shallow water table in the low-gradient basin floor.

During drier conditions, stream-flow was confined to the sand-filled channels and the floodplains developed extensive dessication surfaces. Standing water in shallow pools and ponds formed playas, as evidenced by the extensively mudcracked and mechanically-disrupted mudstone and siltstone fabrics. The evaporative concentration of surface and ground waters produced carbonate (including ferroan dolomite) nodules and irregular layers and, less frequently, evaporite minerals, including halite (Parnell, 1983) and gypsum (Krynine, 1950).

#### Lacustrine Deposits

"Now it is in the shales and sandstones...that we find organic remains - the fishes, the tracks, and the plants."

- E. Hitchcock, 1858, p. 11.

The sedimentary structures, sorting, bedding thickness and geometry, and fossils of the grey sandstone, grey siltstone, and black shale of the Dark Shale Facies are analogous to features of both modern and ancient lacustrine deposits (Link and Osborne, 1978; Hardie and others, 1978).

The grain size of the siltstone, sandstone, and minor conglomerate of Sub-facies 7 correlates with the offshore - onshore relationships in an inferred lake margin depositional environment. Grey, ripple-laminated and wavy-bedded siltstone units were deposited at or below wave base in the sub-littoral zone. The intercalation of laminated silt and clay and thin, fine sand layers is indicative of the variable energy conditions that existed between the nearshore and deeper water environments.

The grey sandstone units contain a wide variety of wave-generated sedimentary structures. Oscillatory ripples are the most common sedimentary structure observed in this unit; some of

the ripples have sheared crests suggesting that they were planed by turbulence from shoaling waves. Planar-stratified and tabular cross-stratified low-angle accretionary sandstone lenses were deposited as nearshore bars in the littoral zone and as linear beach ridges in the swash zone. These sandstone units are very well sorted, indicating a high degree of reworking and winnowing by waves. Fragmental plant material is very abundant and reflects the proximity of heavily-vegetated areas to the lake margin. Beds containing plant fossil layers suggest that detrital plant material may have formed floating and/or submerged masses near shore.

The sandstone beds become coarser and more conglomeratic onshore. Often, large boulders or cobbles are found as isolated clasts in the wave-rippled, well-sorted sandstone units. Large-scale, tabular cross-stratification in the coarse sandstone and minor conglomerate was formed in beach ridges built by wave reworking of alluvial fan sediment (LeTourneau and Smoot, 1985).

Sub-facies 8 is composed of thin-bedded, laminated and microlaminated black shale deposited under reducing conditions in periodically or permanently stratified lake waters (Sanders, 1968; McDonald, 1982). The fine laminations in the black shale indicate that bioturbating organisms were excluded from the substrate. Anoxic bottom conditions probably resulted from the thermal or chemical stratification of the lake waters. Many of the black shale units of the Portland Formation can be classified as oil shales because of the high percentage of kerogenous organic matter (Robbins and others, 1979). Abundant fossil fishes (semionotids and redfieldiids) characterize several of the black shale deposits; the excellent preservation of these fossils confirms the absence of scavenging benthos.

The finely-laminated nature of the black shale indicates deposition below wave base, but the size and depth of the lakes is uncertain. Anoxic or reducing conditions may form in shallow, eutrophic ponds or lakes, as well as in deep, stratified bodies of water. The thickness of the black shale units does not provide a measure of the depth of the ancient lakes of the Portland Formation, but only an indication of the temporal persistence of a particular depositional environment. In all probability, Portland lakes varied widely in size and depth, from ephemeral pools and playas to deep, stratified lakes hundreds of square kilometers in area.

Along the eastern edge of the basin, the lake and lake-margin deposits (Sub-facies 7 and 8) are intercalated with coarse-grained, stream-flow-dominated alluvial fan deposits (Sub-facies 3). Here, the thickest and best exposed transgressive and regressive shoreline sequences are located (Stop 3). The distribution of the Dark Shale Facies rocks (as shown in Figures 2 and 3) suggests that, at times, the ancient lakes drowned the distal margins of the alluvial fans (Fig. 5).

In contrast, the lake and lake-margin deposits identified in north-central Connecticut (Stop 5) are interbedded with

finer-grained, carbonate-rich clastic rocks and form fluvial-deltaic sequences in the central and western portions of the basin.

#### A Depositional Model for the Portland Formation

The combined influence of climate and tectonics on sedimentation in the Portland Formation can be interpreted from the lateral and vertical distribution of the sub-facies described above, the geometry of the various depositional units, and the sedimentary features of the different rock types.

#### The Climate Hypothesis

Two sub-facies assemblages have been recognized in central Connecticut. One sub-facies assemblage (Fig. 6a) is composed of Sub-facies 1, 2, 4, 5, and 6, which formed in alluvial fan and basin floor environments. Sedimentary features in this assemblage are evidence of ephemeral fluvial activity. The presence of debris flow deposits, abundant mudcracks and other dessication features may indicate that this assemblage was deposited during a "dry" phase of sedimentation, probably under arid to semi-arid climatic conditions.

The optimum conditions for the formation of debris flows are arid and semi-arid climates with less than 500 mm/yr of strongly seasonal rainfall (Blissenbach, 1954; Lustig, 1965). The preservation potential of debris flows on alluvial fans is highest in drier climates because low volume, low frequency stream flows often cannot rework the coarse-grained debris flow lobes on the fan surface. Because of these relationships, debris flows in ancient alluvial fans may be reliable paleoclimatic indicators.

Further evidence of "dry" depositional conditions in the Portland Formation is found in the fine-grained rocks of the central and western portions of the basin. Emerson (1917) and Parnell (1983) describe halite hopper-crystal casts in the Portland strata along the Connecticut and Chicopee Rivers in Massachusetts. Saline brines capable of precipitating halite and gypsum are easily formed by the evaporation of standing bodies of water in an arid or semi-arid environment. Gilchrist (1979) identified caliche paleosol horizons in the Portland Formation in north-central Connecticut. Caliche formation is dependent on evaporative concentration of alkaline ground water in the vadose zone; its occurrence implies semi-aridity and seasonal precipitation (Leeder, 1975; Hubert and others, 1978).

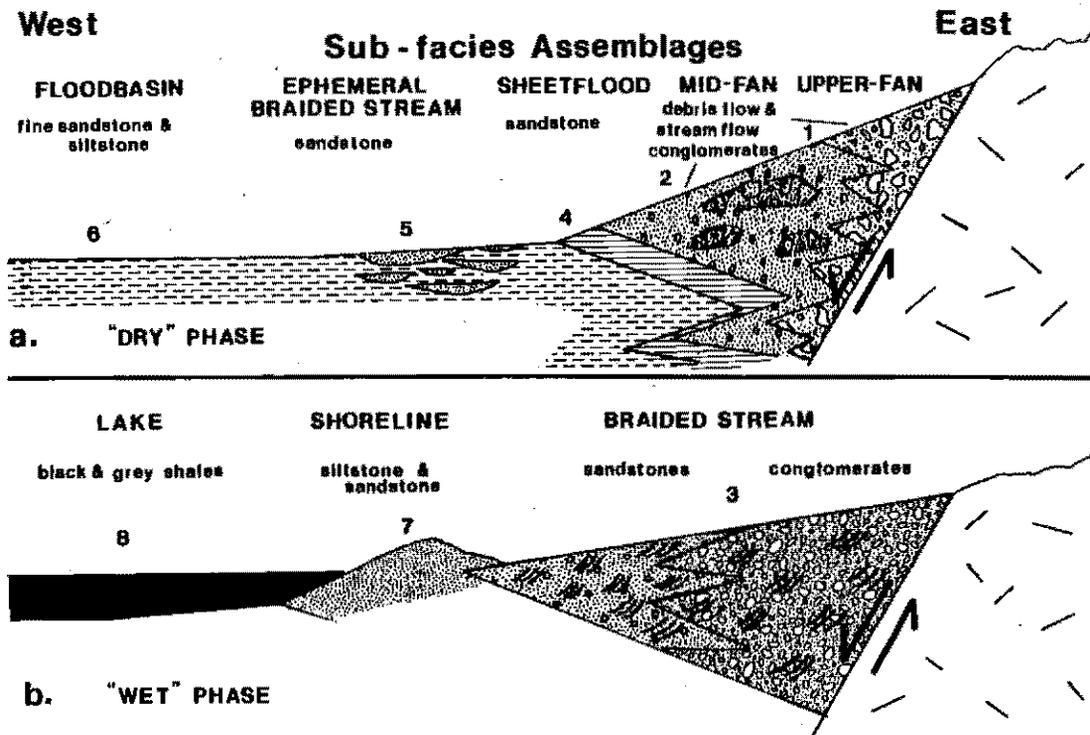


Figure 6. Sub-facies assemblages indicative of (a.) "dry" and (b.) "wet" phases of deposition.

The second assemblage (Fig. 6b) is composed of Sub-facies 3, 7 and 8, which represent: 3) proximal and distal braided stream deposits near the basin margin, 7) lake shoreline and littoral sediments, and 8) the profundal muds of a perennial lake. The laminated black shale (8) and grey sandstone (7) correlate laterally with the cross-bedded sandstone and conglomerate of Sub-facies 3. This assemblage is also found in vertical succession (for example, at Laurel Brook - Stop 3) in the following ascending sequence: 3-7-8-7-3, which represents a single transgressive - regressive lake cycle. This sub-facies assemblage may indicate a "wet" phase of sedimentation, the result of high fluvial discharge under humid climatic conditions. In the central and western portions of the basin, more humid climatic conditions are represented by an assemblage containing Sub-facies 5, 7, and 8.

Along the eastern border of the basin, alternating wet-dry-wet-dry-wet depositional cycles are documented through 700 m of measured section, beginning at Laurel Brook - "wet"

(Stop 3), and continuing through Goleman Road - "dry" (Stop 1) toward the eastern border fault.

The alternation of lacustrine strata with fluvial strata containing caliche, evaporite mineral casts, carbonate nodules, and dessication features provides convincing evidence for wet and dry depositional cycles during much of Portland time. Van Houten (1964) and Olsen (1984) described similar climatic cycles from the Triassic Lockatong Formation in the Newark Basin. Hubert and others (1982) and Gierlowski-Kordesch (1985) recognized cyclic depositional sequences in the Jurassic East Berlin Formation in central Connecticut.

#### The Tectonic Hypothesis

Several lines of evidence support the conclusion that syndepositional tectonic subsidence also controlled the style of sedimentation and the distribution of facies in the Portland Formation.

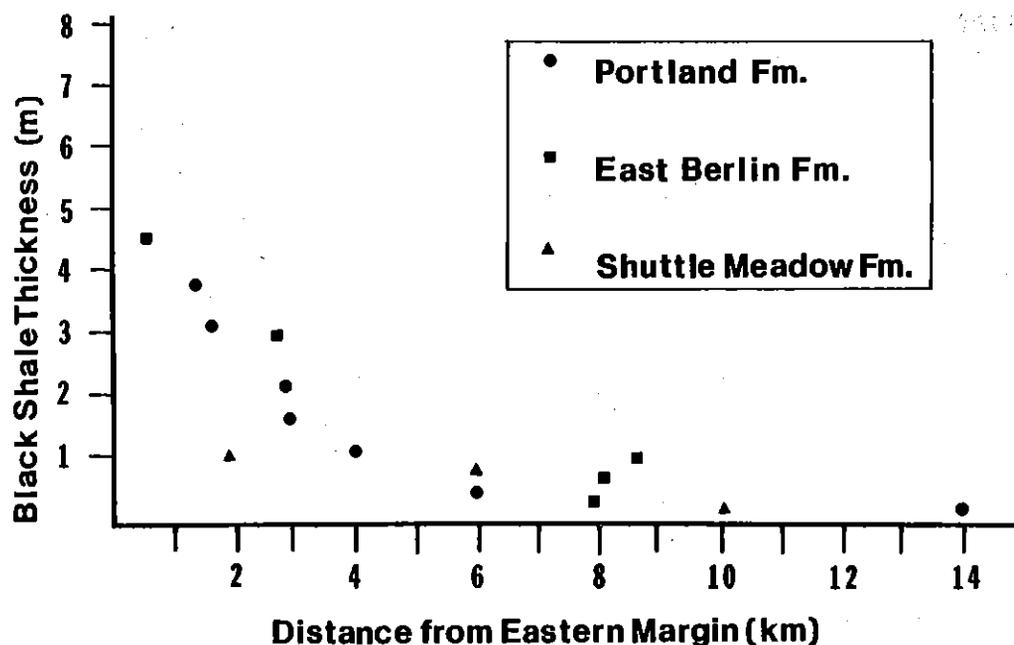


Figure 7. Thickness of black shale beds vs. distance from the eastern fault margin.

The grain-size distribution and the paleocurrent patterns of the eastern margin conglomerate and sandstone delineate discrete alluvial fans with radii of 1 to 3 km. Small, radial alluvial fans are typically associated with rapidly subsiding basin margins, for example, as along the east side of Death Valley, California. In contrast, the western side of Death Valley functions as a "hinge zone" for the asymmetric basin subsidence and the alluvial fans form deeply embayed, coalescing complexes (Hooke, 1972; Steel, 1976; Heward, 1978).

The syndepositional eastward tilting of the floor of the Hartford Basin is also suggested by the geometry of the lacustrine strata. Figure 7 graphs the thickness of the finely-laminated black shale units versus their distance from the present eastern border fault. In each of the formations containing lacustrine strata, the black shale units are thickest nearest to the eastern margin and thin towards the center of the basin. In addition, several lacustrine sequences located within 1 km of the basin margin contain features indicative of syndepositional seismic activity, including slump folds, ductile faults, and thin turbidites (Sanders, 1968; McDonald, 1975).

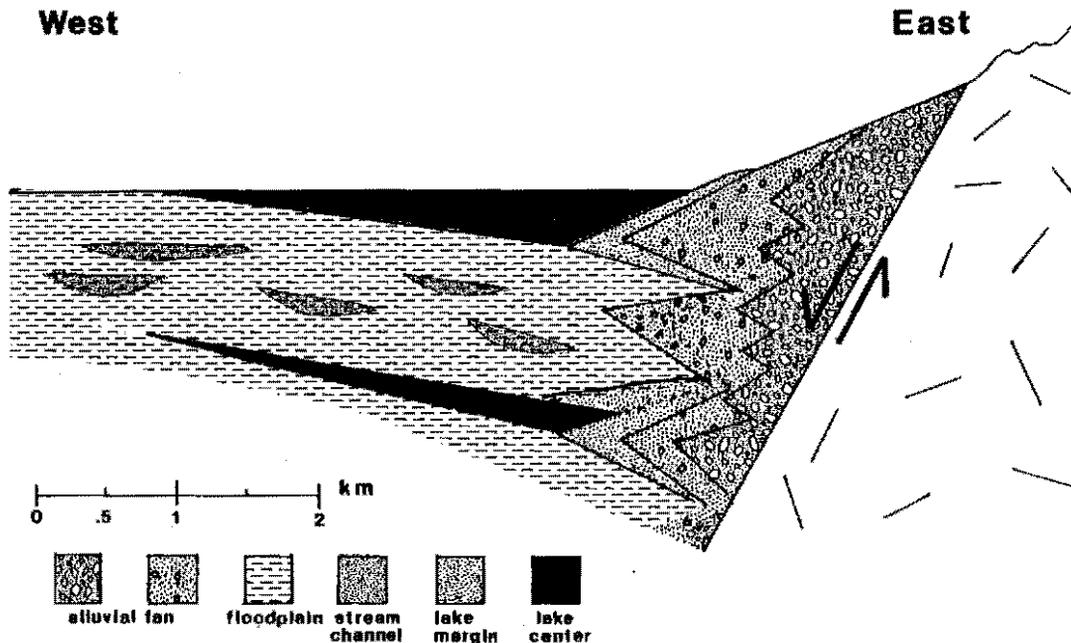


Figure 8. Inferred geometry of an Early Jurassic lake basin and deposits based on facies distribution and the data shown in Figure 7.

Figure 8 diagrams the geometry of a hypothesized Early Jurassic lake basin and its deposits. As a result of asymmetric basin subsidence, the perennial lakes of the Portland, East Berlin, and Shuttle Meadow Formations were deepest and persisted longest along the eastern margin. Shoreline features are well developed in the coarse alluvial fan deposits of the Portland Formation (Stop 3) (LeTourneau and Smoot, 1985).

Further evidence of the active tectonic subsidence of the Hartford Basin is provided by the provenance of basalt clasts in the conglomerate units at Round Hill (Fig. 3 and 5). The geochemical signature of the basalt clasts is comparable to that of the Hampden Basalt (Puffer and others, 1981) which underlies the Portland Formation (Table 1). One or more of the basalt flows in the basin may have crossed the fault margin and continued some distance into the eastern highlands. Subsequent uplift and erosion could then transport the basalt fragments on to the surface of the alluvial fans. At Round Hill, the basalt clasts are restricted to the lowest 250 m of the measured section (Fig. 3). This suggests that only a thin tongue of the Hampden Basalt encroached into the eastern highlands and a limited volume of material was available for transport into the basin.

Coarsening-up sequences of conglomerate and sandstone, 10 to 30 m thick, are the dominant depositional trend in the measured section through the Round Hill Fan (Fig. 3). These repetitive sequences reflect the progradation of alluvial fan lobes due to tectonic rejuvenation of the basin margin and the relative uplift of the source area (Heward, 1978; Gloppen and Steel, 1981).

#### Summary

The depositional model proposed for the Portland Formation is useful for predicting the distribution of lithofacies in the Lower Jurassic sedimentary rocks of the Hartford Basin. This model provides a framework for understanding a rift basin depositional system based on tectonic and climatic controls on sedimentation. Tectonic tilting of the basin floor is the primary control on sedimentation and influences the distribution of the depositional sub-environments in the basin. The vertical distribution of the lacustrine deposits provides a record of cyclic climatic change during the infilling of the basin.

Figure 9 summarizes the depositional system proposed for the Lower Jurassic rocks of the Portland Formation during a hypothetical period of maximum lake expansion.

The eastern parts were dominated by coarse clastic deposition and the greatest sedimentation rates. The easternmost black shales are the thickest and have the best preserved fossil fishes - a result of the well-developed stratification of the lake waters. Episodic influxes of sediment, due to the proximity of alluvial fans, may have contributed to the rapid burial and good preservation of the organisms.

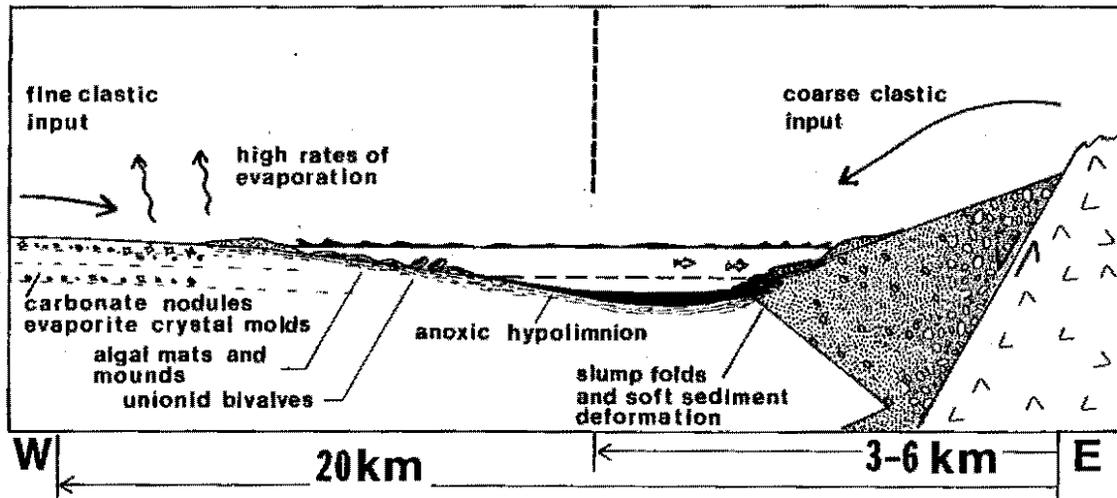


Figure 9. Summary of the depositional system of the Portland Formation during a period of maximum lake expansion.

In the western portions, deposition took place on a broad, low-gradient basin floor. The lake gradually shallowed to the west and fine-grained sedimentation was dominant. Carbonate nodules, limestone beds, algal stromatolites, and evaporite crystal casts were abundant along western shorelines, but rarely formed on the opposite shore. At Stony Brook (Stop 5), western shoreline sandstones of a perennial lake contain abundant micritic intraformational conglomerate, unionid mollusks, and plant debris. Low sedimentation rates encouraged carbonate production by algae, forming stromatolitic mats and mounds in the photic zone. The abundant carbonate nodules and less common evaporite mineral casts in the surrounding floodplain siltstone and shale formed from elevated rates of evaporation on the basin floor. Figure 10 is a reconstruction of Early Jurassic paleogeography based on the sedimentology and paleontology of the Portland Formation.



Figure 10. Paleogeographic reconstruction of the Hartford Basin during Portland time. View from west to east across the center of the basin. For descriptions of the organisms illustrated, see text (Stops 4 and 5) or McDonald (1982).

## PART II.

Stop Descriptions

AUTHORS' NOTE: Do not enter any of the localities listed below without obtaining the permission of the landowner(s). Please respect their property rights or access may be denied to future investigators.

Stop 1. Coleman Road, Middletown.

Exceptional exposures of alluvial fan conglomerate and sandstone are found throughout the Round Hill area in Middletown (Middletown 7.5 min. quadrangle). Several linear ridges are located 250 m east of Coleman Road, 1.5 km south of its intersection with Rt. 17. The vertical section at this stop incorporates coarsening-up, progradational alluvial fan cycles - from sheetflow sandflats of the fan toe (Sub-facies 4), to shallow stream deposits (Sub-facies 2) and debris flows (Sub-facies 1) of the mid-fan. These deposits accumulated during arid to semi-arid climatic conditions. The following discussion focuses on the lower or westernmost ridge in the outcrop area. This ridge extends in an essentially unbroken line over a distance of at least 2.5 km, from Round Hill Road in the south, to just west of Mapleshade Road in the north (Stop 2).

A complete range of wholly to partially preserved debris flow deposits occurs at this location. A "boulder bed" horizon is the most prominent feature in the 10 m high linear cliff. At the north end of the boulder bed, a number of diagnostic debris flow features can be observed, including: inverse grading, matrix-supported clasts in chaotic or random orientations, planar lower contacts, hummocky upper contacts, and abrupt grain-size contrasts with adjacent units. Grey-green, weathered basalt clasts are abundant and are possibly derived from the Hampden Basalt. The high percentage of mud in the debris flow matrix causes these layers to be eroded deeply into the cliff face.

Along the outcrop, approximately 50 m north of the boulder bed, several large boulders over 1 m in diameter are found. Some of these boulders shelter remnant debris flow fabrics beneath them; other large boulders occur as isolated clasts in a pebbly sand matrix. These boulders were originally freighted on to the alluvial fan by debris flows. Subsequent stream flow removed most or all of the finer matrix and smaller clasts and left the largest clasts behind.

The debris flows are surrounded by normal-graded, poorly-sorted, poorly-stratified conglomerate and pebbly sandstone (Sub-facies 2) deposited by braided streams.

Cross-bedding consists of small-scale trough cross-stratification representing "scour-and-fill" surfaces in the active channels. Horizontal-planar and inclined-planar stratification indicates that much of the sediment was deposited by shallow, high-velocity streams. Fine-grained, ripple- and wavy-laminated rocks of Sub-facies 4 are typically found at the base of each coarsening-up sequence.

The next ridge to the east (upsection) is similar; 10 - 30 m coarsening-up cycles repeat through the 700 m sequence at Round Hill. Covered intervals are underlain by fine-grained distal fan deposits that thicken in a basinward direction (toward Stop 2). Paleocurrent indicators at Stop 1 trend north - northwest revealing that this locality is on the northwest flank of the Round Hill Fan complex. The central area of the alluvial fan complex is located near Sunshine Farms and Round Hill Road, approximately 1 km south of Stop 1.

#### Stop 2. Mapleshade Road, Middletown.

Two well-exposed, parallel ridges of alluvial fan conglomerate and sandstone (Sub-facies 2 and 4) are located just northeast of Stop 1 (Coleman Road) at the sharp bend in Mapleshade Road (Middletown 7.5 min. quadrangle). These ridges are laterally correlative with the two ridges at Stop 1.

The ridges at this stop contain two coarsening-up, progradational alluvial fan cycles. At this site, the maximum clast size is smaller and the fining-up beds comprising the fan cycles are thinner than at Stop 1. Debris flow deposits have not been recognized here. Cobbles and boulders occur as clast-supported, imbricated lenses and layers at the base of the normal-graded beds.

A survey of the lower ridge shows that all of the cross-stratification is very small scale. Most of the outcrop is comprised of thin, fining-up depositional units, representing decelerating flow events (Sub-facies 2). In this ridge, a broad, shallow alluvial fan channel can be observed. This channel is less than 1 m thick and is approximately 5 m wide. Its lower contact is concave-upward and scoured into the underlying unit; the upper contact is planar and distinct. The base of the channel is defined by a clast-supported cobble or pebble lag. The channel fill is finer grained and better sorted than the surrounding beds. Large cobbles and boulders are absent in the channel. The channel is composed entirely of well-developed, small-scale trough cross-stratification. A comparison of the sorting and stratification within the channel lens and the surrounding units indicates the subtle contrast between more frequent, longer-duration stream flow in the shallow channel and short-duration, high velocity flow over broader areas of the fan during high discharge events.

The upper (eastern) ridge at this stop shows features similar to those in the lower ridge. A small channel is

exposed at the base of this outcrop. At the north end of the outcrop, the sheetflood sandstones of Sub-facies 4 are well exposed.

It is interesting to note that the covered interval between the two ridges at Stop 2 is thicker than that between the correlative ridges at Stop 1. This indicates that the distal fan deposits pinch out southward toward the central part of the fan complex. The upper ridge is approximately 5 m thick at Stop 2, but 200 m to the north it thins to less than 1 m - a clear illustration of the wedge-like geometry of the alluvial fan lobes. Both Stop 1 and 2 represent "dry" alluvial fan cycles.

### Stop 3. Laurel Brook, Middlefield.

Laurel Brook contains up to 30 m of well-exposed alluvial fan conglomerate, lake margin sandstone, and lacustrine dark shale (Sub-facies 3, 7, 8). The outcrops are 0.5 km south of Laurel Brook reservoir and 0.75 km west of the weigh station/parking area along Rt. 17, south of Middletown (Middletown 7.5 min. quadrangle). This locality illustrates the typical intercalation of Portland alluvial fan and lacustrine strata. Similar sequences can be seen at Prout Brook and Long Hill Brook, Middletown, and along the south bank of the Connecticut River in north Portland.

The complete transgressive - regressive lacustrine cycle at Laurel Brook is composed of the following sub-facies, from base to top: 3 - 7 - 8 - 7 - 3. The strata surrounding the central microlaminated black shale have a marked asymmetry. The shallow lacustrine/shoreline strata (Sub-facies 7) below the shale are thinner than the equivalent rocks above. Evidently, the initial lake transgression occurred over a relatively short interval. In contrast, the regressive phase was of substantially longer duration.

The finely-laminated and microlaminated black shale produces abundant well-preserved fishes (redfieldiids and semionotids). Other fossils at this locality include plants, coprolites, and scarce reptile tracks (McDonald, 1975; 1982).

Red mudstones, poorly-sorted conglomerate, and pebbly sandstone overlie the lacustrine cycle. These sediments were deposited on floodplains/mudflats and alluvial fans surrounding the perennial lake. Shoreline features, such as well-sorted bars and wave-generated beach ridges have been recognized in the basal 1.5 m of the upper conglomerate beds.

The stratigraphic sequence at Laurel Brook illustrates the evolution of a perennial lake system and the migration of its onshore - offshore facies as a function of varying depth. This sub-facies assemblage documents a "wet" phase of deposition during early Portland time.

Stop 4. K-F Quarry, Suffield.

A typical sequence of lower-middle Portland "mid-basin" strata is exposed in the quarry of the K-F Brick Company, on the west bank of the Connecticut River, 3.6 km east of Suffield center (Broad Brook 7.5 min. quadrangle). At present, some 20 m of vertical section are visible along the west wall at the southern end of the quarry. These strata are of particular interest because of their lithologic and paleontologic diversity, for the variety of sedimentary structures they exhibit, and for their copper mineralization.

Laterally-extensive, lenticular or tabular units of red-brown, buff-white, or grey, very fine- to medium-grained stream channel sandstone (Sub-facies 5) make up most of the quarry section. The sandstones are seldom greater than one meter in thickness, and are usually well-sorted, well-indurated, calcareous and micaceous. They display planar and distinct upper contacts; the lower contacts of the units are undulatory and scoured into the underlying beds. Laterally, the sandstone bodies terminate abruptly or interfinger with finer-grained deposits. At a few horizons, the sandstones contain angular to rounded clasts of red or grey mudstone; the clasts vary from sand-sized particles to flattened, polygonal plates up to 50 cm in diameter. Interbedded with the sandstones and locally dominant are millimeter- to meter-scale lenses of red-brown, micaceous floodplain siltstone and silty mudstone. Conspicuous among the quarry redbeds are two 20 cm units of fissile, thinly-bedded, partly microlaminated, fossiliferous, "lacustrine" grey-black shale which occur near the base and top of the section.

The exposed rock faces, the extensive rubble piles, and huge quarried slabs at this locality permit both cross-section and bedding-plane examination of numerous primary and secondary sedimentary structures. Large-scale cross-stratification with meter-scale, planar cross-bed sets can be seen on the south wall of the quarry. Megaripples with wavelengths of 1.5 - 2 m are found on the upper surface of a buff-white sandstone along the west wall. Most of the sandstone and siltstone beds are planar-laminated or ripple cross-laminated, some display climbing-ripple lamination. Oscillatory ripples, current ripples, and parting lineations are abundant on certain bedding surfaces, as are polygonal dessication cracks, flute casts, raindrop imprints, foam impressions, swash and wrinkle marks, and a variety of tool marks. Syndepositional soft-sediment deformation structures include load casts, ball-and-pillow structure, and convoluted bedding. Later deformation produced small faults with well-developed slickensides, gentle folds, and small, orthogonal joint sets. Manganese oxide dendrites and calcite druse coat fracture and bedding surfaces in some units; dolomitic nodules and mottling occur in some of the red mudstone-siltstone beds.

The quarry exposures contain a variety of fossils. Trace fossils are conspicuous in the red-brown siltstone and mudstone;

most abundant are horizontal and vertical burrows of the Scoyenia type, probably made by non-insect arthropods, possibly crayfish (Olsen, 1980; Gierlowski-Kordes, 1985). These linear burrows are from 0.1 to 1.0 cm in diameter and up to 30 cm long, are often branched, and possess irregular, knobby, longitudinal ridges on the external surface. Faint meniscate infillings are visible in a few specimens. An unusual, sinuous crawling trace 21 cm long and 1 cm wide was collected at the site. It has lobate furrows made by appendages, and may have been produced by the arthropod responsible for the Scoyenia burrows. Smooth, usually horizontal, cylindrical burrows, 1-2 mm in diameter are also present. Indistinct reptile footprints occur sparingly on bedding surfaces; types represented include the tracks of small and large dinosaurs (Grallator, Anchisauripus, Eubrontes) and possibly those of small, crocodile-like reptiles (Batrachopus).

Most of the non-red rocks at the quarry contain copper-rich, coalified plant remains, mainly woody stem and branch fragments up to 0.5 m long. A few stem fragments display the longitudinal striations and constricted nodes characteristic of the horsetail Equisetites. Well-preserved, carbonized leafy shoots of the conifers Brachyphyllum and/or Pagiophyllum are found at several horizons, sometimes accompanied by ovuliferous cones and cone scales referable to Hirmerella.

In addition to plant fossils, the two thin, grey-black shale units preserve dissociated scales and bones of the holostean fish Semionotus, large coprolites, and rare conchostracans and ostracodes. A 1.5 cm zone of chocolate-brown silty claystone immediately above the upper grey-black shale is particularly rich in invertebrate remains, and has produced conchostracans (Cyzicus and ?Cornia), darwinulid ostracodes, and possible beetle elytra.

The existence of copper minerals in the Suffield - Enfield area is mentioned in colonial land transactions (Gray, 1982). Subsequent discoveries of copper-rich, coalified plant remains in the region led to explorations for productive coal beds and copper deposits (Silliman, 1818; Shepard, 1837), most of which were soon abandoned when the limited extent of the "coal" and copper mineralization was realized. At the K-F quarry, chalcocite and bornite and their oxidation products coat bedding planes and surround and replace carbonized wood fragments in the grey, buff, and white sandstones and organic-rich grey-black shale units (Gray, 1982). Malachite is most conspicuous in mineralized samples; azurite, chrysocolla, covellite, and cuprite occur in small amounts. The concentration of copper minerals in the non-red units suggests hydrothermal emplacement. Local reducing conditions in the sediments were created by decaying organic matter and as a result of hydrogen sulfide production by anaerobic bacteria. This facilitated the precipitation of insoluble sulfides from copper-rich groundwaters. Gray (1982) proposes that the copper mineralization took place prior to significant compaction of the sediments.

The lithic character, sedimentary structures, and fossils of the K-F quarry strata collectively suggest deposition in braided river - floodplain environments. The geometry of the sandstone bodies and the dune-scale bedforms present in some units characterize broad, meandering, usually shallow, sand-bed rivers with braided channels delineated by bars. During frequent periods of flooding, rivers undercut the fine-grained margins of the main channels and incised smaller, secondary, crevasse-splay channels in the floodplains. Mud clasts of various sizes and abundant plant debris were incorporated into the channel sands. During the waning stages of flood events, fine sand, silt, and clay formed climbing-ripples and planar beds on the adjacent floodplains. The abundant dessication features, mud-draped surfaces, and burrowed zones found on the upper contacts of many sandstone/siltstone units indicate that the episodes of fluvial activity were punctuated by short-duration (probably seasonal) dry intervals. The density of well-preserved, little-transported conifer remains in some beds suggests that at least parts of the floodplain and possibly stream channel bars and islands were heavily vegetated. The thin beds of red and grey-black "lacustrine" shale may have formed in oxbow lakes or small floodplain ponds rich in organic debris.

#### Stop 5. Stony Brook, Suffield.

One of the largest and most informative exposures of the lower-middle Portland Formation in north-central Connecticut is revealed along the north and south banks of the Stony Brook ravine, 3.4 km southeast of the town of Suffield and 0.8 km west of the intersection of Stony Brook with the Connecticut River (Windsor Locks and Broad Brook 7.5 min. quadrangles). More than 60 meters of section outcrop on the near-vertical walls of the gorge.

The bulk of the Stony Brook section is composed of thinly-bedded to massive, irregular- to wavy-laminated, pale red to brick red, calcareous stream channel sandstone, floodplain siltstone, and minor red shale. Typically, these redbeds are highly micaceous and intensely burrowed; ripple marks, dessication cracks, and reptile footprints are locally abundant on bedding surfaces. The succession of oxidized fluvial deposits is interrupted by at least two large, asymmetrical cycles of grey-black lacustrine strata and a number of smaller non-red units.

The best-exposed lacustrine cycle is found along the south bank of the stream 0.2 km west of the Rt. 159 bridge, and is noteworthy because of its unusual lithologies and abundance of fossils. The 6 m thick lacustrine sequence is characterized by an alternation of well-indurated lenses of well-sorted, ripple cross-laminated, fine- to medium-grained, calcareous grey sandstone and interbedded units of friable, thinly-bedded, red-grey, grey-green, and black calcareous siltstone, shale, and

limestone. Though the sandstone beds form prominent ledges in outcrop (up to 0.6 m thick in the western portions of the exposure), they thin markedly east or downdip. Paleocurrent directions derived from current or climbing ripples and linear scour marks in the sandstone units trend N 30°E to N 50°E. The siltstone and claystone units usually thicken in an easterly direction. Near the base of the section are two thin lenses of grey-black, algal-laminated silty limestone with well-defined "loop structures" and/or planar laminations. The limestone beds pinch and swell, but do not exceed 5 cm in thickness. Normally, they have distinct mudcracked, burrowed, or loaded upper and lower contacts. Three meters from the base of the cycle, a 0.4 m disrupted bed of grey, sandy siltstone contains nodular carbonate septaria and massive, elongate nodules of silty limestone up to 30 cm in length. These carbonate nodules or lenses may be stromatolitic mounds.

Three thin lenses of calcareous intraformational conglomerate are the most distinctive units in the lower lacustrine cycle at Stony Brook. These conglomerate lenses vary in thickness along the length of the outcrop and range from 0.5 cm to 25 cm. The three lenses are similar in structure and overall composition, but differ slightly in clast size and lithology and matrix composition. Typically, the intraformational conglomerate is composed of a diverse assortment of the following: 1) very angular to well-rounded, sand- to pebble-sized limestone or dolomitic limestone peloids; 2) concentrically-zoned ooids, pisoids, and probable oncolites; 3) well-rounded grey siltstone or black shale clasts up to 9 cm in diameter; 4) micrite clasts enclosing fish scales, bones, and coprolites; 5) cylindrical, tufa-coated, micrite-filled or hollow structures (probable plant stem or branch casts); and 6) large, complex intraclasts and fragments of the foregoing. These sediments are contained in a matrix of grey or red-grey, micaceous, highly calcareous, fine-grained sandstone or siltstone. The middle and upper conglomerate lenses display poorly-defined graded bedding. The lower lens is reverse graded. The upper and lower contacts of the intraformational conglomerate lenses and the enclosing sandstones and siltstones are undulatory and sharp. Because of the abundance of carbonate cement, the lenses of conglomerate are very well indurated; the fresh rock is very hard and splits along irregular fractures parallel to bedding or along vertical, planar joint surfaces. When weathered, the carbonate clasts are readily dissolved and the rock assumes a characteristic vesicular appearance.

A large portion of the lacustrine strata at Stony Brook is fossiliferous. Networks of branching, smooth-walled, mm-scale, horizontal and vertical burrows are common on many bedding surfaces; larger burrows of the *Scoyenia* type are less common. Carbonized plant stem, branch, and leaf fragments are found throughout the cycle and are particularly abundant in some of the grey sandstone and intraformational conglomerate units. Much of the plant material consists of finely-macerated,

taxonomically-indeterminate debris, but Brachyphyllum-like (conifer) leafy shoots and Equisetites (horsetail) remains can sometimes be recognized. The lenses of intraformational conglomerate enclose masses of very large, diamond-shaped or square fish scales and isolated skull, shoulder girdle, and fin bones referable to some of the more robust species of Semionotus. Black, oval, scale-filled coprolites up to 5 cm in length occur in some conglomerate beds; these may indicate the presence of the coelacanth Diplurus longicaudatus (McDonald, 1982) or fish-eating reptiles. Thin, highly calcareous, ooid-filled zones in the middle lens of intraformational conglomerate contain scarce, well-preserved darwinulid ostracodes and ovate conchostracan (clam shrimp) valves. The conchostracans, referable to Cyzicus sp., range up to 4 mm in length and possess calcite-replaced shells with distinct concentric costae.

The most notable fossils in the Stony Brook rocks are the bivalved mollusks contained in the conglomerate lenses and associated grey sandstones. These pelecypods are the only known in-situ mollusks from the Portland Formation and the locality is one of only three sites in the Connecticut Valley Newark where mollusks have been found (McDonald, 1982; 1985). The bivalves occur as detailed internal and external molds and casts and compare favorably with the previously described species of Unio (freshwater clams or mussels) found in a glacially-transported boulder near Wilbraham, Massachusetts (Emerson, 1900; Troxell, 1914). The Stony Brook specimens are narrowly elliptical in outline, with a gently convex beak and long, well-defined lateral teeth; typical individuals are about 5 cm long and 2 cm high. The concentric growth lines of the shell are preserved in several specimens. Nearly all the unionids are fully articulated with closed and unbroken valves. Individuals preserved in cross-section display minute dissolved shell voids about 0.2 mm in thickness. The articulated, undamaged condition of the bivalved remains and their abundance in certain layers strongly implies that these fossils are autochthonous. This conclusion is further supported by the discovery of presumed mollusk dwelling/escape burrows in the uppermost intraformational conglomerate and the immediately overlying grey sandstone. The larger burrows are straight, roughly cylindrical, unbranched, vertical to sub-vertical structures up to 10 cm long and 2.5 cm in diameter. Along the edge of the burrows, the sediment laminae are bent downward, a diagnostic feature of escape traces (Reineck and Singh, 1980). The deepest portions of many burrows are excavated into the upper few centimeters of the intraformational conglomerate; most burrows terminate upward in ripple cross-laminated, calcareous silty sandstone. Mollusks have not yet been found within the burrows, but they are common at the appropriate horizons. The present-day Unio is a filter-feeding bivalve with an incompletely fused mantle; it lives exposed on lake and stream channel floors, or it can burrow a short distance into the sediment (McKerrow, 1978). Evidence suggests that the

Stony Brook bivalves were semi-infaunal, in-situ residents in the limy conglomeratic and sandy substrate.

The strata exposed in the basal lacustrine cycle at Stony Brook are interpreted as shoreline and nearshore deposits which formed at the western margin of a large, perennial, Portland lake. Periodic fluctuations of lake level account for the interfingering of coarse and fine, clastic and chemical, oxidized and reduced sediments. During deposition, the fluvial-deltaic, mudflat-sandflat, and shoreline environments of the Stony Brook lake were alternately subjected to varied-length episodes of shoaling and subaerial exposure followed by periods of lake expansion.

The lenses of westward-thickening, well-sorted grey sandstone which dominate the section may represent prograding delta lobes which built into the lake from the southwest. The load structures that often mark the sandstone-shale contacts were developed from the accumulation of coarse sediment on top of water-saturated and non-compacted mud. The siltstone-shale units are usually bioturbated and probably formed on broad, partially-oxidized nearshore mudflats or perhaps in shallow lagoons along the shoreline.

Wave agitation and local reworking of the carbonate sediments produced the peloids, ooids, pisoids, oncolites, and intraclasts which were later incorporated into intraformational conglomerate lenses. The presence of large clastic and carbonate rip-up clasts in the intraformational conglomerate suggests that the conglomerate may have formed during a transgressive lacustrine phase when low-standing lake waters re-advanced over previously deposited sediments. Another possibility is that the conglomerate beds are storm deposits or "tempestites". The variety of clast types, the combination of very angular and very well-rounded clasts, the lack of imbrication, the occurrence of massive plant fragments and dissociated fish remains, and the presence of localized, small-scale hummocky cross-stratification supports this conclusion. The character of these deposits closely conforms to the criteria established by Kreisa (1981) for the recognition of storm deposits.

Poorly represented in the Stony Brook section are the reduced, sulfide-organic-rich, rhythmically-laminated, fossiliferous, "deep water" black shale strata that characterize many other Connecticut Valley Newark lakes. However, rounded, pyrite-rich clasts of black shale occur in some of the intraformational conglomerate layers, so this facies no doubt existed offshore, presumably to the east.

The diverse invertebrate community and the very large, abundant semionotid fishes imply that the Stony Brook lake was ecologically hospitable and of substantial size. The sediments, sedimentary structures, and fossils are closely analogous to those described by Link and Osborne (1978) for the marginal lacustrine units of the Pliocene Ridge Basin Group of southern California.

## PART III.

Road LogMiles

- 00.0 Assemble at WESLEYAN UNIVERSITY, Middletown, at the loading dock behind the SCIENCE TOWER, on Pine St. Proceed left (south) on Pine St. to the intersection of Pine St. and Randolph Rd.
- 1.5 Turn left (east) on Randolph Rd.
- 1.6 Turn right (south) on Rt. 17.
- 2.0 Bear left on Coleman Rd. Ascend the north flank of the Round Hill Fan complex. Note outcrop of conglomerate and sandstone along the left (east) side of the road.
- 3.0 Park just before the intersection of Kelsey St. and Coleman Rd. Walk northeast along a dirt trail through woods to encounter several linear ridges of alluvial fan conglomerate.
- STOP #1: Coleman Road. Alluvial fan deposits and debris flows.
- Return to cars and proceed to intersection of Coleman Rd. and Kelsey St. Turn left (east) on Kelsey St.
- 3.3 Turn left (north) on Mapleshade Rd. Note bedding plane - dip slope exposures of conglomerate along the left (west) side of the road.
- 4.0 Park on Mapleshade Rd.
- STOP #2: Mapleshade Road. Alluvial fan deposits.
- Proceed into fields on left (west) side of road and continue northwest into lower field. Follow outcrop until the south end of the lower ridge is encountered. The stop description begins here. After examining the lower outcrop, walk back to upper field and continue east to the prominent cliff in the northeast corner

of the field.

Return to cars. Continue along Mapleshade Rd. (north).

- 4.5 Turn left (west) on Randolph Rd. Proceed to traffic light at intersection of Randolph Rd. and Rt. 17.

Turn left (south) on Rt. 17. Note excellent exposure of coarse sandstone (Sub-facies 5) behind Monte Green Inn along the left (east) side of road. Round Hill (elevation 582 ft.) can be seen along the east side of the road, just past Dooley Pond.

- 6.6 Bear right into the abandoned weigh station/parking area opposite the Pizza King restaurant. Park and walk due west (right) across fields approximately 1/2 mile to encounter north-south-trending Laurel Brook. Follow brook north (downstream) to prominent chasm or gorge.

STOP #3: Laurel Brook. Alluvial fan and lacustrine strata.

Return north on Rt. 17 from parking area.

- 8.9 Turn right (east) on Randolph Road (Rt. 155). (10.0 mi.) This slope parallels the east-dipping bedding of the conglomerates on the northeast flank of the Round Hill Fan complex. (10.9 mi.) This steep, west-facing hillslope marks the approximate location of the eastern fault margin of the Hartford Basin. To the southwest the Round Hill Fan complex can be seen.

- 11.1 Turn right into the entrance ramp for Rt. 9 North (Hartford). Exposures of the Paleozoic basement rocks are visible on both sides of the ramp. Note the panoramic view of the valley (11.4 mi.). The prominent ridge on the western horizon is formed by the Holyoke Basalt, the thickest of the three basalt units in the basin.

- 18.9 Bear right on to I-91 North (Hartford).

Exit 23: Optional side trip to Dinosaur State Park (20.6 mi.) for excellent exposures of dinosaur trackways in lake margin sandstone and siltstone of the Lower Jurassic East Berlin Formation.

(20.8 mi.) The contact of the Hampden Basalt with the underlying East Berlin Formation can be

seen along the right (east) side of I-91.  
(21.0 mi.) Exposures of the East Berlin Formation  
are located along the right side of the highway.

- 28.3 Keep to the right and follow I-91 North past  
Hartford.
- 46.4 Take Exit 17 West (Rt. 190) to Suffield. Note  
outcrops of the Portland Formation (Sub-facies 6)  
along the highway just before the bridge over the  
Connecticut River.
- 48.0 Turn left (south) on Rt. 159.
- 48.6 Turn left into unmarked dirt road and entrance  
to K-F Quarry.

STOP #4: K-F Quarry. Braided stream and floodplain strata.

NOTE: This is an active quarry and access  
to this stop is restricted.  
Permission must be obtained in  
advance.

Continue south on Rt. 159.

- 50.3 Park along the road just before bridge over Stony  
Brook. Descend to brook along left (east) side of  
bridge and walk west (upstream) approximately 1/4  
mile to large exposures on south bank of stream.

STOP #5: Stony Brook. Fluvial/lacustrine deposits.

-END TRIP-

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**Trip C-1: Geology of Southern Connecticut, North-South Transect.**

John Rodgers and Brian J. Skinner.

This trip crosses the following quadrangles, published quadrangle maps of which are listed below:

	<u>Bedrock Geology</u>	<u>Surficial Geology</u>
(New Haven, only on way to first stop)		CG&NHS QR 18
(Mt. Carmel, only on way to first stop)	USGS GQ-199	CG&NHS QR 12
<b>Southington</b> (Stop 1)	USGS GQ-200	USGS GQ-146
<b>Meriden</b> (Stops 2 to 6)	USGS GQ-738	USGS GQ-150
<b>Middletown</b> (Stops 6 and 7)	CG&NHS QR 8	
<b>Hartford South</b> (Stops 7 and 8)		CG&NHS QR 20
<b>Durham</b> (Stops 9 and 10)		USGS GQ-756
<b>Guilford</b>	CG&NHS Bull 86	CG&NHS QR 28
<b>Branford</b> (Stops 11 and 12)		CG&NHS QR 14

Stratigraphic Column: Newark group in Hartford basin (Central Lowland).

Lower Jurassic

Equivalent dikes, according to Philpotts and Martello, ms.

Portland arkose

"Meriden formation" of  
Krynine (1950)

Hampden basalt (3rd flow)  
East Berlin formation  
Holyoke basalt (2nd flow)  
Shuttle Meadow formation  
Talcott basalt (1st flow)

Bridgeport-Pelham dike  
Buttress-Ware dike  
Fair Haven and Higganum dikes

## Upper Triassic

New Haven arkose

**DON'T FORGET TO SET YOUR ODOMETER TO 0.0 BEFORE STARTING OUT!**

(The mileage for this itinerary was measured on a car whose odometer is about 1% fast.)

To reach STOP 1 from New Haven, take Whitney Avenue north.

**Mileage**

- 0.0** Corner of Whitney and Humphrey Streets, at outlet of Kline Parking lot. (Turn left if leaving parking lot.)
- 1.1** Enter town of Hamden. As road ahead bends to left, one sees to the right the dam of Lake Whitney, which covers the original dam for water power for the works of Eli Whitney, the inventor of standardized parts and the cotton gin. An industrial museum has been established at the site, the location of his gun factory. The dam was located where Mill River had a natural waterfall over the Mill Rock dike, which connects the East Rock sill, visible to the right (east), through Mill Rock, to the left (west), and Pine Rock to the south end of the West Rock sill in West Rock. The dike is exposed on the left (west) side of Whitney Avenue at 1.4.
- 4.2** Enter Mount Carmel quadrangle.
- 4.8** Village of Centerville. From here north Whitney Avenue is Route 10.
- 7.5** Traffic light in village of Mount Carmel. At this point the road cuts diagonally across the feeder dike for the Mt. Carmel or Sleeping Giant stock, which is visible on the right just beyond the corner.
- 10.0** Enter town of Cheshire.
- 13.1** Village of Cheshire; enter Southington quadrangle. Continue on Route 10.
- 16.5** The West Rock sill, which for about 13 miles north of West Rock in New Haven lies near the base of the Triassic New Haven arkose, at its north end cuts upward in the sequence (though not as sharply as along the Mill Rock dike at its south end). Here it expands into a sill that is like but much smaller than the East Rock sill. Outcrops are visible in the little ridge just west of the road for the next mile.

- 17.5 Enter town of Southington. The sill here crosses the road obliquely; it disappears about a mile farther north-northeast.
- 17.6 Intersection of Routes 10 and (old) 66. Keep straight or bear right where Route 10 bears left; at traffic light, turn left (west) on Route 66.
- 18.9 Intersections with Route I-84. Continue straight (route becomes Route 322 but isn't marked).
- 19.5 Traffic light in village of Marion. Turn right (north) on Marion Road.
- 20.2 Blinker. Marion Road bears right; turn left on Mt. Vernon Road.
- 20.3 Turn sharp right with Mt. Vernon Road (road ahead is dead end).
- 22.6 Corner of Roaring Brook Drive; turn left (west) up drive and go to circle at end (22.9).

MILEAGE NOW STARTS AT 0.0 AGAIN

**STOP 1:** This is Locality 10 of Wheeler (1937), 39 of Krynine (1950; see Roaring Brook in index). Proceed across vacant lot to right (north) of circle and, bearing right, enter woods and descend into gorge of Roaring Brook. At first, outcrops are of (Devonian?) Southington Mountain schist (the type locality is the mountain rising to the west), but downstream a short distance, the schist on the north (left) bank is overlain unconformably by the basal conglomerate of the Triassic New Haven arkose. Note that the unconformity, projected upstream, will not pass over Southington Mountain, which is schist to the top. Hence a fault must pass between this locality and the mountain. Moreover, according to Fritts (GQ 200), another fault passes between this outcrop and a prominent outcrop of New Haven arkose (also conglomeratic) not far downstream on the south (right) side of the brook. These faults are interpreted as western branches of the main Mixville fault, whose existence at the western border of the Connecticut Valley Mesozoic basin is proved at several places by drilling or excavation (as in the Bristol copper mine, 13 km (8 miles) north of this locality), though it leaves the border and cuts into the older rocks about 20 km (12 miles) south of here. Krynine's petrographic work confirmed Barrell's original conclusion that the Triassic arkose is not derived from the underlying schist to any considerable degree, but comes from the Eastern Highlands, 27 km (17 miles) east of here.

- 0.0 Return down Roaring Brook Drive.
- 0.3 Turn right (south) on Mt. Vernon Road.

- 2.6 Right-angle turn to left (east).
- 2.7 Stop sign and blinker. Turn right (south) on Marion Road.
- 3.4 Traffic light in village of Marion. Turn left (east) on Route 322.
- 4.0 Crossing I-84 (at its Exit 27). (Just beyond, road becomes Route 66 at present; this may soon change.) Fine view of Hanging Hills of Meriden; cliffs are Holyoke (2nd) lava flow. This flow, dipping east, forms the (irregular) north-south mountain (Talcott Mtn.) to the north (left), but in the Hanging Hills it dips gently north. The prominent bench (marked by pines) halfway down is held up by the Talcott (1st) flow with the same dips.
- 5.3 Intersection with Route 10; continue east on 66.
- (5.7- (Brief excursion into town of Cheshire and return into  
5.9) Southington.)
- 6.1 Enter Meriden quadrangle.
- 7.7 Traffic light at exit from Route 66 (present west end of four-lane divided highway). Park on right at light (until it becomes the entrance ramp for 66 west).

**STOP 2:** Cross to far side of exit ramp at light and walk east to outcrops. Excellent display of relatively fine-grained stream (channel and flood-plain) deposits of New Haven arkose. Dip here is gently north under the lava flows of the Hanging Hills, instead of the usual slightly steeper east dip.

Proceed on old Route 66 (Main Street of Meriden).

- 7.9 Enter town (and city) of Meriden.
- 8.6 Traffic light; exit from Hubbard Park on left (north). Continue on Main Street; we will return here.
- 8.8 Entrance to Hubbard Park on left (north). Continue on Main Street.
- 9.3 First of three traffic lights; continue (slightly left) on Main Street.
- 9.5 Third traffic light; intersection with Centennial Avenue. Turn right (south) on Centennial Avenue and proceed past another traffic light (at which street becomes Oregon Road).

- 11.1 Oregon Road crosses bridge (Quinnipiac River) and ends in T-junction. Park where you can (best on right, just beyond corner).

**STOP 3:** Hanover Pond (pond is just to east). Outcrops on south side of road across river; all are fairly conglomeratic New Haven arkose. Close to intersection (and along Quinnipiac River to west) they dip gently north, but to east along road, they dip more and more steeply west, quite unusual for the Triassic-Jurassic rocks of the Connecticut basin, and farther east several rather small faults can be demonstrated, mostly with east side upthrown. As shown on Hanshaw's map (GQ 738), we are close to the projection of perhaps the largest fault in this part of the Connecticut Mesozoic basin (Cathole Mtn.-Lamentation Mtn. fault, to name it for the two offset portions of the Holyoke flow). Turn around and proceed back (north) along Oregon Road.

- 12.7 Traffic light at Centennial Avenue and Main Street; turn left (west) on Main Street.
- 12.9 Traffic light; bear right on Main Street. Go slow and watch for next turn, which is blind.
- 13.4 Just beyond projecting outcrop, turn right (north) into Hubbard Park.
- 13.7 Road intersections in park; continue straight ahead under overpass of Route 66 and up hill beyond.
- 14.0 Quarry to right is in Talcott flow, which displays pillows and poison ivy. We will see the pillows better at a later stop.
- 14.2 South end of Lake Merimere.

**STOP 4 (Optional):** We are here close to the top of the Talcott flow, which can be seen, dipping north, at the top of the outcrops on the right side of the road. The cliffs above us on both sides of the lake are the Holyoke flow, also evidently dipping north. The little island in front of us appears to be a horst of the Talcott flow; the lake is artificial but the valley was excavated along one or more minor faults.

Continue north along east side of lake.

- 14.4 Outcrops of sediments on right; Shuttle Meadow formation. Rather fine-grained and thin-bedded flood-plain or lake sediments. (Enter town of Berlin.)
- 14.6 Outcrops of base of Holyoke flow on right.
- 14.9 Road climbs to north-dipping top of Holyoke flow and descends dip slope.

- 15.1 North end of Lake Merimere. Turn left (west) on rather poor road and follow it, first on strike along foot of dip slope (to 15.6), then (re-entering town of Meriden) up the dip slope south.
- 16.5 Forks; take left fork towards East Peak. Note cliffs below to right.
- 16.9 Bear right into East Peak parking lot; park at upper end of lot (17.0) if feasible.

**STOP 5:** East Peak of the Hanging Hills. The base of the tower stands about 955 feet above sea level, whereas Hanover Pond (at STOP 3), visible off to the southeast, is 87. We are standing approximately on the upper surface of the Holyoke lava flow, which holds up all the highest hills in the Valley except those on the West Rock, Mount Carmel, and Barndoor intrusives. The characteristic double jointing in the Holyoke flow is particularly well displayed here. The older columnar joints show narrow altered (silicified?) selvages, which weather in slight relief above both the normal rock and the joint itself; the younger systematic joints cut undeflected across the columnar joints, implying that they had been entirely healed. Large boulders of this rock showing the characteristic jointing pattern are scattered over the countryside from here to Long Island Sound. The well known "Judges Cave" on West Rock in New Haven is a group of such boulders so placed as to provide some shelter; in 1661, three regicides (Dixwell, Goffe, and Whalley) hid there from the agents of Charles II.

The Talcott flow forms the bench at our feet, between us and the lake in Hubbard Park. The Hampden flow forms low ridges in the country to the northeast, beyond the dip slope on the Holyoke flow, but they are not clearly seen from here. The city of Meriden, spread out before us to the southeast, and all the country to the south is underlain by the New Haven arkose, drained by the Quinnipiac River to Long Island Sound at New Haven Harbor.

In the Hanging Hills, the Holyoke flow and the beds above and below strike nearly east-west and dip gently north, in strong contrast to their normal north-south strike and moderate east dip; the change is evidently associated with the particularly intense faulting in the Meriden region (see USGS GQ 738, 494; CG&NHS QR 8) and especially with the large sinistral offset in map pattern caused by the largest of these faults (sinistral offset does not prove sinistral strike slip, of course).

To the east across Lake Merimere is South Mountain, Holyoke lava displaced only a little to the left from what we are standing on; half hidden behind it is Cathole Mountain, displaced somewhat more. The largest fault (or group of faults) then offsets the Holyoke flow 8 miles to the northeast, to the north end of the Lamentation Mountain, the northernmost of the north-south mountains in the middle distance to the east, in which the Holyoke flow resumes its normal strike and dip. The main Hartford line of the New Haven Railroad and the Berlin Turnpike (visible at the foot of the mountain) go through the gap between, the lowest divide into the Connecticut River drainage (175 feet). Chauncey Peak, the south end of Lamentation Mountain, is slightly offset from the

rest. Another fault then displaces the flow another 5 miles to the northeast, to the north end of Higby Mountain; Rt. I-91 goes through this gap. Smaller gaps are visible in the ridge from Higby Mountain south, each caused by a smaller offset along a similar fault; Rt. 66 and the old Air Line of the New Haven Railroad follow two of these gaps toward Middletown. From here it is not easy to pick out the larger gaps between Pistapaug Mountain (gap used by Rt. 17) or between Totoket Mountain and Saltonstall Ridge (used by Rt. 80); the latter gap is not caused by faulting within the Triassic but by a transverse anticline that abuts southeastward against the eastern border fault, interrupting the outcrop of the Holyoke flow (we will pass along it after 69.1).

Off to the north of Lamentation Mountain is Cedar Mountain, again upheld by the Holyoke flow brought up along a northern branch of the fault behind Cathole Mountain. The Hampden flow east of Cedar Mountain can be traced onto the Trinity College campus in Hartford. On a good day, the insurance towers of Hartford can be seen behind Cedar Mountain.

Behind Higby and Lamentation mountains and off to the northeast are the Eastern Highlands, separated from the Central Valley by the eastern border fault and the chief source of the Triassic and Jurassic sediments. Due east of us, one can make out the break in the Highlands at Middletown, where the Connecticut River turns away from the valley to find its way through the Highlands to the Sound.

In the opposite direction, the Holyoke flow extends west to West Peak (1,024 feet above sea level) and then turns abruptly north, resuming its normal strike and dip. Thence it extends north for many miles, though broken and somewhat offset by faults, forming Talcott Mountain west of Hartford and reaching Mt. Tom and Mt. Holyoke, on opposite sides of the Connecticut River in central Massachusetts - these can be seen from the top of the tower on a very clear day. (This spring the tower was closed to visitors, because the iron staircase is rusting out.)

Beyond West Peak is the valley underlain by the New Haven arkose, and behind that the Western Highlands. To the south, however, the West Rock sill appears, first as low hills within the valley, then higher and higher in front of the Western Highlands until Mount Sanford reaches the skyline and hides them. Out in the valley southwest of Mount Sanford is the large mass of Mount Carmel or the Sleeping Giant, an irregular sill or stock higher in the New Haven arkose than the West Rock sill and probably nearly above the main basement feeder dike.

Just to the left of and behind Mount Carmel, on a clear day one can see the Civil War monument on top of East Rock in New Haven and behind that the waters of Long Island Sound and Long Island. Thus one can see entirely across the 55-mile width of Connecticut, to points in Massachusetts and New York State 95 miles apart.

From the latitude of the Hanging Hills south, the hills upheld by the Holyoke lava and the Mount Carmel and West Rock sills reach to heights that decline steadily southward, reaching sea level around New Haven Harbor; the slope is about 45 feet per mile (8 meters per kilometer). From the Hanging Hills north, however, no peaks on the Holyoke flow reach 1,000 feet until Mt. Tom (1,200 ft.) and the Holyoke Range; the slope

from West Peak to Mt. Tom would be only about 4 feet per mile (less than a meter per kilometer). The sloping hill-top surface to the south is continuous with the surface beneath the Cretaceous rocks on Long Island - for this reason, from the Hanging Hills Long Island Sound and Long Island appear higher than any of the hills between - and it therefore represents the Fall Zone surface or facet (Flint, 1963); even the highest hills to the north have been reduced by erosion well below this surface. One can therefore imagine that when that erosion was going on, Cretaceous rocks still reached inland as far as Meriden. As Barrell pointed out long ago, it is probably no coincidence that the Connecticut River deserts the Connecticut Valley just at this latitude. Return to Lake Merimere, south along it, and back down under the overpass of Route 66.

- 20.3 (= 13.6 above) Corner in Hubbard Park; must turn right; then take next left (20.4) and proceed to traffic light on Main Street.
- 20.6 (= 8.6 above) Turn right (west) on Main Street.
- 21.5 Turn left onto ramp for Route 66 East; we will drive by the outcrops of STOP 2 and others, all dipping gently north under the Hanging Hills and its lower bench (Talcott flow).
- 23.3 Exit 5; continue on Route 66. Road here approaches base of Talcott flow, which crops out in the northern entrance ramp of Exit 5. Just beyond, however, the road crosses the Cathole-Lamentation fault, upthrown to the east, and the cuts are in beds much lower in the New Haven arkose, which dip east at a fairly high angle. Higby Mountain, east-dipping Holyoke flow, bars the view eastward, and the bench formed by the Talcott flow is visible beneath it. Continue to exit 12.
- 26.8 Take Exit 12 off Route 66.
- 27.2 Turn left (north) on Preston Avenue.
- 28.0 Take entrance ramp onto I-91 North (Exit 19), and park near top of ramp but not on main highway.

**STOP 6:** Proceed north to outcrops along right (east) side of I-91. These are the Talcott flow, partly vesicular, in places showing excellent pillows, especially on glacially smoothed and striated surfaces. (The top of the flow could be seen still farther along, but the outcrops have been degraded by weathering and geologists.)

- 28.2 Continue north onto I-91, following it to Exit 21. Enter town of Middletown and Middletown quadrangle. At first, road follows swampy belt over Shuttle Meadow formation between bench (Talcott flow) to west and Higby Mtn. (Holyoke flow) to east, but about where it begins to descend, it approaches a major fault (offsetting Higby Mtn. from Lamentation Mtn. or its southern satellite, Chauncey Peak) that cuts off the Talcott flow to the west and then bevels the Holyoke flow, which crops out along the road. At about 31.3, one catches a glimpse of Westfield Falls in a gorge just east of the road; the falls is over virtually the north tip of the Holyoke flow as it is cut off by the fault.
- 32.9 Mattabassett River; enter town of Cromwell.
- 33.0 Exit 21; turn off toward Route 72.
- 33.3 Traffic light at Route 72. Turn left (east).
- 33.7 Traffic light; turn left (north) on Coles Road.
- 34.3 Turn left (northwest) on North Road Extension, which goes under I-91 and becomes North Road.
- 35.1 Cross-roads; continue straight ahead on road marked No Outlet.
- 35.5 Park as road descends toward brook, and climb well beaten path into brush to right.

**STOP 7:** Large highway cuts built for projected extension of Route 9 west of its intersection with I-91. (These cuts lie astride Middletown-Hartford South quadrangle line.) They display about the upper half of the East Berlin formation and its contact with the overlying Hampden (3rd) flow. The East Berlin formation exposed here consists of a triple alternation of fine-grained red flood-plain deposits and even finer grained gray and black lake deposits (the cycles are about 10 meters thick, except the uppermost). At the top of each group of lake beds is a sandstone layer. Carbonate (mainly dolomite) is common, especially in the lake beds and the overlying sandstone. The sandstone at the top of the highest lake beds is the level of the dinosaur trackways at Dinosaur State Park, our next stop. A few tracks have been observed here, but the nature of the outcrops (vertical faces instead of pavement outcrops) is unfavorable to finding them. Plant remains and fish scales have also been found in this cut, but in general fossils are rare here.

The Hampden flow is formed of several flow-units (Chapman, 1965), as can be seen here. Vesicles and amygdaloids are visible in the tops and bottoms of some of the flow-units. Fine specimens of amethyst have been found in the Hampden flow in this cut. Turn around and return to I-91, turning onto ramp for I-91 N at 37.7 (=33.3 above). For next 3 miles, route follows east foot of dip slope of Hampden flow.

- 39.2 Intersection with Route 9; the cross-ramps, over and under, lead to the cuts visited at STOP 7. Continue on I-91. Enter Hartford South quadrangle.
- 39.9 Enter town of Rocky Hill.
- 41.0 Take Exit 23; just ahead, dip slope swings across I-91 (good cuts at crest of hill ahead), and exit ramp climbs up it.
- 41.4 End of ramp at West Road, on crest of ridge on Hampden flow. Turn right (east) on West Road.
- 41.8 A small fault offsets the ridge to the right, and road drops off it. Dip here is south, off Rocky Hill anticline.
- 42.1 Entrance to Dinosaur State Park on right.

**STOP 8:** In 1966, the State Highway Department chose this site for a central Highway Department Research Laboratory, close to but not on Interstate I-91 near the geographic center of the state. One Friday afternoon in August, one of the bull-dozer operators, Mr. Ed McCarthy, engaged in clearing the overburden to bedrock before construction, turned up flat slabs of sandstone on which he recognized some large dinosaur footprints (such prints have of course been well known in the Connecticut Valley for 150 years). After investigating, the project engineer, Mr. Tom Jeffreys, stopped excavation in the area and called the Yale Peabody Museum, the University of Connecticut, and the newspapers; later an announcement was broadcast on TV, and the Saturday Hartford Courant carried the story. As word of the find spread, many persons came down over the weekend to pick up examples for their patios, rock-gardens, etc. The news also reached Ms. Jane Cheney, Director of the Children's Museum in Hartford, who went directly to Governor John Dempsey (about to stand for re-election) and persuaded him that the find was exceptional and should be preserved. At a meeting of state officials on Monday morning, it was agreed that Peabody Museum would direct the bull-dozer operators while they determined the size and significance of the deposit; Prof. John Ostrom of the Museum and Prof. Joe Webb Peoples of Wesleyan University, then Director of the Connecticut Geological and Natural History Survey, were in general charge. Later the Governor declared the locality The Dinosaur State Park. A news item concerning the dinosaur trackway appeared on the front page of the Hartford Courant for twelve straight days. Clearing continued for several weeks, until a single surface of sandstone displayed over two thousand tracks. Testing elsewhere on the property showed that the layer with the tracks was even more extensive; moreover it is only one of five layers within about 2 meters of rock that display tracks.

By this time, it was thought that enough had been uncovered to make a spectacular display, and the work was stopped; the main concern after that was to preserve the tracks against the approaching winter's freezes and thaws. The tracks were therefore covered up, and, except for one or two brief spells, the main discovery site has not been uncovered since. On the other hand, in 1967 a more modest area was uncovered west of the

main site (in the same layers), which could be covered by a temporary structure (a plastic bubble kept up by excess air pressure), and this area became the main exhibit at the Park. Later the temporary structure was replaced by the present more permanent structure, but the original plan to build a larger museum over the main, original discovery has never been carried through. In any case, the Park was duly dedicated in 1967 by Governor Dempsey; honor was paid to Mr. McCarthy, the original finder of the tracks; and the Rocky Hill High School Band played a new piece of music called "Dinosaur", written for the occasion by its Director.

As the Hampden flow forms the ridge immediately south of the trackway area, the stratigraphic position is known exactly; when the I-91 cuts about 2 1/2 miles to the southwest were opened, the trackway levels were pinpointed there (see description of STOP 7). While the main trackway was still uncovered, a trench was dug down dip to the south, which showed that the trackway layers are cut down dip by a small thrust fault, dipping south more steeply than the beds, so that the trackways layers are brought back up closer to the land surface. If a museum is ever built over the original site, this trench could be reopened and the thrust fault displayed. Its westward extension is clearly responsible for the right offset of the Hampden ridge between the Park and I-91 (noted at 41.8).

The Highway Department, deprived of their original site, had to recommence operations about a mile farther east, and rumor has it that the bulldozer operators were given strict orders to stop for nothing.

Continue east on West Road.

- 43.0 Intersection with Route 99; turn right (south). Shortly we cross the Hampden flow into the Portland arkose.
- 44.2 Enter town of Cromwell.
- 45.1 Enter Middletown quadrangle.
- 48.0-  
48.3 Route 99 joins Route 9 (south).
- 48.9 Enter town (city) of Middletown. Continue on Route 9 past Middletown exits to exit for Route 17 south.
- 50.1 Take exit for Route 17 S.
- 50.6 Turn left (south) with Route 17 around traffic circle.
- 54.6 Enter Durham quadrangle.
- 54.9 Enter town of Middlefield.
- 55.1 Enter town of Durham. Continue through village of Durham on Route 17.

- 57.2 Forks of Routes 17 and 79; bear right (southwest) on 17.
- 57.4 Forks of Routes 17 and 77; bear left (south) on 77 and prepare to stop.

57.7 **STOP 9:** Outcrop on left (east) side of road. We are still in the Portland arkose, which is normally medium to fine grained, but here it is a polymict fanglomerate. At this locality, the eastern border fault of the Connecticut basin is about a quarter mile to the southeast; as we will see in the rest of the trip, all sedimentary units of the Newark group turn into fanglomerate as they approach the fault, which, as Barrell saw clearly 70 years ago, must therefore have been active during the entire depositional history of the group. The pebbles and cobbles (none larger than about head-size here or anywhere along the fault) are made of rocks from the Eastern Highlands, and indeed by far the largest part of the Newark group in the Connecticut basin is from that source. At this locality, however, a few cobbles of basalt can be found, showing that lavas (volcanoes?) were exposed on the Eastern Highlands, perhaps over the Higganum dike, which can be followed across Connecticut within the Eastern Highlands and here is about 4 miles to the southeast. Or the basalt might be derived from "Foye's volcano" (Foye, 1930), which is exposed about 0.6 miles southwest of here just west of the border fault.

John L. Rosenfeld, during his two years at Wesleyan, instigated a study of the cobbles in the fanglomerates along the border fault and reported (pers. comm.) that the grade of metamorphism in them increases upward; i.e., that the lowest Newark strata contain fragments from lower grade metamorphic rocks, now entirely removed by erosion, and the highest strata contain rocks much more like those to be seen today in the Highlands. Unfortunately the results of this work have not been adequately published (but see Eaton and Rosenfeld, 1960, p. 174).

- 58.1 Road bears to right, swinging to follow the eastern border fault, here marked by a pronounced valley between the schist and gneiss (Collins Hill formation - Ordovician?) to the left (east) and strata of the Newark group to the right (west). "Foye's volcano" is exposed in an overgrown quarry at the foot of the first spur to the right. Our route now follows the border fault, more or less, for about 11 miles, as far as the village of North Branford.
- 59.3 Road swings a little to east into Eastern Highlands. The outcrops around and just north of the farm on the right are of mylonitized and silicified schist along the fault.
- 59.8 Enter town of Guilford.
- 61.5 Road returns to fault at the pass. The low outcrop left of the road is schist and pegmatite; the quarry right of the road is in the Holyoke flow dipping south at the east, cut-off butt end of Totoket Mtn. Within the lava is a large block of phyllite, interpreted as a block that slid off the fault scarp to the east into the flow. We cannot stop here with a large group; those who wish to stop on their own are warned that the quarry is full of poison ivy.

- 61.8 Road cut on right once displayed the eastern border fault, but unfortunately weathering in the forty years since it was cut has obscured the evidence. The main face is in badly weathered Holyoke lava (probably close to the top of the unit), at the south end of the face is even more badly weathered schist, and fresher pegmatite can be found in the floor of the cut. The cut was described in detail by Digman (1950) when it was fresh; between the lava and the schist on the face was about a foot of gouge, into which a shovel handle could easily be driven. The fault dipped 55° west. Beyond, schist and pegmatite are exposed on the left of the road; then the road swings back across the fault.
- 62.2 North end of Lake Quonnipaug, which here fills the fault-line valley. Up a private road to the right (west), there used to be exposed a mass of metamorphic rock surrounded by fanglomerate of the East Berlin formation. As suggested by de Boer (1968, Trip C-5; p. 10), the mass is probably a large landslide block, comparable to but much larger than the mass of phyllite in the Holyoke flow at 61.5. Continue south on Route 77 for 0.4 mi. down the west side of Lake Quonnipaug.
- 62.6 Park either just beyond or just opposite a wall made of large quarried blocks of granite (Stony Creek granite, probably from the Norcross quarry in Branford).

**STOP 10:** Outcrops south of and above the granite wall are the Hampden flow, here dipping 40° south and hence striking perpendicular to the border fault. North of the wall are outcrops of coarse conglomerate like that at STOP 9 (but no basalt is reported). Stratigraphically these beds are exactly at the level of those seen beneath the Hampden flow at STOP 7, but the difference in grain-size is striking. Pegmatite and gneiss of the Eastern Highlands make the outcrops across the lake.

Continue south on Route 77. The route continues to follow the west side of the valley over the fault, which makes several blunt angles. As pointed out by Wheeler (1939), where the fault makes an angle convex to the east, the strata west of it form a basin; where convex to the west, they form an arch. Thus the Hampden flow butts against the fault four times in the next four miles of its course, dipping alternately south (here and at about 65.0) and northeast or east (at about 63.8 and north of 67.3). The fault, a wide gouge zone carrying much water, was exposed in the tunnel of the New Haven Water Company from Lake Hammonnasset to Lake Gaillard, crossing our route a little south of the south end of Lake Quonnipaug (Walton and Lundgren, personal communication).

- 64.1 Enter Guilford quadrangle.

- 65.2 Road corner; exposure of conglomeratic Portland arkose in second of the two basins. At this point the fault angles sharply to the right (west), more or less followed by the road to the right, but we continue straight ahead on Route 77 into the Eastern Highlands rocks, again the Collins Hill schist.
- 65.8 Road crosses Higganum dike within Eastern Highlands (briefly mentioned at STOP 9).
- 65.9 Intersection with Route 80; turn right (west) on Route 80.
- 66.1- Road crosses dike again, this time obliquely.  
66.2
- 67.3 Enter town of North Branford.
- 67.5 Route 80 approaches and enters fault-line valley; road from corner at 65.2 rejoins it. Conglomerate has been exposed or dug up at several points on the north side of the road from here to village of North Branford, and large blocks may be seen here and there.
- 67.7 Enter Branford quadrangle.
- 69.1 First traffic light in village of North Branford; continue on Route 80, which shortly bends to northwest around the south, cut-off butt end of Totoket Mountain, made of the Holyoke flow dipping northeast (cf. 61.5). The road here abandons the border fault and enters an anticlinal valley over Shuttle Meadow formation between Totoket Mtn. (Holyoke flow dipping northeast) and Saltonstall Ridge (Holyoke flow dipping southwest).
- 70.4 Entrance to North Branford traprock quarry, exploiting Holyoke flow.
- 70.7 Junction with Route 22; continue west (bearing left) on Route 80. Route now enters valley over Shuttle Meadow formation between scarp slope of Holyoke flow in Saltonstall Ridge (which outlines another large basin) and dip slope of Talcott flow (rather broken up by faulting); in general the road hugs the foot of the dip slope for the next 6 miles.
- 71.8 Traffic light in Totoket village; continue straight on Route 80.
- 72.3 Enter town of East Haven.
- 72.5 Road bends to right around outcrop of Talcott flow. Near west end of this cut a clastic dike of (baked) arkose cuts the lava flow, which shows chilled margins against it. This stop is worth making but is quite impracticable for a large group. Those wishing to stop should go beyond the outcrop and park as soon as possible, but be very careful on the outcrop as the road is busy and the corner is blind.

**73.2** Junction of Routes 80 and 100 in village of Foxon; bear left on Route 100. Road continues to follow foot of dip slope of Talcott flow; Holyoke flow makes Saltonstall Ridge across the valley. The sinuous course of the valley, ridges, and road reflects the sinuous strike of the strata.

**76.7** Pass under Route I-95.

**76.8** Traffic light at N end of bridge over railroad. Bear left over bridge onto High Street.

(**Optional Stop:** Turn right at light onto Laurel Road and proceed about 0.1 miles to overpass of Route I-95. Park and climb up into cut of I-95 through Talcott flow. There are pillows here, but they aren't as good as those at STOP 6. West end of cut on north side of road appears to be a dike (or dikes), but at one place halfway up the cut the dike rock seems to grade into the (pillow) lava. Return to traffic light and turn right (south) across bridge.)

**77.05** Traffic light (first of two). Turn left (east) on nameless street (in front of Old Stone Church).

**77.3** Stop sign at intersection. If optional STOP 11 is to be made, turn left (north) with care. Otherwise turn right (south) and pick up itinerary at 77.9.

**77.4** Traffic light at U.S. 1. Cross straight across U.S. 1 onto Estelle Street, which shortly bends around to the right.

**77.6** End of Estelle Street.

**STOP 11:** (Optional) Climb up to locked gate on small bridge over railroad and I-95. The big cut ahead is in the east-dipping Holyoke flow at the south end of Saltonstall Ridge; the reservoir to the right, Lake Saltonstall, fills the valley over the East Berlin formation. The Hampden flow forms the hills east of the lake. But the pass utilized by the railroad and highway lies on a cross-fault that offsets the strata 0.1 mile to the left; on our side of the bridge the flow forms the hill to our right, and the railroad cut below and to the right of the bridge is in the Shuttle Meadow formation (pieces of fresh-water limestone have been found there). Detailed mapping by John Sanders and students in the Yale Field Geology course, during the construction of I-95, showed that the fault zone is composite, being made of several nearly vertical parallel (east-west) faults, not all with left-handed offset. The little ridge between the railroad and I-95 is worth visiting as evidence of the composite faulting, if the gate key can be obtained from the New Haven Water Company.

(A precisely similar [or slightly larger] offset 0.35 miles to the south is utilized by U.S. 1. The cuts on the north side of U.S. 1 there show rather coarse or even finely conglomeratic sediments of the Shuttle Meadow formation cut by numerous small faults. The Holyoke flow does not crop out at the roadside but can be found around the corner on the dirt road on the west side of Lake Saltonstall at its outlet. On the other

hand, if one climbs along the hillside west from that point one remains in lava almost all the way to the west end of the cuts along the road below. The main cross-fault therefore lies only a few meters north of the roadcuts and is roughly parallel to them.)

(The site of the first iron smelter in Connecticut is believed to lie on the south side of Route 1, just below the dam of Lake Saltonstall. The ore used was presumably bog iron from swamps lying upstream from the smelter and now beneath the lake.)

Turn around and retrace route to U.S. 1 (77.8) and corner beyond.

77.9 Bear left (onto Hemingway Street, not marked).

78.0 Traffic light at intersection with Main Street. Proceed straight ahead. (To see road cuts along U.S. 1 mentioned in description of STOP 11, turn left here and proceed 0.3 miles east until street approaches U.S. 1. Stop there, look across U.S. 1, and observe the relations. Although a traffic light marks the intersection, it is quite dangerous to try to cross U.S. 1 on foot and also dangerous to work along the outcrops because the road curves around the outcrop and the cars, travelling at high speed, cannot see you until they are almost on you.)

78.1 Turn left on River Street and proceed to end.

78.4 Parking lot is behind Trolley Museum office on left.

**STOP 12:** East Haven-Branford Trolley Museum. We will board the trolleys, cross East River into town of Branford, and proceed south about 3/4 mile, following the west foot of Beacon Hill, formed by the Holyoke flow south of the offset at U.S. 1. Where the ridge ends and the trolley track turns left (east) and ends, disembark. You are now exactly on the eastern border fault again. Quartz veins in highly silicified gneiss (Light House gneiss, probably latest Precambrian) can be seen south of the tracks, and the Holyoke flow is exposed in a large quarry north of them. Coarse conglomerate can be found on the upper west slope of the ridge beneath the basalt at the summit; it is already considerably coarser than the rocks exposed along U.S. 1 less than a mile to the north.

When you are through looking, hail a trolley. On the trip back, don't fail to stop to see the main trolley museum.

End of field trip. To return to Kline Geology Laboratory, retrace itinerary to U.S. 1 (77.8 above) and turn left on it. In a little more than a mile, a left exit leads onto I-95. Follow I-95 to Exit 48 (right exit) onto I-91, then take Exit 3 (Trumbull Street). At traffic light at end of ramp, turn right on Orange Street, turn left at next traffic light onto Humphrey Street, and continue straight at next traffic light (Whitney Avenue) into Kline parking lot. KGL is to the left, at the south end of the parking lot. Distance from trolley museum is about 5 3/4 miles.

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**GEOLOGY IN THE VICINITY OF THE HODGES COMPLEX AND  
THE TYLER LAKE GRANITE, WEST TORRINGTON, CONNECTICUT**

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INTRODUCTION AND TRIP PURPOSE

Modern stratigraphic and structural studies of lower Paleozoic metamorphic rocks in the New England Appalachians have defined distinctions between miogeoclinal, transitional, and eugeoclinal tectonostratigraphic units (Hatch and others 1968, Cady 1969, Hall 1976, 1980, and Robinson and Hall 1980). Many workers in western Connecticut have noted the abundance of Proterozoic Y gneiss and autochthonous lower Paleozoic miogeoclinal cover rocks to the west of Cameron's Line and the abrupt eugeoclinal character of the Hartland Formation to the east (Agar 1927, Cameron 1951, Rodgers and others 1959, Hatch and Stanley 1973, and Merguerian 1983).

Recent interpretations suggest that Cameron's Line is an important deep-seated ductile fault in the medial Ordovician Taconic suture zone separating transitional rocks (Waramaug Formation) from essentially coeval eugeoclinal rocks (Hartland Formation).

Intruded along Cameron's Line in West Torrington, Connecticut, the Hodges mafic-ultramafic complex and the Tyler Lake granite are the products of medial Ordovician plutonism. In collisional orogens, the mechanisms of deep-seated suturing and plutonism remain enigmatic. This trip examines the structures developed at Cameron's Line, the geologic relations of the Hodges Complex and the Tyler Lake granite, and the metamorphic stratigraphy and structure of the lower Paleozoic Waramaug and Hartland wallrocks.

REGIONAL SETTING

The crystalline terrane of western Connecticut consists of a diverse assemblage of Proterozoic to lower Paleozoic metasedimentary and metaigneous rocks which can be traced from New York City northward into the Connecticut Valley-Gaspé synclinorium (Fig. 1). Separated by Cameron's Line, a major ductile shear zone in the New England Appalachians, two major tectonostratigraphic terranes compose the geologic framework of western Connecticut (Fig. 2).

Cameron's Line delimits the easternmost exposures of Proterozoic Y gneiss and overlying lower Paleozoic miogeoclinal rocks (Rodgers and others 1959, Hall 1980, Merguerian 1983). Together they represent deformed North American craton and overlying shelf deposits. Included in the western terrane are metamorphosed Cambrian to Ordovician allochthonous rocks such as the Waramaug Formation and the Hoosac Schist, deposited transitionally between shallow and deep water realms (discussion below).

The Hartland Formation comprises the eastern terrane and occurs to the east of Cameron's Line. The Hartland is a metamorphosed sequence of

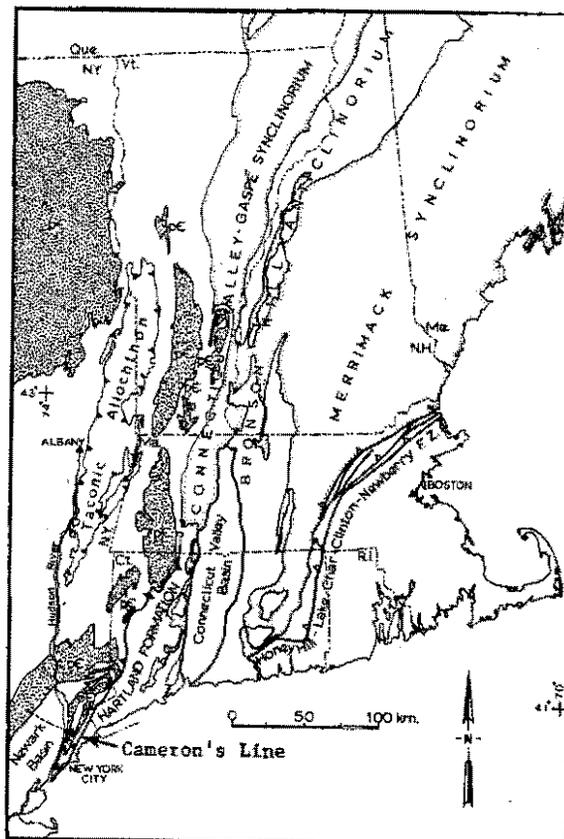


Figure 1 - Regional index map of New England (after Williams 1978).

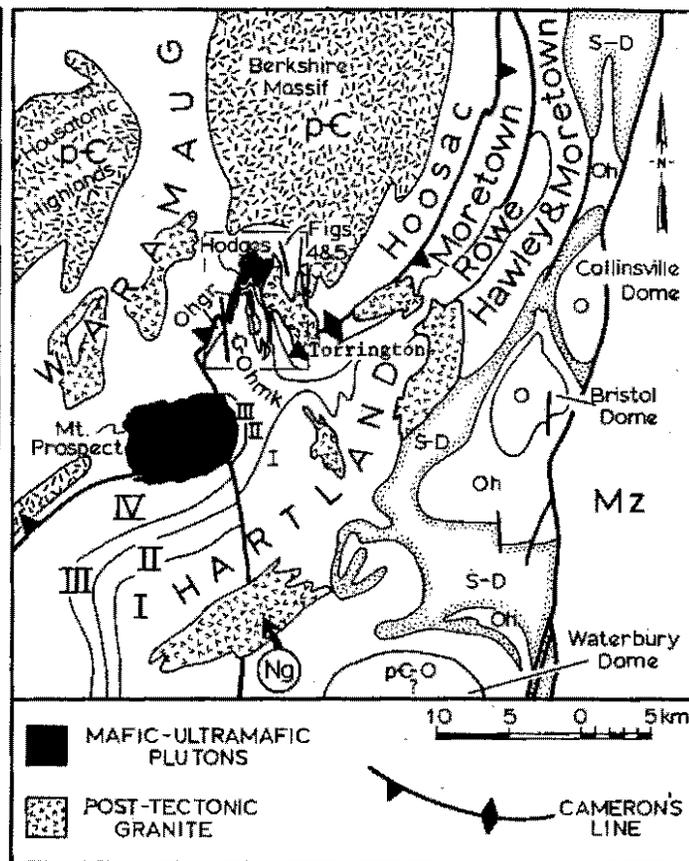


Figure 2 - Geological sketchmap showing the lithostratigraphic correlations between rocks of the West Torrington area and adjacent regions. Oh=Hawley Fm., Om=Moretown Fm., Ng=Nonewaug granite. After Hatch and Stanley (1973), Gates (1967), Merguerian (1983), and Rodgers and others (1959).

eugeoclinal rocks formerly deposited on oceanic crust. Judging by metamorphic minerals the western and eastern terranes were juxtaposed at depths of  $\approx 20$  km along Cameron's Line during lower Paleozoic times. The force behind such deep-seated deformation presumably resulted from a collision between a volcanic arc terrane and the passive margin of North America. Presently the arc terrane is exposed in the Bronson Hill Anticlinorium and its extension southward into central Connecticut (Fig. 1). Note the northwest to southeast stratigraphic variation from miogeoclinal to eugeosynclinal rocks from eastern New York State to central Connecticut (Fig. 3).

A number of lower Paleozoic calc-alkaline plutons occur both in the western and eastern terranes in southern New England. Near West Torrington the Hodges mafic-ultramafic complex and then the Tyler Lake granite were sequentially intruded across Cameron's Line (Merguerian 1977). They are interpreted as late syn-orogenic plutons due to their formerly elongate shapes and since the regional metamorphic fabrics related to the development of Cameron's Line in both the bounding Waramaug and Hartland Formations are contact metamorphosed. The recognition of significant medial Ordovician plutonism across Cameron's Line (Mose 1982, Mose and Nagel 1982, Merguerian and others 1984, Amenta and Mose 1985) establishes a Taconian or possibly older age for the formation of Cameron's Line and the syntectonic development of regional metamorphic fabrics in western Connecticut.

#### STRATIGRAPHY

The Waramaug Formation of Gates (1952) forms a belt up to 10 km wide from Torrington southward to New Milford, Connecticut (Fig. 2) where Clarke (1958) correlated the Waramaug and Manhattan Formations. Mapping by Jackson (1980), and Jackson and Hall (1982) near Kent, Connecticut, by Alavi (1975) near Bedford, New York, by Hall (1968a,b) in White Plains, New York, and by Merguerian and Baskerville (1986) in New York City support this correlation.

Regional correlations are shown in Fig. 3. The Waramaug has been redefined in northwestern Connecticut by Dana (1977) and Hall (pers. comm., 1981) where it is not as extensive as originally defined (Gates 1952). The Waramaug Formation is correlative and physically continuous between Connecticut and Massachusetts with the Late Proterozoic (?) to Cambrian Hoosac Schist (Hall 1971, 1976, Hatch and Stanley 1973, Merguerian 1977).

In the vicinity of West Torrington, the Waramaug (p6-Owg) crops out west, north, and northeast of the Hodges Complex (Fig. 4) and consist of a heterogeneous assemblage of rusty-, gray-, and locally maroon-weathering gneiss, mica schist, and granofels with subordinate amphibolite gneiss, amphibolite, and calc-silicate rocks (Stops 4, lunch stop and 5). Outcrops are massive and indistinctly layered with a nubby weathered surface due to resistant quartz+aluminosilicate segregations.

The Hartland Formation (Cameron 1951, Gates 1951, 1952) consists dominantly of aluminous metasedimentary and interlayered metavolcanic rocks. They are bound to the west by Cameron's Line and to the east,

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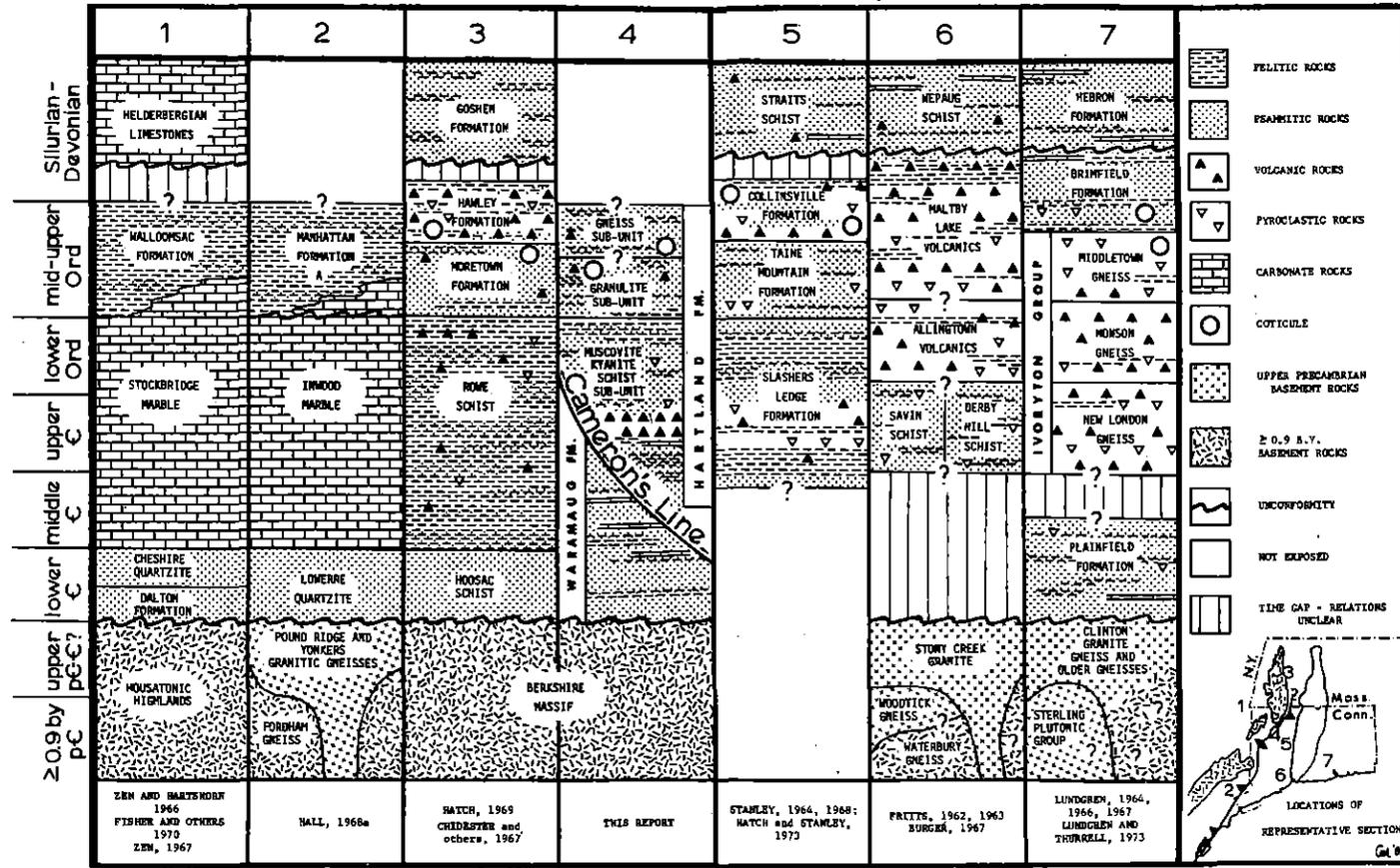


Figure 3 - Stratigraphic correlation chart for southern New England showing the interpreted protoliths of the lithologies depicted. Mixed symbols indicate relative, non-quantitative abundances of various pre-metamorphic lithologies.

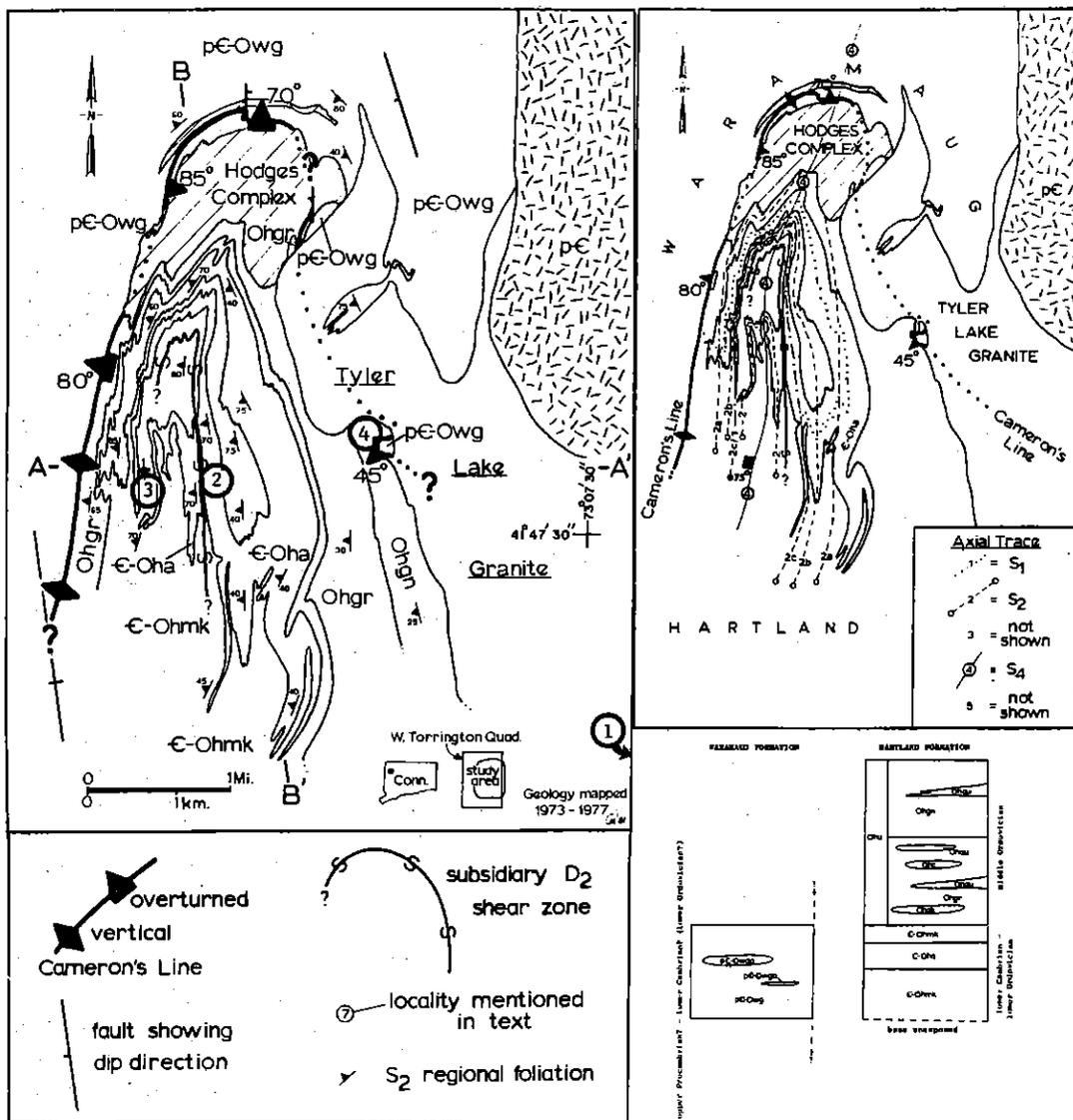


Figure 4 - Simplified geological map (showing Steps 2-4), axial surface map, and stratigraphic column of a part of the West Torrington quadrangle (after Merguerian 1977, 1983).

near the Collinsville, Bristol, and Waterbury domes (Fig. 2), are overlain by metamorphosed rocks of probable Silurian to Devonian age (Hatch and Stanley 1973). Rocks mapped as Hartland extend from New York City (Seyfert and Leveson 1969, Baskerville 1982, Merguerian and Baskerville 1986) through southeastern New York (Hall 1968a, Pelligrini 1977) and western Connecticut, northward to the Massachusetts state line (Hatch and Stanley 1973).

In western Connecticut the Hartland Formation consists of a thick sequence of interlayered muscovite schist, micaceous gneiss and granofels, amphibolite, and minor amounts of calc-silicate rock, serpentinite, and manganiferous garnet-quartz granofels (coticule). Hartland rocks are correlative with metamorphosed eugeoclinal Cambrian to Ordovician rocks found along strike in New England (Fig. 3). A pre-Silurian minimum depositional age for the sequence is indicated since the Hartland is overlain by Silurian and Devonian metamorphic rocks and the dominant regional foliation in the Hartland is truncated by the 383±5 m.y. Nonewaug granite (Mose and Nagel 1982).

In the vicinity of Torrington, the Hartland is subdivided into upper and lower members based on stratigraphic position (Figs. 3, 4). The lower member (6-Ohmk) consists of lustrous gray-weathering muscovitic schist typically containing large (up to 10 cm) porphyroblasts of garnet, biotite, staurolite, and kyanite (Stop 3). A texturally and mineralogically diverse assemblage of thick, laterally variable amphibolites (6-Oha) are interlayered within the lower member (Stop 2). The lower member grades, with some lensing, into the upper member.

The upper member (Stops 1, 4) consists of lustrous pin-striped muscovitic gneiss (Ohgn), well-layered quartzofeldspathic granofels and schist (Ohgr), amphibolite (Ohau), and subordinate quartzite, coticule, and calc-silicate rocks (Ohc) and lenses of muscovite-kyanite schist (Ohmk) (Figs. 3, 4). The upper and lower members of the Hartland Formation are correlative with the Rowe-Moretown-Hawley eugeoclinal sequence of western Massachusetts (Fig. 3).

Significant tectonic intercalation at Cameron's Line and intense regional isoclinal folding under amphibolite-grade metamorphic conditions create uncertainties in distinguishing between the Waramaug and Hartland Formations near Cameron's Line. Elsewhere, their unique lithologic character makes identification simple. Waramaug rocks are generally rusty- to gray-weathering, coarse- to medium-grained, gneissic, and granular to foliated with quartz-biotite and plagioclase the dominant minerals. Muscovite is typically present but not as abundant as in the Hartland. The Waramaug contains thin layers of amphibolite and amphibolitic gneiss which are typically granular and gray-green to black in color and rare tremolite-quartz calc-silicate layers.

In contrast, the Hartland rocks are gray-weathering and well-layered, fine- to coarse-grained, and typically schistose with interlayers of granofels, amphibolite, and rare coticule. The rocks are very rich in muscovite and quartz and, to a lesser extent, plagioclase and contain thin to thick layers of greenish amphibolite.

The absolute ages of the Waramaug and Hartland Formations are unknown but they are basically considered time-stratigraphic correlatives by most workers (Hall 1976). They represent dominantly transitional slope-rise (Waramaug) and adjacent oceanic (Hartland) sequences deposited in the lower Paleozoic near the North American shelf and were subsequently deformed into metamorphic rocks and juxtaposed at depth along Cameron's Line.

#### THE HODGES COMPLEX AND THE TYLER LAKE GRANITE

Metamorphosed mafic and ultramafic rocks of the Hodges Complex underlie a 2.5 km<sup>2</sup> area (Figs. 4, 5). The complex is a steep-walled, folded, mushroom-shaped pluton with a hornblende gabbro core and dioritic chilled margin. A stock-like central intrusion and many smaller separated masses of pyroxenite and hornblendite crosscut the main gabbro-diorite pluton as well as foliated amphibolites of the Hartland Formation stretching to the south. The pluton is in direct contact with both the Waramaug and Hartland Formations and is surrounded by a narrow contact aureole.

All rocks of the Hodges Complex have been metamorphosed with recrystallization and overprinting of original igneous textures. Relict olivine, enstatite, hypersthene, augite, and hornblende are corroded and replaced by tremolite, actinolite, anthophyllite, cummingtonite, hornblende, magnesian chlorite, calcite, talc, and serpentine minerals. Gabbros and diorites are relatively unaltered although recrystallization of plagioclase and replacement of hornblende by biotite and chlorite has occurred.

The Hodges is subdivided into three mappable metaigneous rock units - Pp = pyroxenite, hornblendite, Hg = gabbro, and Di = diorite. The ultramafic rocks (Pp) are typically highly-altered, medium-grained to pegmatitic, dense and deeply iron-stained, silver-green to dark-green to black hornblende orthopyroxenite, biotite-tremolite-orthopyroxenite, orthopyroxene hornblendite, hornblendite, and biotite hornblendite (Stops 6, 7).

The main gabbroic mass of the Hodges Complex (Hg) is composed of medium- to very coarse-grained, dark gray-weathering, hornblende-plagioclase-biotite±quartz gabbro (Stop 6). Labradorite (An<sub>50-55</sub>) is generally clouded with oscillatory zoning and is about equal to the hornblende content. The hornblende contains pyroxene ghosts defined by opaques suggesting that pyroxene was an important mineral phase prior to metamorphism.

The dioritic rocks (Di) are by far the most variable in texture and mineralogy (Stops 5-7). They are greenish to black and white, poor- to well-foliated, fine- to medium-grained, banded hornblende-plagioclase-biotite±quartz diorites. Alternating layers of subhedral hornblende together with euhedral to subhedral laths of plagioclase together define an igneous flow layering. The diorites form the flow-layered chilled margin of the main gabbroic mass of the Hodges.

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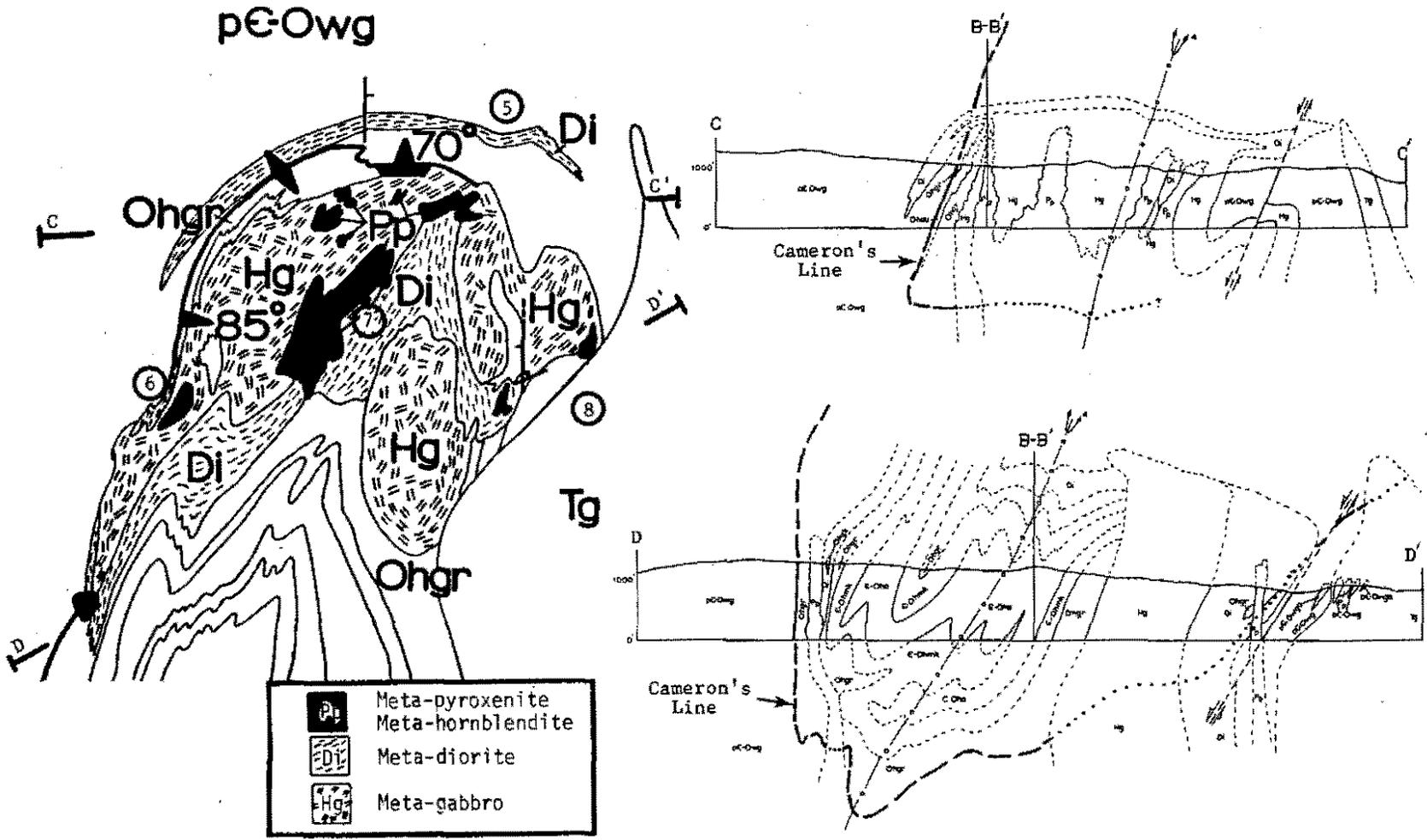


Figure 5 - Geologic map and sections of the Hodges Complex showing Steps 5-8.

The Tyler Lake granite (Stop 8) is an elongate pluton initially described by Gates and Christensen (1965) as the eastern mass of the Tyler Lake granite. It intrudes across Cameron's Line and includes xenoliths of the Hodges rocks at their contact zone.

The  $S_1 + S_2$  regional foliation exerted a strong control on the original geometric form of the Hodges and Tyler Lake intrusives generating sheetlike, rather than equidimensional forms. As discussed later, minerals of the Hodges contact aureole postdate the  $S_2$  regional foliation in the Waramaug and Hartland wallrocks. Post-intrusive folding has deformed Cameron's Line and the plutons, causing abundant metamorphic alteration of original igneous textures (Table 1).

#### STRUCTURAL GEOLOGY, INTRUSIVE RELATIONSHIPS, AND METAMORPHISM

The wallrocks of the Hodges Complex and the Tyler Lake granite have experienced a complicated Phanerozoic structural history that began with two phases of isoclinal folding ( $F_1$  and  $F_2$ ) yielding two subparallel regional foliations ( $S_1$  and  $S_2$ ).  $F_1$  folds are rare and usually developed in amphibolites which were less ductile than the surrounding schistose rocks during subsequent deformation (Fig. 6a). However, an  $S_1$  foliation and parallel compositional layering is commonly deformed by  $F_2$  folds in both the Waramaug and Hartland Formations. The  $D_1$  and  $D_2$  events are similar in orientation, style, and metamorphism (amphibolite grade) and are considered progressive. They mark the initial prograde metamorphic pulse ( $M_1$  in Table 1) that culminated during the formation of Cameron's Line.

Cameron's Line is a 15-90 m wide zone of intense localized isoclinal  $F_2$  folds with limbs sheared parallel to  $S_2$ , transposition of  $S_1$  fabrics, and regional truncation of Hartland subunits. The synmetamorphic shear zone includes layers of mylonitic amphibolite intercalated with both Waramaug and Hartland rocks and, locally, deformed slivers of serpentinite (Stop 4). Away from Cameron's Line,  $D_2$  resulted in the development of a penetrative regional foliation ( $S_2$ ) in the Waramaug and Hartland Formations. While it is unclear whether motion along Cameron's Line initiated during  $D_1$ , the regional parallelism of  $S_2$  axial surfaces and the trace of development of Cameron's Line in West Torrington (Fig. 4) strongly suggest that the development of Cameron's Line and  $S_2$  in the wallrocks are essentially coeval. A subsidiary  $D_2$  shear zone, marked by mylonitic amphibolite (Stop 2) and a soapstone-talc body (optional Stop 2a) are also developed within the Hartland.

A secondary regional metamorphic pulse ( $M_2$  in Table 1) occurred after the juxtaposition of the Waramaug and Hartland Formations since porphyroblasts of garnet, staurolite, and kyanite overgrow the  $S_2$  foliation. The  $M_2$  metamorphic pulse was reached after  $D_2$  but before  $D_4$  since  $M_2$  porphyroblasts are deformed by  $S_4$  cleavage. The contact aureole of the Hodges Complex also overprints the  $S_2$  foliation and it is likely that the intrusion of the Hodges was synchronous with the regional  $M_2$  event as shown in Table 1. Both the  $M_1$  metamorphic pulse and the development of the Hodges contact aureole are pre-medial Ordovician events based on a  $466 \pm 12$  m.y. Rb/Sr age on the Tyler Lake granite reported by Mergerian and others (1984). It is possible that the  $M_2$  event is also

DEFORMATIONAL EVENT	LINEAR FEATURES	PLANAR FEATURES	IGNEOUS ACTIVITY	METAMORPHISM
D <sub>1</sub>	F <sub>1</sub> isoclinal folds of compositional layering. L <sub>1</sub> quartz ribbing in gneisses and schists. Hornblende lineation in amphibolites.	S <sub>1</sub> gneissic layering in gneisses or hornblende-plagioclase foliation in amphibolites. Generally not recognized in schists.		Amphibolite-grade M <sub>1</sub>
D <sub>2</sub>	F <sub>2</sub> penetrative isoclinal folds of early S <sub>1</sub> structures and compositional layering. L <sub>2</sub> mineral streaking in schists and gneisses.	S <sub>2</sub> regional foliation composed of oriented phyllosilicates+kyanite or sillimanite developed axial planar to F <sub>2</sub> folds.	HODGES COMPLEX TYLER LAKE GRANITE 466±12 m.y.	Amphibolite-grade M <sub>2</sub>
D <sub>3</sub>	F <sub>3</sub> shallow SW to NW plunging, open to tight, crenulate folds of the S <sub>2</sub> regional foliation. L <sub>3</sub> intersection lineation in massive rocks; crinkle axis in micaceous rocks.	S <sub>3</sub> crenulation or slip cleavage developed axial planar to F <sub>3</sub> folds. Oriented NW to WSW with shallow dips.	PEGMATITES.	
D <sub>4</sub>	F <sub>4</sub> steep SW plunging dextral synformal folds of the S <sub>2</sub> regional foliation	S <sub>4</sub> crenulation cleavage, slip cleavage, or spaced schistosity developed axial planar to F <sub>4</sub> folds Orientation - N20°E, 75°NW.		Biotite-grade M <sub>3</sub> (retrograde)
D <sub>5</sub>	F <sub>5</sub> open folds and warps with variable hingelines. L <sub>5</sub> intersection lineation.	S <sub>5</sub> slip cleavage and rock cleavage axial planar to F <sub>5</sub> folds oriented NW to W with variable dip.		continued retrograde

G2-10

Table 1 - Linear and planar structural features and chronology of folding, igneous activity, and metamorphism in Torrington, Connecticut area.

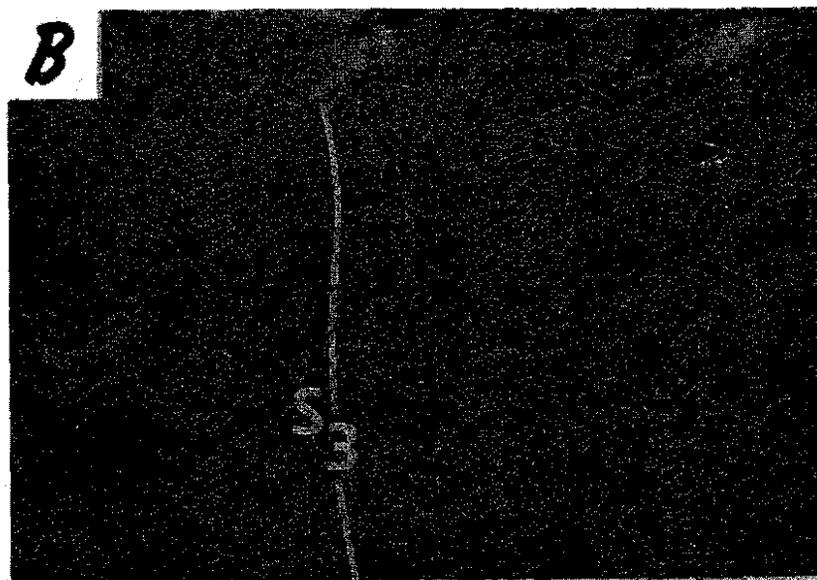
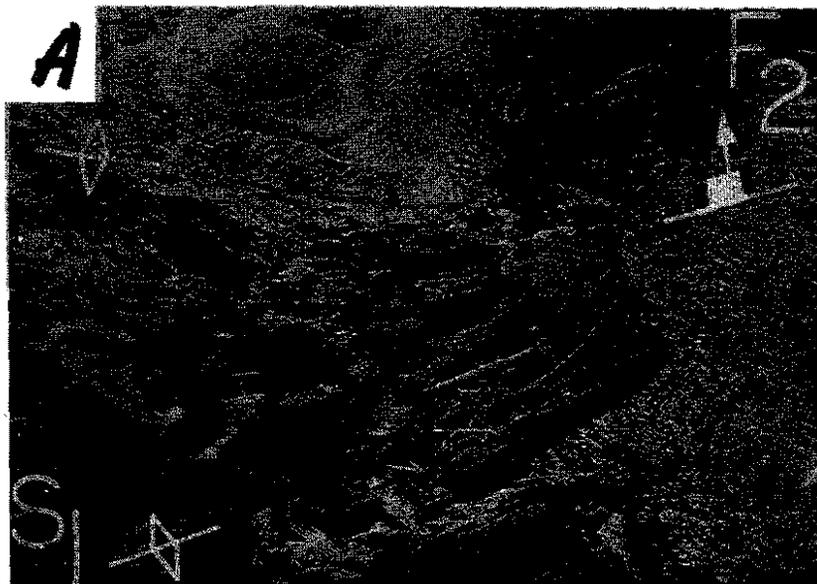


Figure 6 - a) Typical  $F_2$  isoclinal fold of foliated ( $S_1$ ) amphibolite away from Cameron's Line.  $S_1$  can be traced into the axial surfaces of upright isoclinal  $F_1$  folds. b) Typical  $F_3$  folds in Hartland gneiss (Ohgn) near the contact with the Tyler Lake granite. c)  $F_4$  fold in Hartland amphibolite  $\epsilon$ -Oha showing axial planar slip cleavage ( $S_4$ ). Hammer is 45 cm long.

of Ordovician vintage despite the fact that most workers in Western Connecticut attribute the growth of large post-regional foliation porphyroblasts to Acadian (middle Devonian) Barrovian metamorphism documented in Massachusetts (Hatch 1975, Stanley 1975, Robinson and Hall, 1980).

Open to tight, crenulate  $F_3$  folds occur dominantly in the vicinity of the plutons (Fig. 6b). Often their axial surface cleavages are parallel to the margins of the plutons and they have little regional effect on the map pattern. They are interpreted as syn-intrusive folds; their axial surfaces are not shown in figure 4.

Near West Torrington the  $S_1 + S_2$  foliations, Cameron's Line, and the Hodges and Tyler Lake plutons were strongly deformed by dextral  $F_4$  folds (Fig. 6c) and cut by an associated axial planar spaced schistosity.  $S_4$  is characterized by the growth of idioblastic biotite and hornblende and recrystallized quartz, by parting in  $M_2$  garnet, staurolite, and kyanite porphyroblasts, and by brittle deformation of plagioclase twin lamellae. The  $S_4$  schistosity crosscuts the large  $M_2$  staurolite±kyanite±garnet porphyroblasts as well as the  $S_1 + S_2 = M_1$  regional foliation (Table 1). Metamorphism ( $M_3$ ) during the  $D_4$  event fosters retrograde biotite and amphibole recrystallization. In addition, the  $D_4$  event caused widespread metamorphic recrystallization, serpentization, and chloritization in the Hodges Complex and recrystallization and domainal shearing in the Tyler Lake granite (Stops 5-8).

A fifth, and possibly sixth, deformation is suggested by the warping of the  $S_4$  axial surface trace (Fig. 4) and by local open to crenulate folds with variable plunges and shallow NE and NW to W-trending axial surfaces. This deformation is low grade and is marked by recrystallized quartz and chlorite±white mica. These crosscutting structural, metamorphic, and intrusive relations are summarized in Table 1 and discussed in greater detail in Merguerian (1977, 1983). The  $D_x$ ,  $F_x$ ,  $S_x$ , and  $M_x$  nomenclature (to denote deformational event, fold generation, axial surface fabric, and metamorphic event, respectively) will be utilized in later field descriptions.

#### STEREOGRAMS AND STRUCTURE SECTIONS

Stereograms of the major structural features described above are shown in figure 7. Stereogram 1 shows poles to  $S_2$  in both the Waramaug

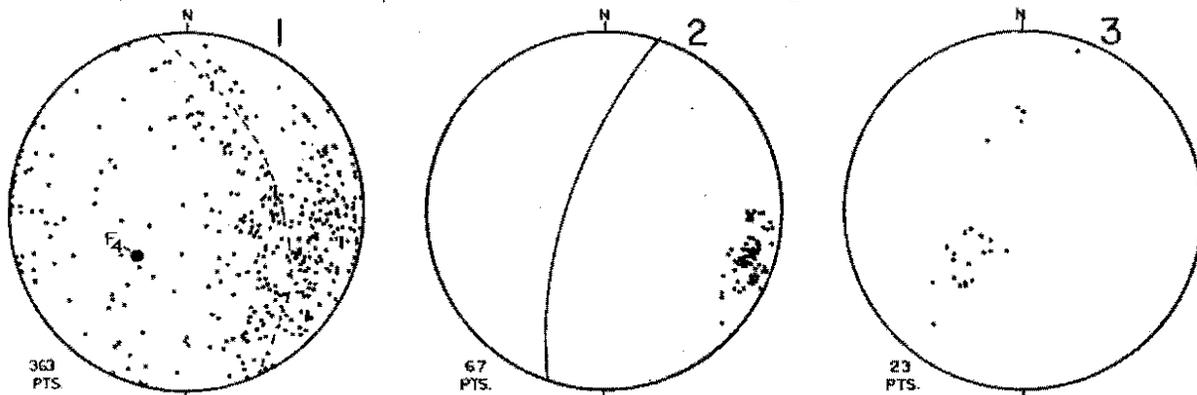


Figure 7 - Stereograms of structural elements.

and Hartland Formations. The wide scatter of poles distributed about a NW-SE girdle indicates the presence of post-D<sub>2</sub> deformation. Poles to S<sub>4</sub> (Stereogram 2) and F<sub>4</sub> fold axes and L<sub>4</sub> intersection lineations (Stereogram 3) shows a consistent trend for S<sub>4</sub>  $\cong$  N19°E, 72°NW and F<sub>4</sub>  $\cong$  S50°W @ 60°. Clearly, the girdle distribution of S<sub>2</sub> poles is largely the result of F<sub>4</sub> folding. Some scatter due to local F<sub>3</sub> and F<sub>5+</sub> folds may have also occurred.

Sections in figure 8 are drawn from map data and axial surface traces shown figure 4 but the exact configuration of F<sub>1</sub> closures in the subsurface is hypothetical due to extensive D<sub>2</sub> transposition. The major obvious structure in section A-A' is a dextral F<sub>4</sub> synform with a steep

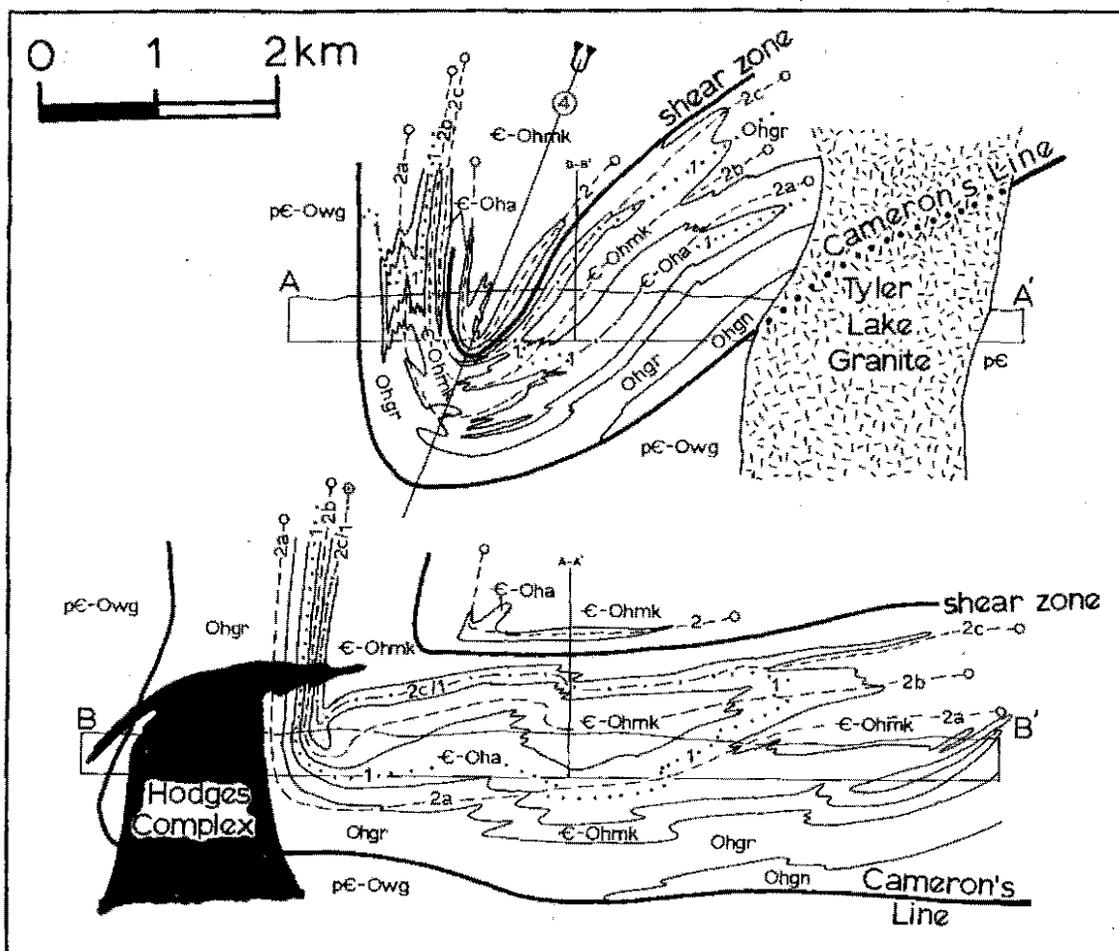


Figure 8 - Geologic structure sections. Section lines are shown in figure 4. No vertical exaggeration.

western limb (vertical to locally overturned toward the east) and a shallow west-dipping eastern limb. The interference of F<sub>1</sub> and F<sub>2</sub> folds yields a complex interdigitating map pattern of Harland subunits. The section shows superposition of F<sub>4</sub> on the older structures, folding of Cameron's Line and the subsidiary D<sub>2</sub> shear zone, the truncation of Hartland subunit Ohgn against Cameron's Line, and the cross-cutting relationship of the Tyler Lake granite.

Section B-B' shows a north-south view roughly parallel to the trace of S<sub>4</sub>. Again, the complicated fold geometry of the Hartland subunits, truncation of Hartland subunit Ohgn, and crosscutting relationship of the Hodges Complex is indicated.

#### ACKNOWLEDGMENTS

The main phase of field work for this study was performed in 1973-1977 but subsequent study and remapping has continued to the present. Financial and technical support was provided by the Connecticut State Geological and Natural History Survey (Grants 81-506 and 82-506), Hofstra University, and Duke Geological Laboratory. I gratefully acknowledge assistance in the field by G.V. Bennett, L. Bean, and P. LaJuke and the help and support of Dr. Nicholas M. Ratcliffe, Ray Wadhams, and John A. Carter during the early stages of this study.

ROAD LOG

From New Haven, Connecticut, travel west on State route 34 for roughly 6 miles to the intersection with State route 8 near Derby. Take route 8 northward through Waterbury and Thomaston for roughly 32 miles to exit 42 (Litchfield-Harwinton). Bear right (east) on route 118 at the end of the exit ramp and park in the commuter parking lot. Plan to arrive at the assembly point (Stop 1) at 9:00 AM sharp. We will pause for lunch between stops 4 and 5 and there will be places to pick up food if necessary. All of the stops are within the West Torrington 7-1/2 minute quadrangle except for Stop 1 (Torrington quadrangle).

Stop 1 - Hartland Formation (upper member) granofels, schist, and amphibolite.

A convenient place to initiate today's fieldtrip, the outcrops forming the roadcut across from the commuter lot were originally described by Martin (1970). Here, 2-15 cm-scale very well-layered muscovite-biotite-plagioclase-quartz-(hornblende)-(garnet) granofels occurs with interlayered schist of similar mineralogy. The major minerals are listed in order of decreasing abundance; those in parentheses are not found in all exposures. The abundance of muscovite in the granofels and schist creates a lustrous sheen from foliation surfaces reflecting sunlight. A 2 m thick layer of hornblende-plagioclase-biotite-epidote-quartz-(garnet) amphibolite is exposed on the south-facing portion of the outcrop. The pervasive interlayering of granofels and schist, high muscovite and plagioclase content, and presence of amphibolite suggests that protoliths of these rocks were volcanoclastic graywackes and interlayered shale with subordinate basalt flows. The Hartland upper member is similar to and correlative with the Moretown Formation of western Massachusetts (Fig. 3).

The dominant layering is parallel to the composite  $S_1 + S_2$  regional foliation, all striking roughly  $N65^\circ E$  with dips of  $60^\circ NW$ . The  $S_1 + S_2$  foliation is deformed by crenulate  $F_3$  folds with axial surfaces oriented  $N30^\circ E, 26^\circ SE$ . The  $F_3$  hingelines are expressed as  $L_3$  crinkle-axis lineations in highly micaceous layers and as  $L_3$  intersection lineations in more massive granofels. The  $F_3$  and  $L_3$  elements trend  $N55^\circ E$  and plunge  $9^\circ$ . Note the upright warping of  $S_3$  axial surface traces and the decrease in wavelength of  $F_3$  folds in mica-rich interlayers. A 2 m thick pegmatite intrudes across the  $S_3$  axial surfaces, locally rotating  $F_3$  folds and older fabrics. Note the  $F_4$  "z" folds with  $N20^\circ E, 84^\circ NW$  axial-planar slip cleavage.

Near the eastern end of the roadcut,  $L_2$  lineations are deformed by subhorizontal  $F_3$  folds and overprinted by  $L_3$  lineations. The associated  $S_3$  axial surfaces ( $N40^\circ E, 20^\circ SE$ ) are warped by late  $F_4$  crenulations with axial surfaces oriented  $\approx N25^\circ E, 70^\circ NW$ . Broad arching by later  $F_{5+}$  is also evident.

On the west-facing portion of the roadcut adjacent to the north-bound entrance ramp for route 8,  $F_2$  intrafolial folds occur in thinly layered granofels. The  $S_2$  axial surface strikes  $N55^\circ E$  and dips  $56^\circ NW$  and  $F_2$  hingelines are subhorizontal, trending  $N55^\circ E-S55^\circ W$ . The  $F_2$  folds

deform a pre-existing  $S_1$  mica foliation. Many amphibolite layers are exposed to the north along the roadcut.

### MILEAGE

<u>Total</u>	<u>Interval</u>	
0.0	0.0	Exit the commuter lot and take the northbound ramp for route 8 toward Torrington. Note the upright $F_2$ folds in outcrops 0.6 miles from the starting point. At exit 44 (Routes 4 and 202 - Downtown Torrington) follow the exit ramp to the traffic signal.
3.3	3.3	Make a left travelling westward on route 202 (East Main Street) past three traffic lights.
3.9	0.6	At the fourth traffic light bear right ( <u>across</u> Main Street) up the hill onto Water Street. Follow Water Street past the railroad tracks to the traffic light.
4.3	0.4	Turn left onto Church Street. Drive over the Naugatuck River which separates Proterozoic Y gneiss of the Berkshire massif on the east from the Tyler Lake granite on the west. Follow Church Street to the small traffic triangle.
4.6	0.3	Turn left (west) driving uphill onto Highland Avenue. For the next 0.9 miles, the Tyler Lake granite crops out in wooded areas away from the road. Pass Allen Road on the right (5.2 mi.) and Stop 4 near Patterson Pond (5.6 mi.). Continue west, now driving on Ohgr and pass the radio towers to the right (6.3 mi.) which essentially mark the contact between the upper and lower members of the Hartland.
6.4	1.8	Pass Westside Road on the right and Rossi Road on the left and continue uphill to the massive outcrops on either side of the road (here known as Soapstone Hill Road!).
6.8	0.4	Park in the bend of the road just past the outcrops.

### Stop 2 - Hartland Formation (lower member) amphibolite and subsidiary $D_2$ shear zone.

The roadside outcrops consist of fine- to medium-grained, dark-green hornblende-plagioclase-biotite-(quartz)-(epidote)-(chlorite)-(garnet) amphibolite with lineated prismatic hornblende. Elliptical quartz segregations up to 4 cm thick lie within the  $S_2$  foliation. Elsewhere, 1-2 m thick felsic granofels#hornblende, #biotite, #chlorite are interlayered with the amphibolite and muscovitic schist.

The  $S_2$  foliation strikes  $N15^\circ W$  and dips  $67^\circ SW$  with a prominent  $L_2$  hornblende lineation trending  $N80^\circ W$  and plunging  $65^\circ$ .  $S_1$  is essentially coplanar with  $S_2$  due to intense transposition and overprinting during  $D_2$ .  $S_2$  and  $L_2$  parallel the axial surfaces and hingelines, respectively, of rootless  $F_2$  isoclinal folds exposed on gently northeast-dipping joint faces. The  $S_1$  foliation, composed of hornblende and plagioclase, is locally preserved in  $F_2$  hinges.

The amphibolite tends to break into wedge-shaped pieces due to the coplanar  $S_1 + S_2$  foliation and an oblique  $S_4$  spaced cleavage striking N-S and dipping  $75^\circ W$ . Oriented samples show  $S_4$  defined by idioblastic biotite, the product of  $M_3$  metamorphism.

Walk 110 m west on Soapstone Hill Road where amphibolite outcrops exhibit  $S_2$  mylonitic layering ( $N14^\circ W$ ,  $82^\circ SW$ ). Between these outcrops the muscovite schist is phyllonitic and thin (Figs. 4, 8). The mylonitic textures may mark a subsidiary  $D_2$  shear zone which imbricates the Hartland amphibolite. Alternatively, shearing could simply be the result of ductility contrasts developed across the amphibolite-schist contact.

#### Stop 2a (Optional) - Soapstone Quarry

Walk north on a dirt trail immediately west of the parking area for Stop 2. Along the way, ridges are composed of amphibolite and the intervening valleys are underlain by muscovitic schist. Roughly 700 m north the trail ends at a 90 m long by 20 m wide pit of a former soapstone quarry. (They don't call it Soapstone Hill Road for nothing, you know!)

The excavation, which is oriented parallel to  $S_2$  in the bounding muscovite-chlorite schist, produced commercial quantities of soapstone. Blocks from the tailings pile include talc-tremolite schist, chlorite schist, and very-coarse grained amphibolite rich in opaque minerals.

The quarry is on-strike with mylonitic amphibolite to the south. The elongate shape parallel to  $S_2$  and foliated nature of the altered serpentinite body suggests that the soapstone-talc body represents ultramafic rock deformed during  $D_2$  and possibly  $D_1$ . It may mark a syn-tectonic ultramafic intrusion (Gates and Christensen 1965) or a small sliver of ophiolite.

6.8    0.0    Continue west on Soapstone Hill Road and pull into a large clearing to the left.

7.3    0.5    Outcrops for Stop 3 are in the woods north of the road.

#### Stop 3 - Hartland Formation (lower member) muscovite-kyanite-staurolite schist.

The lower member Hartland schist crops out less than 50 m north of the road. The rocks are highly lustrous, gray-weathering, medium- to coarse-grained, quartz-muscovite-plagioclase-biotite-opaque-(garnet)-(chlorite)-(apatite) schists often containing 1-10 cm porphyroblasts of

kyanite, staurolite, and garnet, and more rarely plagioclase and biotite. Quartz and muscovite are roughly equal in proportion, together composing more than half the rock. Deeply-eroded outcrops have a knotted appearance due to differentially weathered porphyroblasts. Granular, clear- to smoky-gray quartz pods are conspicuous and occur flattened into  $S_2$ . The lower member becomes markedly hornblende, chlorite, and/or biotite-rich near amphibolite contacts.

The muscovite schist, amphibolite, and rare felsic granofels of the lower member are probably derived from metamorphosed pelitic sediments with interlayered basalt and rare volcanoclastic layers. The rocks are correlative with the Rowe Schist of western Massachusetts (Fig. 3).

The large, non-oriented porphyroblasts of kyanite, staurolite, and garnet overgrow the  $S_2$  foliation and represent the  $M_2$  metamorphism in Table 1. Kyanite tends to occur mimetically within  $S_2$ . Staurolite tends to form spongy porphyroblasts, sometimes twinned, protruding randomly from the schist.

7.3 0.0 Backtrack east on Soapstone Hill Road.

9.0 1.7 Make a sharp right onto a partly hidden dirt road just past Patterson Pond. Pull up as far as possible.

#### Stop 4 - Cameron's Line and dismembered ophiolite.

The mylonitic amphibolite (Ohau) outcrop on the dirt road occurs within Cameron's Line, a 90 m zone consisting of highly-sheared, tectonically intercalated lithologies of the Hartland (upper member) and the Waramaug Formations (Fig. 9). In the amphibolite, an  $S_2$  mylonitic foliation ( $N85^\circ W, 70^\circ NE$ ) is parallel to the axial surfaces of  $F_2$  folds with sheared out limbs plunging  $40^\circ$  into  $N75^\circ W$  (Fig. 10a). An  $S_1$  foliation composed of aligned hornblende occurs in the  $F_2$  hinge areas. A specimen collected from this outcrop after blasting in 1973 shows an  $F_1$  isocline refolded by  $F_2$  with significant shearing and recrystallization parallel to  $S_2$  (Fig. 10b).  $F_3$  folds with subhorizontal axial surfaces warp  $S_2$ . The mylonitic amphibolite is interlayered with lustrous muscovitic gneiss (Ohgn).

The Hartland crops out to the north, west and south (Figs. 4, 9). Follow the dirt road south, stopping first to examine the Hartland gneiss outcrops on the subdued knob to the east. Here,  $S_2$  dips shallowly toward the west due to  $F_3$  folding. About 120 m to the south note mylonitic amphibolite in the woods to the east.  $S_2$  strikes  $N40^\circ E$  and dips  $37^\circ NW$ ;  $L_2$  trends  $S80^\circ W$  and plunges  $27^\circ$ . The Tyler Lake granite crops out farther south and east (Fig. 9).

Walk back to Highland Avenue and walk east roughly 100 m. The first outcrop to the right (before the creek) consists of massive, rusty-weathering quartz-plagioclase-biotite-sillimanite-muscovite-garnet-chlorite-tourmaline gneiss of the Waramaug Formation. The laminated  $S_2$  mylonitic foliation strikes  $N65^\circ W$  and dips  $32^\circ SW$ . The Waramaug

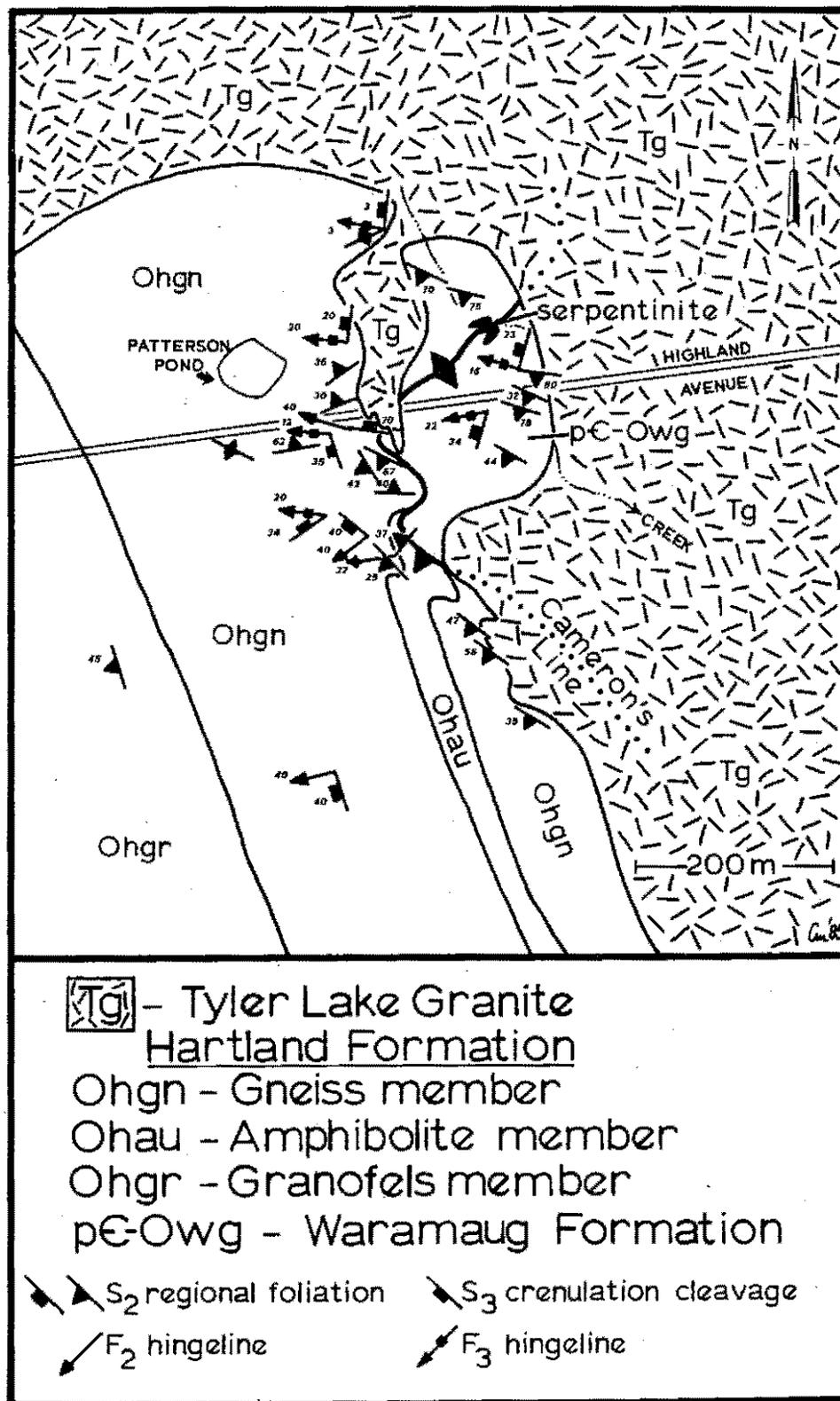


Figure 9 - Geological sketchmap in the vicinity of Stop 4.

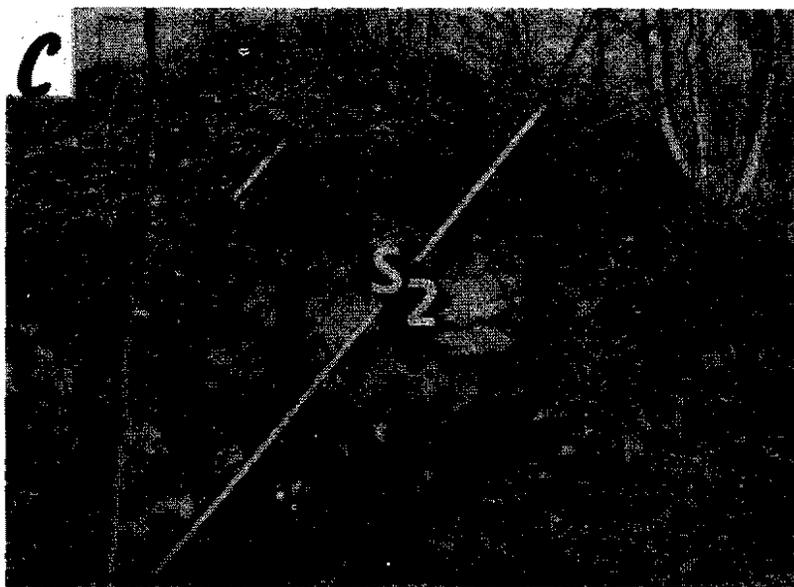
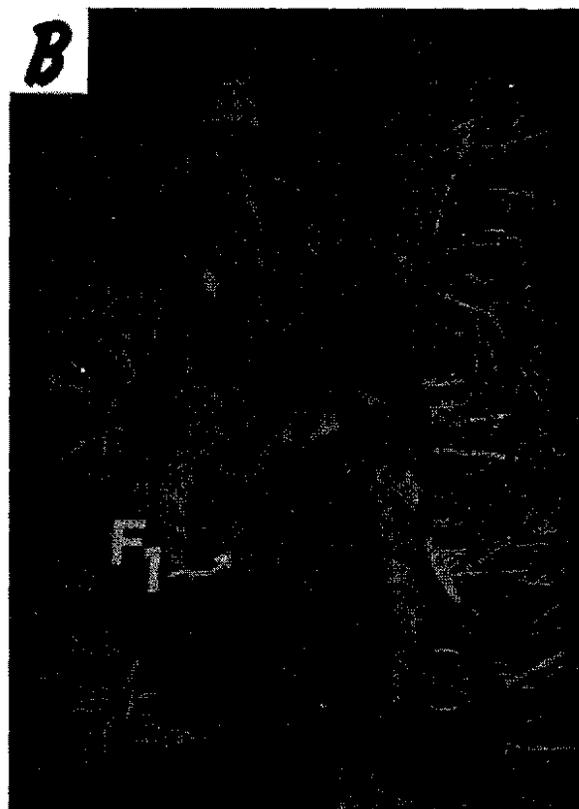
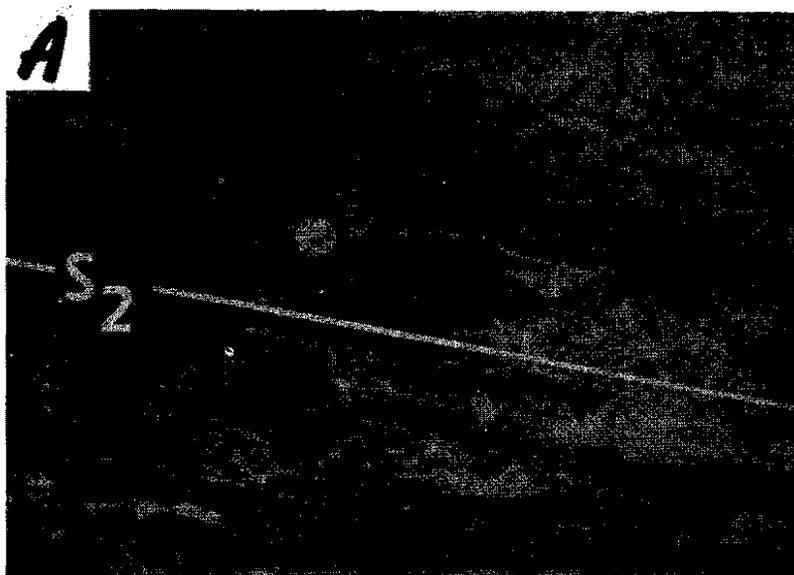


Figure 10 - a)  $F_2$  folds in mylonitic amphibolite (Ohau) near Cameron's Line with limbs sheared parallel to  $S_2$ . b)  $F_1$  isoclinal fold refolded by  $F_2$  and sheared parallel to  $S_2$ . c) Isoclinally folded serpentinite occurring at Cameron's Line with a steep  $S_2$  axial surface trace. The serpentinite is compositionally zoned and highly altered.

crops out to the south and north, thus Cameron's Line occurs between these outcrops and the cars.

Walking north, adjacent to the small creek, note the overturned  $F_3$  fold with pegmatite intruded along  $S_3$ . The small hill to the west is underlain by southwest-dipping Waramaug gneiss. Trace the creek to where a 10 m isoclinally folded serpentinite body separates Waramaug rocks to the southeast from Hartland rocks to the west and northwest. Folded by  $F_2$  folds (Fig. 10c), the serpentinite is zoned and highly-altered containing relict olivine and orthopyroxene. The zoning (compositional? or tectonic?) is due to relative enrichment of greenish, intergrown cummingtonite and tremolite in the upper part of the body compared to the dense, black serpentine- and anthophyllite-enriched lower part.

The body is distinct in mineralogy and texture from ultramafic rocks of the Hodges Complex. The overall eugeoclinal nature of the Hartland and the  $D_2$  (and possibly  $D_1$ ) deformation expressed in the serpentinite suggests that the body represents dismembered ophiolite (Merguerian 1979).

- |      |     |  |
|------|-----|--|
| 9.0  | 0.0 | Drive east on Highland Avenue back to the turnoff from Church Street.  |
| 10.0 | 1.0 | At the stop sign, turn left onto Riverside Avenue. Note the exposures of Waramaug amphibolite (p6-Owga) at the entrance to Charlene Susan Besse Park (10.9 mi.). Outcrops of Waramaug occur on the hilltop roughly 500 m due south. Continue north on Riverside Avenue to the intersection with State route 4. |
| 11.3 | 1.3 | Turn right onto route 4 and drive 0.25 miles east to Scarpelli's Drive-In. Those without lunch can pick something up here but we'll lunch at a nearby picturesque spot. With supplies in hand drive west on route 4 to the traffic light.  |
| 11.8 | 0.5 | Turn right onto route 272 toward Wrightville.  |
| 12.6 | 0.8 | Turn right onto Brass Mill Dam Road.   |
| 12.9 | 0.3 | <u>LUNCH STOP.</u> Park in the wide area south of the dam. There are nice places to lunch to the northwest.  |

#### Lunchstop

Outcrops of the Waramaug occur in the woods to the west and excellent exposures occur in the spillway north of the cars. Here, the Waramaug and interlayered amphibolite are deformed into 15 m amplitude  $F_2$  folds with  $N42^\circ E$ ,  $57^\circ NW$  axial surfaces and hingelines trending  $S40^\circ W$  at  $15^\circ$ .  $F_3$  crenulate "z" folds trend W at  $37^\circ$  with  $N15^\circ W$ ,  $52^\circ SW$  axial surface cleavage. Stillwater Pond may be fault controlled as abundant  $N20^\circ W$ ,  $75^\circ SW$  closely-spaced fractures cut the Waramaug exposures.

- 12.9 0.0 Backtrack on Brass Mill Dam Road.
- 13.2 0.3 Drive across route 272 uphill onto Hodges Hill Road.
- 13.7 0.5 At the stop sign, turn right onto University Drive (Town Farm Road).
- 14.1 0.4 Turn left onto John Brown Road and continue to dirt trail turnoff to the left.
- 14.3 0.2 Pull in as close as possible - we're at Stop 5!

Stop 5 - Waramaug Formation and contact relations of Hodges diorite sill

Walk roughly 60 m south on the dirt trail to an outcrop of massive but internally laminated, gray-weathering quartz-plagioclase-biotite-sillimanite-muscovite-garnet gneiss. Similar to the Waramaug at Stop 4, this outcrop exhibits a nubby-weathered surface due to differentially eroded quartz and sillimanite. A 30 cm layer of garnet amphibolite is isoclinally folded by  $F_2$ . Here,  $S_2$  strikes  $N80^\circ W$  and dips  $83^\circ SW$  with  $L_2$  lineations trending  $N85^\circ W$  at  $26^\circ$ . A spaced  $S_4$  slip cleavage oriented  $N26^\circ E$ ,  $84^\circ NW$  deforms  $S_2$ .

The Waramaug crops out to the northeast and southwest (Fig. 11). Those to the northeast show  $S_1$  metamorphic layering trending  $N20^\circ E$ ,  $17^\circ NW$  folded by upright antiformal  $F_2$  "m" folds with axial surfaces oriented  $N75^\circ W$ ,  $90^\circ$  and hingelines trending  $N75^\circ W$  at  $17^\circ$ . Outcrops to the southwest illustrate the typical massive, nubby-weathered appearance of the Waramaug Formation.  $S_2$  strikes  $N80^\circ W$  and dips  $83^\circ NE$  with a prominent  $L_2$  lineation trending  $N70^\circ W$  at  $36^\circ$ .  $L_2$  is produced by intersection with  $S_1$  which is locally preserved at a small angle to  $S_2$ . Commonly,  $S_1$  is transposed into parallelism with  $S_2$  but here  $S_1$  dips  $70^\circ$ - $75^\circ NE$  due to  $F_2$  isoclinal folding. The axial surface traces of  $F_2$  isoclines are indicated in figure 11.

Note the abundance of garnet at the southern margin of the gneiss. The enrichment marks the contact effects of adjacent flow-layered diorite of the Hodges Complex. The diorite forms a small sill intruded parallel to  $S_2$  in the Waramaug (Fig. 11). The microscope shows that the contact garnets (up to 1 cm) have grown across  $S_2$  in the Waramaug wall-rocks. Garnet enrichment also occurred in the diorite suggesting limited alumina metasomatism from the wallrocks took place.

In the woods to the east and west, Hodges diorites were intruded as sills and lit-par-lit injections (typically less than 10 m thick) along  $S_2$  in the Waramaug. Flow-layering, defined by oriented hornblende and biotite set in a plagioclase matrix, is regionally parallel to  $S_2$ . In addition, garnetiferous diorite sills to the east are folded by SW-plunging  $F_4$  folds and cut by a spaced  $S_4$  biotite schistosity (developed during  $M_3$ ) oriented  $N25^\circ E$ ,  $75^\circ NW$ .

The near east-west trend of  $S_2$  in this area is due to the effects of the major dextral  $F_4$  fold in the area (Fig. 4). On the flanks of the

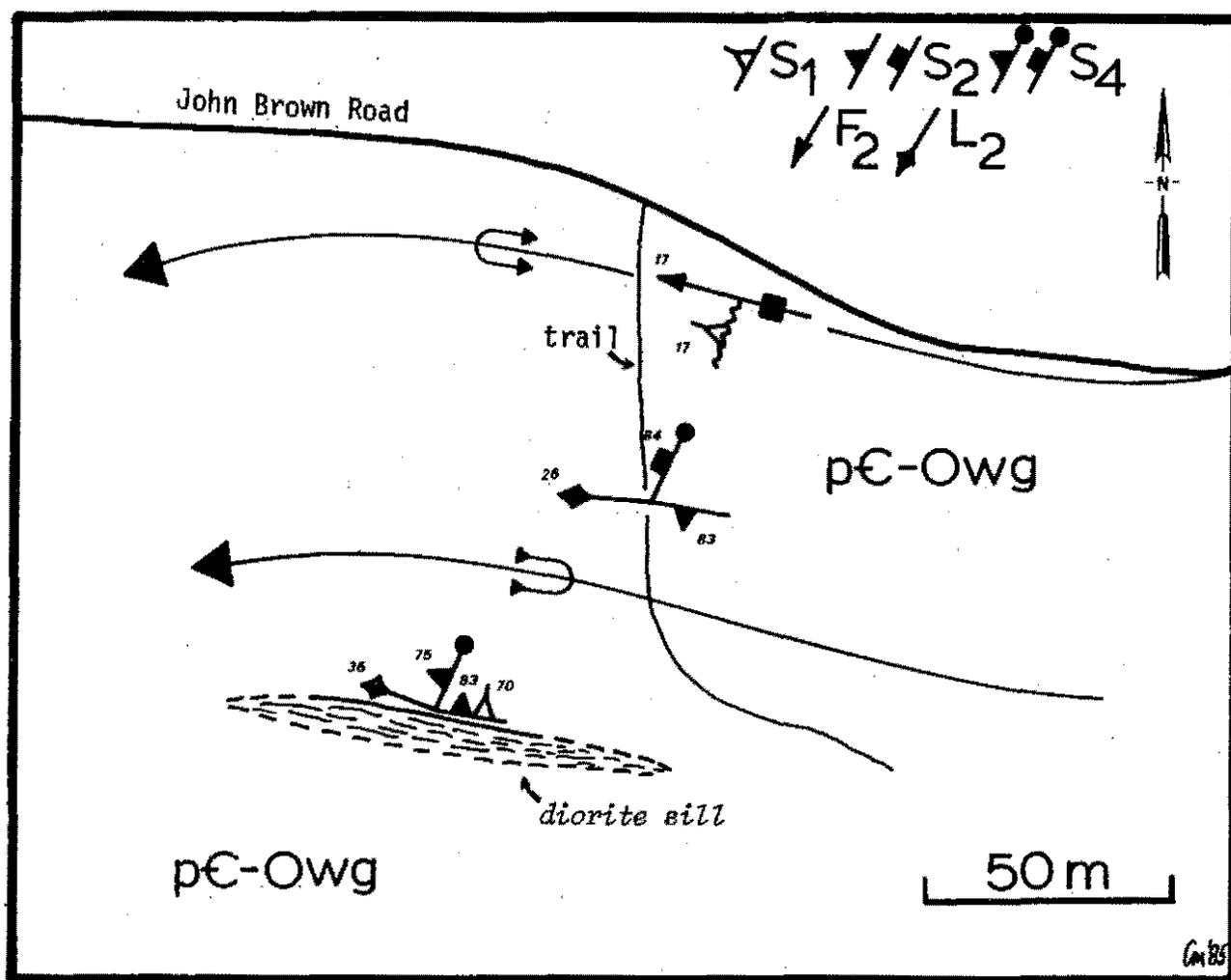


Figure 11 - Geological sketchmap of Stop 5 showing outcrops described in text and trace of major subvertical  $S_2$  axial surfaces.

$F_4$  fold (Stops 2-4)  $S_2$  trends approximately NNE but in the  $F_4$  hinge areas (Stops 1, 5) near east-west trends occur.

The post- $S_2$  field and textural evidence on the contact relations of the diorites fixes the time of intrusion to late-syn to post- $D_2$ .  $D_4$  crosscutting deformation and  $M_3$  overprinting places an upper relative age limit on the Hodges intrusive episode.

- |      |     |  |
|------|-----|--|
| 14.3 | 0.0 | Backtrack east on John Brown Road and turn right (south) on University Drive toward route 4. |
| 15.4 | 1.1 | Turn right (west) on route 4 and continue past Klug Hill Road and Wright Road to Weed Road.  |
| 16.4 | 1.0 | Turn left (south) onto Weed Road and park about 0.1 miles south along the side of the road.  |

16.5 0.1 The 1320' hill to the west is Stop 6.

Stop 6 (Optional) - Mafic and ultramafic rocks of the Hodges Complex

Walking up the overgrown trail westward from Weed Road, the hill to the west is primarily composed of hornblende gabbro which is locally melanocratic with porphyritic textures. A west-dipping flow-layering is defined by mafic mineral concentrations and oriented hornblende. Coarse-grained pyroxenite and hornblendite crops out near the top of the hill and in a small pod to the south (Fig. 5).

Cameron's Line is here masked by intrusives but based on detailed tracing of screens and xenoliths, it traverses across the top of the hill in a S20°W direction. The Hodges rocks are in contact with both the Waramaug and Hartland to the west and east, respectively.

To the east, many xenoliths and screens of the Hartland (upper member) are contact metamorphosed with the development of cordierite, kyanite, sillimanite, staurolite, and garnet and the elimination of muscovite. Garnets up to 3 cm across overgrow the S<sub>2</sub> foliation in the Hartland amphibolite. Float draping the slopes to the east of Weed Road shows hornblende porphyroblasts overgrown on S<sub>2</sub> in Hartland amphibolite. Contact metamorphic assemblages are fully discussed at the next stop (7).

To the west near outcrops of the Waramaug, flow-layered diorite trends N25°E with vertical to steep easterly dips. The Waramaug is a dense hornfels peppered with garnet but the characteristic nubby-weathering is still preserved. Along the western slope of the 1320' hill the Waramaug contains white tremolitic calc-silicate layers.

16.5 0.0 Return to route 4 and turn right. Drive slow and prepare to make a left at Wright Road.

17.1 0.6 Follow Wright Road north and park near the first barn on the left.

17.2 0.1 The ridge to the northwest is Stop 7.

Stop 7 - Mafic and ultramafic rocks of the Hodges Complex

Walk from the barn northwestward onto a ridge where flow-layered diorites grade westward into gabbroic rocks. The diorites were multiply intruded as thin sill-like masses parallel to S<sub>2</sub> in the Harland Formation. Both the Hartland and the diorites are strongly enriched in garnet in the contact zone. Flow layering in the diorites is oriented N65°E, 70°NW. Massive gabbroic rocks occur near the top of the ridge and extend toward the northeast (Fig. 5).

To the west, beyond the ridgecrest, a large NE-trending mass of pyroxenite and hornblendite crops out (Fig. 5). The ultramafic mass, which crosscuts the diorite-gabbro contact and truncates flow-layering in the diorite, is interpreted as the youngest intrusive in the Hodges Complex. Locally, the ultramafic rocks are strongly sheared and trans-

formed into laminated serpentinite. The shear zone, oriented N36°E, 75°NW, cuts across the central part of the Hodges and is due to shearing along the axial surface of the major dextral  $F_4$  fold which deformed Cameron's Line, the Hodges Complex, and the Tyler Lake granite into a broad dextral flexure. In fact, the plutons may have acted as immobile plugs localizing the  $F_4$  hinge area. To the southwest the  $S_4$  shear zone is marked by zones of serpentinitization and chloritization although domains of relatively unaltered pyroxenite and hornblendite are preserved.

Contact metamorphism of the Hodges wallrocks. The contact mineralization of the Hodges Complex overprints  $S_2$  in both the Waramaug and Hartland Formations. In the contact aureole, typical foliated textures in the wallrocks are replaced by dense, finer-grained, garnet-rich, hornfels (Fig. 12a). Amphibolite in the contact aureole contains post- $S_2$  garnet and hornblende as discussed at Stop 6 (Fig. 12b). Where gabbroic rocks intrude the Hartland Ohc subunit, a randomly oriented colorless amphibole of the cummingtonite-grunerite series forms. To the northeast of Klug Hill, gabbroic rocks intrude the Hartland granofels (Ohgr) producing a unique cordierite-kyanite-staurolite-garnet-biotite-plagioclase ( $An_{23}$ ) assemblage (Fig. 12c). Abundant staurolite occurs in the contact aureole of Stop 6 and garnet there contains sillimanite microlites.

Table 2 shows the contact assemblages in the wallrocks of the Hodges Complex compared to regional assemblages outside the aureole. Randomly oriented contact phases, the assemblage cordierite-kyanite-staurolite-garnet, and the absence of muscovite are characteristics of the Hodges aureole. These traits indicate that the Hodges was statically intruded between 5-8 kb (20-25 km) and 675°-700°C, near the  $Al_2SiO_5$  triple point (Merguerian 1977).

There is no clear overprinting of contact minerals by  $M_2$  kyanite, staurolite, or garnet porphyroblasts. Rather, identical minerals formed in the contact aureole suggesting that intrusion of the Hodges and regional  $M_2$  metamorphism were coeval (Table 1). Local temperature increases adjacent to the Hodges Complex produced cordierite+sillimanite and fostered the breakdown of muscovite.

- |      |     |  |
|------|-----|--|
| 17.2 | 0.0 | Drive back to route 4 and turn left. Follow route 4 east to Lovers Lane.                                   |
| 18.2 | 1.0 | Turn right onto Lovers Lane and another immediate right into Ducci Electrical Contracting Co. parking lot. |
| 18.3 | 0.1 | The Tyler Lake granite crops out in the creek bed along the south edge of the lot.                         |

End of log

Stop 8 - Tyler Lake granite

Tan-weathering, medium-grained, foliated quartz-microcline-plagioclase-muscovite-biotite-garnet-(chlorite)-(apatite) granite is exposed near the creek bed. An X-ray fluorescence analysis by Dr. D.

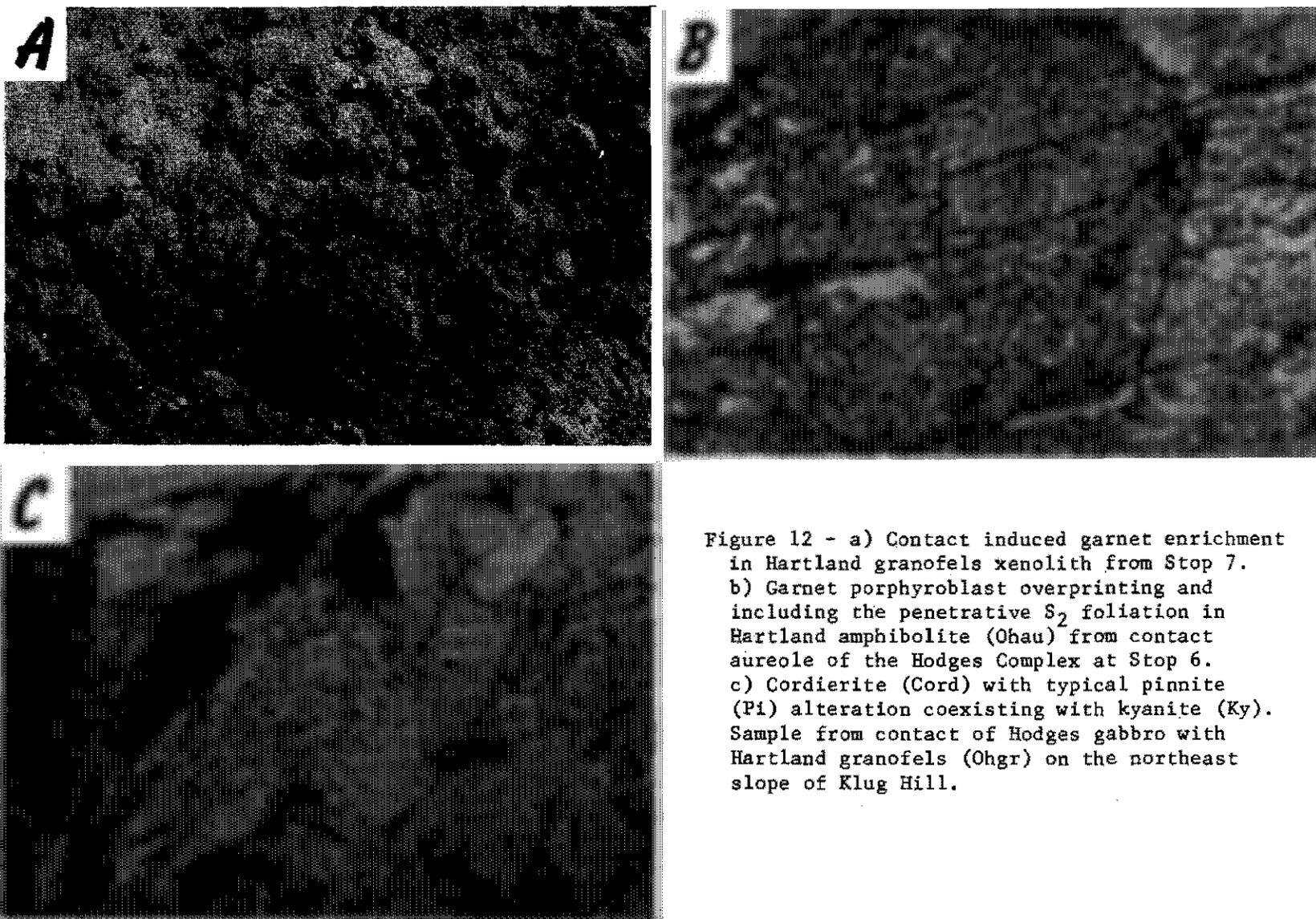


Figure 12 - a) Contact induced garnet enrichment in Hartland granofels xenolith from Stop 7. b) Garnet porphyroblast overprinting and including the penetrative  $S_2$  foliation in Hartland amphibolite (Ohau) from contact aureole of the Hodges Complex at Stop 6. c) Cordierite (Cord) with typical pinnite (P1) alteration coexisting with kyanite (Ky). Sample from contact of Hodges gabbro with Hartland granofels (Ohgr) on the northeast slope of Klug Hill.

<u>SAMPLE</u>	<u>UNIT</u>	<u>CONTACT ASSEMBLAGE</u>	<u>REGIONAL ASSEMBLAGE AWAY FROM CONTACT</u>	<u>REMARKS</u>
H-169	Ohgr	cord-ky-st-gt-bi- plag-qtz-chl	bi-musc-qtz-plag±ky	Contact with gabbro on Klug Hill. Sample was 5 feet from contact.
H-31	Ohgk	cord-ky-st-bi-qtz	musc-bi-qtz-plag-ky±st	Contact of gabbroic and ultra- mafic rocks with a screen of Hartland rocks. From the SW of the intersection of Weed Road and Route 4.
H-68	Ohc	st-gt-qtz-bi-sill	qtz-musc-plag	
H-69	Ohgk	cord-ky-gt-st-sill- bi-qtz-chl	musc-bi-qtz-plag-ky-±st	
H-30b	Ohau	hyp-hb-plag-bi-op	hb-plag-bi-op	
H-43	pE-Owg	gt-plag-qtz-bi	gt-plag-qtz-bi	From contact of Waramaug and dio- rite NNW of the Hodges Complex.
H-52A	pE-Owg	ky-bi-qtz-plag-gt	musc-gt-ky-qtz-plag	
H-116	Ohc	grun-qtz	musc-bi-qtz-plag	From contact with diorite N of Route 4 and Klug Hill Road.
H-56	pE-Owga	hb-plag-bi-gt-op	hb-plag-bi-op	Direct from contact with diorite N of the Hodges Complex.
H-36	Ohau	hb-plag-bi-gt-op	hb-plag-bi-op	Direct from contact with ultra- mafic rocks N of the Hodges Nickel Prospect.

KEY:

cord = cordierite  
ky = kyanite  
st = staurolite  
gt = garnet  
bi = biotite

plag = plagioclase  
qtz = quartz  
chl = chlorite  
sill = sillimanite

hyp = hypersthene  
hb = hornblende  
op = opaques  
grun = grunerite

TABLE 2 - Contact assemblages in country rocks near the Hodges Complex compared with the regional assemblage outside the aureole.

Radcliffe of Hofstra University produced the following result:  $\text{SiO}_2 = 73.0$ ,  $\text{Al}_2\text{O}_3 = 14.2$ ,  $\text{Fe}_2\text{O}_3 = 1.4$ ,  $\text{MgO} = 0.6$ ,  $\text{CaO} = 0.8$ ,  $\text{K}_2\text{O} = 5.5$ ,  $\text{Na}_2\text{O} = 3.1$ ,  $\text{TiO}_2 = 0.2$ ,  $\text{MnO} = 0.1$ , loss on ignition = 0.7 (total = 99.6). The granite is foliated by cm-spaced micaceous layering ( $S_4$ ) oriented  $N36^\circ E$ ,  $60^\circ NW$ . The  $S_4$  foliation is cut by a faint slip cleavage ( $S_6?$ ) oriented  $N10^\circ E$ ,  $42^\circ SW$ .

The Tyler Lake granite contains xenoliths of the Hodges rocks and is in direct contact with all major metamorphic units in the area (except G-Ohmk and E-Oha) suggesting a young intrusive age. Widely separated sample suites from the granite yield a well-defined  $466 \pm 12$  m.y. Rb-Sr isochron with initial Sr 87/86 =  $0.7082 \pm 0.0011$  (Merguerian and others 1984). Since the Hodges was intruded following or nearly synchronous with  $D_2$ , this medial Ordovician age is proof of a Taconian or possibly older age for Cameron's Line. The Sr 87/86 data implies that the Tyler Lake granite was derived either from anatectic melting of the continental crust or materials derived from the crust (i.e.-Waramaug and Hartland sequences). Assimilation may have also been an important process during intrusion of the granite. The Hodges Complex may have been emplaced during the late stages of the Taconic orogeny when oversteepened subduction into a thick orogenic welt tapped mafic and ultra-mafic magmas which ascended along Cameron's Line.

Thanks for coming on the trip - hope you enjoyed yourself!!



Route 4 (east) leads back to the center of Torrington and then to route 8.

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GEOLOGY OF  
THE MT. PROSPECT REGION, WESTERN CONNECTICUT

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INTRODUCTION

Cameron's Line which extends through the Mt. Prospect region, is the map trace of a thrust fault that forms a major tectonic boundary in western Connecticut (Fig. 1). Rocks east of it are believed to have been deposited on oceanic crust and to have been transported westward onto North American continental crust during the closure of the Iapetus Ocean basin in the Ordovician. The Mt. Prospect Complex is a series of dominantly mafic intrusive igneous rocks east of Cameron's Line and thus intrusive into rocks originally deposited in the Iapetus Ocean. Numerous other mafic intrusives are present east of Cameron's Line (Fig. 1), and the time of emplacement of all these intrusives has been a long standing question in piecing together the geologic history. Recently some (Robinson and Hall, 1980, Hall and Robinson, 1982) have favored early emplacement of the mafic intrusives prior to major motion on the Cameron's Line thrust fault, while others (Merguerian and Ratcliffe, 1977, Merguerian, 1983) have argued that at least one of the mafic intrusives crosscuts Cameron's Line and thus postdates the major thrust movement. Evidence in the Mt. Prospect region indicates that the Mt. Prospect Complex was emplaced prior to major transport along the Cameron's Line thrust fault.

Five phases of deformation have been recognized in the rocks of the Mt. Prospect region. The earliest phase recognized is evident as a foliation in the diorite and as minor folds in the country rocks. Isoclinal folds and associated prominent regional axial plane foliation mark the second phase and third phase isoclinal folds re-fold the second phase folds. Fourth phase folds produce variations in the trends of all earlier structural elements, and the fifth phase of deformation is marked by northwest trending open folds and a conjugate set of folds that trends east to northeast. Major transport along the Cameron's Line thrust fault and associated thrust faults is considered to have occurred during the second phase of deformation.

## STRATIGRAPHY

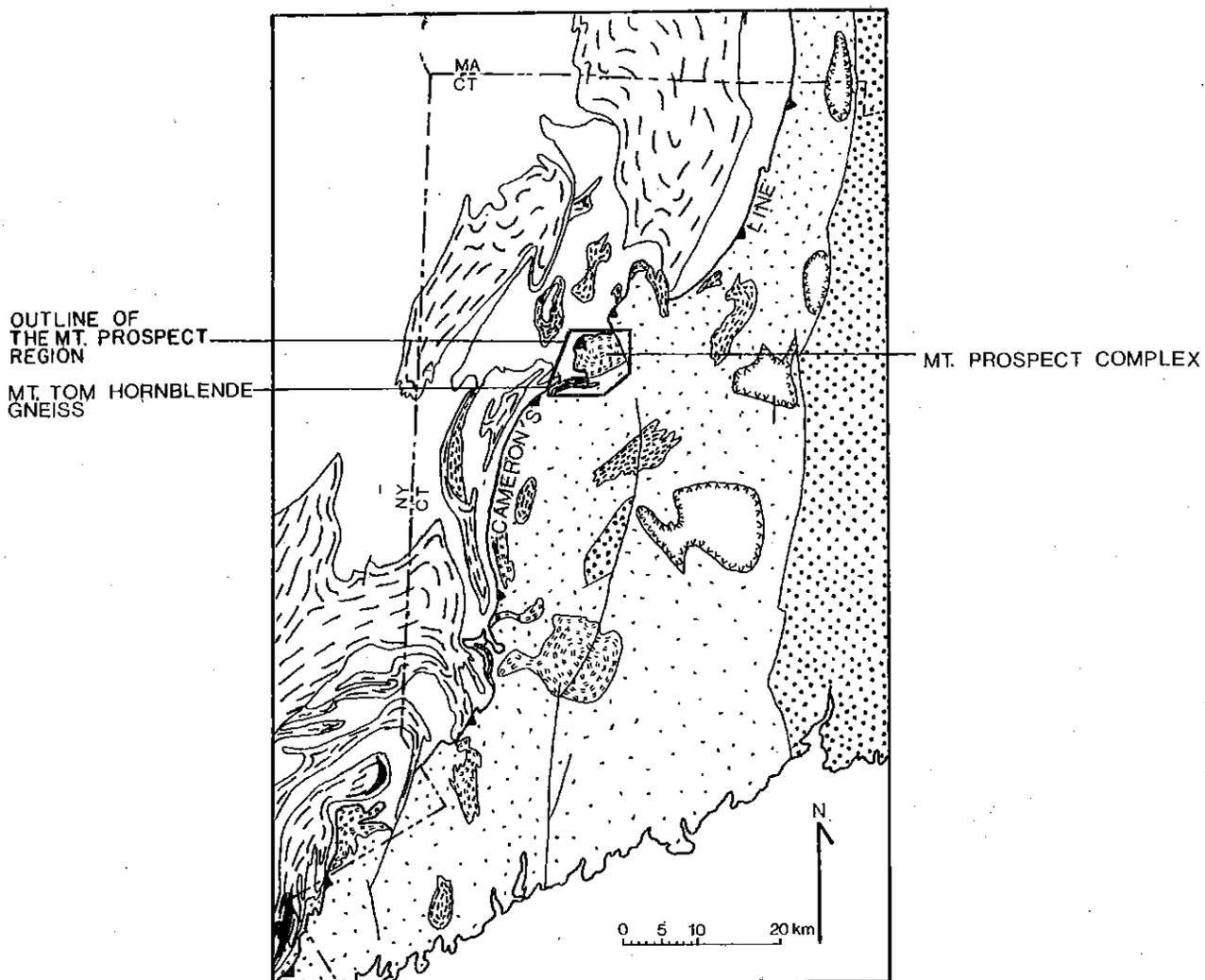
The rocks of the Mt. Prospect Region have been divided into 21 mappable units ranging in age from Precambrian to Ordovician. There are four stratigraphic-tectonic groups consisting of the Precambrian basement with its autochthonous Paleozoic cover rocks and three thrust sheets, the Waramaug Sheet, Above All Sheet, and Cameron's Line Sheet.

Rocks of the Grenvillian basement are dominantly gray, biotite-quartz plagioclase gneisses.

The autochthonous section consists of Lowerre Quartzite, Inwood Marble, and Manhattan A Schist. Lowerre Quartzite is made up of bedded, white to tan, locally slabby, quartzite, siliceous granulite and microcline-bearing schistose granulite. Inwood Marble is well bedded, white, calcite-tremolite-dolomite marble with thin tremolite-rich beds, and white, calcite-cemented, dolomite marble. Manhattan A is rusty-weathering, sillimanite-garnet-muscovite-plagioclase-biotite-quartz schist with minor, siliceous granulite beds.

The Waramaug Thrust Sheet contains the Shepaug Member, Schistose Granulite Member, an Amphibolite Member, and the Garnetiferous Biotite Schist Member of the Manhattan C. The Shepaug Member is interbedded orthoclase-garnet-plagioclase-biotite-quartz schistose gneiss, rusty-weathering, schistose gneisses and schists with sillimanite rods and subordinate granulite beds with sillimanite nodules similar to those in the Schistose Granulite Member. The Schistose Granulite Member is dominantly a fine-grained, well bedded, garnet-muscovite-biotite-plagioclase-microcline-quartz schistose granulite or granulite with locally large, muscovite-sillimanite-quartz ellipsoids or nodules, and subordinate, rusty-weathering, biotite schist with sillimanite quartz rods. The Amphibolite Member is a fine-grained, dark-green to black, plagioclase-biotite-hornblende gneiss with local thin layers of fine-grained, plagioclase-rich gneiss. The Garnetiferous Biotite Schist Member is largely charcoal-gray, garnetiferous, sillimanite-muscovite-quartz-biotite schist with sillimanite nodules that protrude on the weathered outcrop. These schists are interbedded with minor thin beds of laminated gneisses.

Manhattan C in the Above All Thrust Sheet is divided into an Amphibolite Member and the Warren Member (Dana, 1978). The Amphibolite Member is a lineated to foliated quartz-labradorite-hornblende gneiss. The Warren Member includes mainly interbedded, dark-gray, muscovite-garnet-chlorite-plagioclase-biotite-quartz schist, amphibolite, siliceous granulite, and finely-layered schistose gneiss.



**EXPLANATION**

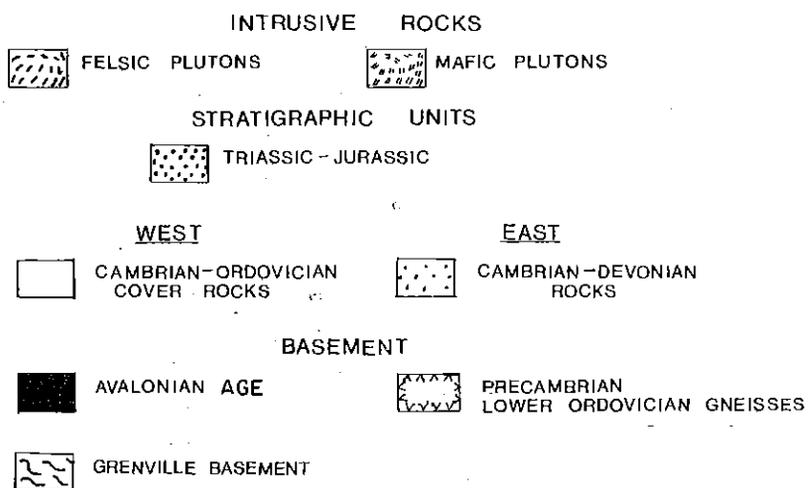


Figure 1: Generalized geologic map of western Connecticut showing the location of the Mt. Prospect area (modified after Hall, 1980).

Explanation:

ROCK UNITS

C3-4

- Above All Thrust Sheet**
- Emw - Warren Member
  - Emwa - Amphibolite Member
- Above All Thrust Fault**
- Waramaug Thrust Sheet**
- Embs - Garnetiferous Biotite Schist Member
  - Em a - Amphibolite Member
  - Emgr - Schistose Granulite Member
  - Em s - Shepaug Member
- Waramaug Thrust Fault**
- Oma - Manhattan A
- Middle Ordovician Unconformity**
- Ei - Inwood Marble
  - Ei - Lowerre Quartzite
- Unconformity**
- pEg - Precambrian Gneisses

SYMBOLS

- ② Field Trip Stop
- Thrust Fault (Teeth in Allochthon)
- Late High Angle Fault (Teeth on Downthrown Side)

Cameron's Line Thrust Sheet

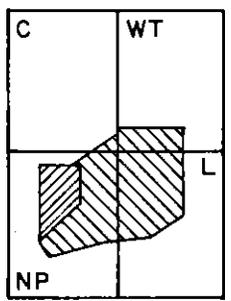
*Mt. Prospect Igneous Complex*

- ?Op - Quartz Monzonite Porphyry
  - ?Oon - Olivine Norite, Mafic Norite, Peridotite, Pyroxenite
  - ?Oqn - Quartz Norite, Norite, Gabbro, Hornblende-Biotite Gneiss
  - ?Ol - Layered Diorite
  - ?Od - Homogeneous Mafic Diorite
- } Late Intrusives
- } Early Diorites

*Sedimentary and Volcanic Rocks*

- ?COhb - Bee Brook Member
  - ?COmt - Mt. Tom Amphibolite
  - ?COhm - Muscovite Schist Member
  - ?COhr - Rusty Schist Member
  - ?COa - Amphibolite Member
- } Hartland Formation
- } Mt. Prospect Formation
- Cameron's Line Thrust Fault**

Manhattan C



QUADRANGLES

- C - Cornwall
- WT - West Torrington
- L - Litchfield
- NP - New Preston
- Mapped by Panish
- Mapped by Dana (1978)



The Cameron's Line Thrust Sheet contains the Mt. Prospect Formation, the Hartland Formation, and the Mt. Prospect Complex. Rocks of the Mt. Prospect Formation are gray to brown, commonly rusty-weathering, thinly layered or laminated, sillimanite-garnet-muscovite-plagioclase-quartz-biotite schistose gneiss with subordinate, commonly laminated, siliceous granulite, gray quartzite, biotite schist, local coticule, and garnet-cordierite-gedrite granulite. A mappable amphibolite horizon in the Mt. Prospect Formation is dark-green, slabby, well foliated, quartz-labradorite-hornblende gneiss. The Hartland Formation is divided into the Rusty-weathering Schist, Muscovite Schist, Mt. Tom Amphibolite, and Bee Brook Members. The Rusty Schist Member of the Hartland consists of gray or silver gray, rusty-weathering, laminated, garnet-plagioclase-muscovite-biotite-quartz schist that contains irregular quartz veins which stain red and thin, sandy, quartz layers, black, rusty-weathering, highly fissile, biotite schist, and subordinate, siliceous granulites. The Muscovite Schist Member is made up of silver-gray, garnet-biotite-oligoclase-quartz -muscovite schist with subordinate layers of biotite-rich schist and laminated quartzites, locally abundant staurolite and/or kyanite, and minor coticule layers. Dark-green to black, fine-to coarse-grained, massive, lineated, or foliated, quartz-andesine-hornblende gneiss constitutes the Mt. Tom Amphibolite. The Bee Brook Member is a thinly bedded unit of fine-grained, gray, massive to laminated, quartz granulites and quartzites, silver gray, muscovite schist, and subordinate, dark-gray, muscovite-biotite schist. Coticule layers are common.

The stratigraphy of the Mt. Prospect region is most easily understood by considering the rocks in the North American plate separately from those of oceanic affinity (Figs. 3, 4a). All known occurrences of Grenvillian basement in western Connecticut are west of Cameron's Line (Fig. 1). These basement rocks are unconformably overlain by an autochthonous section of clastic and carbonate rocks, the Lowerre Quartzite and Inwood Marble, which were deposited along the margin of the North American plate during the Late Precambrian through Early Ordovician. Middle Ordovician sulfidic schists of Manhattan A were deposited unconformably upon these older autochthonous rocks. Manhattan C is a facies equivalent of the basal clastics deposited eastward of the autochthonous section and then tectonically transported westward, locally in two thrust sheets, onto the autochthonous section along thrust faults that root into the Cameron's Line thrust fault.

The clastic and minor volcanic, Cambrian to Lower Ordovician Mt. Prospect and Hartland Formations are inferred to have been deposited east of the Manhattan C on oceanic basement.

The Mt. Prospect Igneous Complex was intruded into the Hartland and Mt. Prospect Formations. The Complex is thus considered to have an oceanic crustal or possibly mantle derivation.

The Mt. Prospect and Hartland Formations, and the Mt. Prospect Igneous Complex were thrust westward as a package along the Cameron's Line thrust fault onto North American basement and its cover rocks, including the allochthonous Manhattan C, during the Taconian Orogeny.

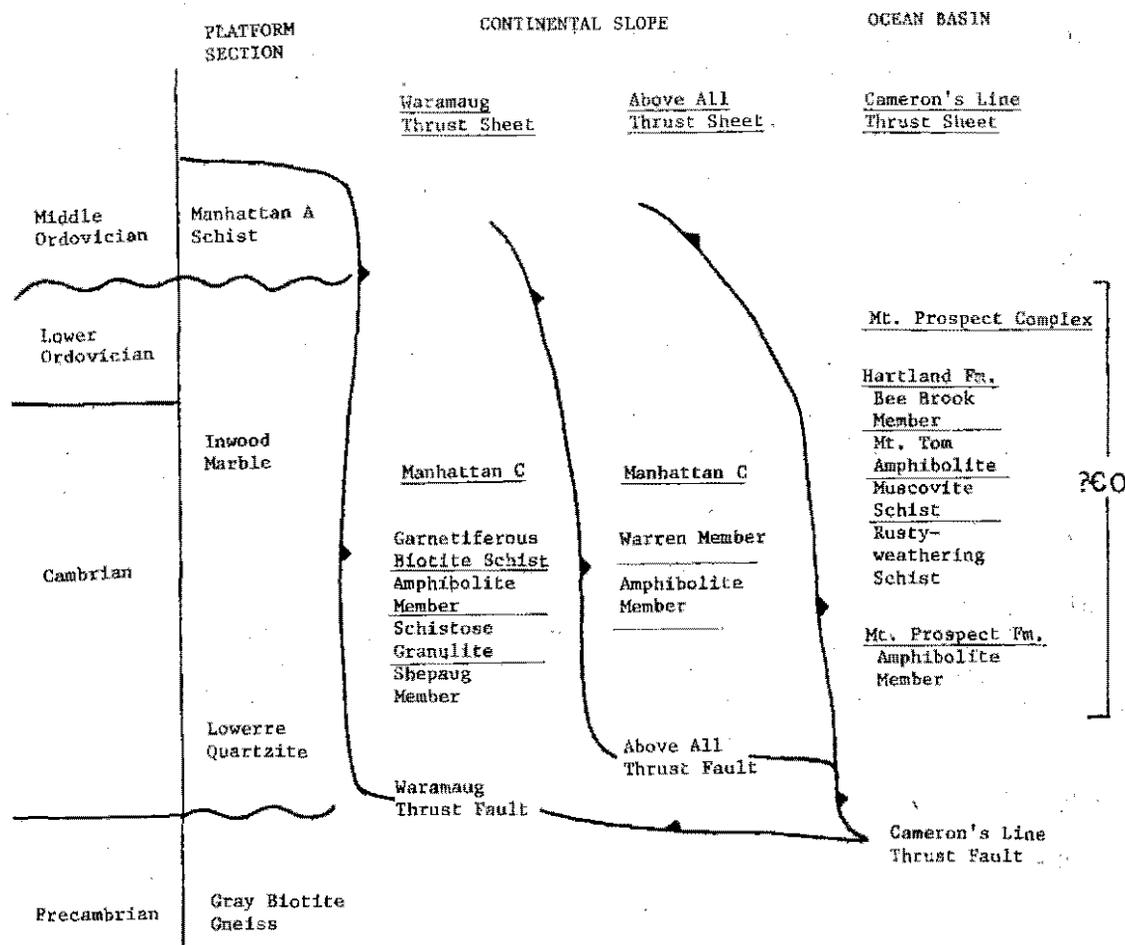


Figure 3. Stratigraphic summary chart showing the tectonic relationships among the rocks of the Mt. Prospect area.

## INTRUSIVE IGNEOUS ROCKS

The Mt. Prospect Complex is a highly deformed sheet of related igneous rocks. It was previously mapped by Cameron (1951) whose data were incorporated into quadrangle maps by Gates (1951) and Gates and Bradley (1952). It is a series of intrusive rocks covering a 33 sq. km area. Present mapping indicates that the intrusive rocks were emplaced into the Hartland and Mt. Prospect Formations in the following order: diorites, gabbros and norites, minor peridotites and pyroxenites, and quartz monzonites (Fig. 2). Minor diorite dikes and small hornblendite bodies are the youngest intrusives in the Complex. Cameron's Line trends northeast across the region and the northwest edge of the Mt. Prospect Complex is locally in thrust contact with rocks to the northwest where Cameron's Line is at the northwest boundary of the Complex (Fig. 2).

## MT. PROSPECT IGNEOUS COMPLEX:

Early Intrusives

## Homogeneous Mafic Diorite

This is a medium-to coarse-grained, indistinctly to well foliated, quartz-andesine-hornblende-biotite gneiss. Either biotite or hornblende may be the dominant mafic mineral. Accessory clinopyroxene may be diopside or augite. One minor additional rock type in this unit is a biotite-rich, scapolite-bearing gneiss with bright green diopside.

## Layered Diorite

This consists of interlayered gray, quartz-oligoclase- or andesine-hornblende-biotite gneiss layers of different modal composition and grain size, quartz monzonite layers, layers composed essentially of 2-4 mm oligoclase grains and matrix biotite, and wispy, centimeter thick, biotite-hornblende layers.

Late Intrusives

## Quartz Norite, Norite, Gabbro, Hornblende-Biotite Gneiss

This unit consists of several igneous rock types dominated by norite and gabbro, and displaying sharp to indistinct contacts with one another. For simplicity these rock types are shown as one unit on the map (Fig. 2). The oldest rock type, which is also the least abundant, is a hornblende-biotite gneiss that is only present near the summit of Mt. Prospect. It has the distinctive characteristic of being friable and apparently deeply weathered in outcrop even where

adjacent igneous rocks are completely fresh. This extensive weathering is probably due to exploitation of alteration products of later igneous rocks by weathering processes.

The norites form several igneous bodies with rock types differing in composition from quartz norite to mafic norite to gabbro. Outcrops may appear massive at first sight, but commonly some sort of foliation is present. This foliation can be an incomplete separation of mafic and felsic minerals into thin layers, preferred orientation of sedimentary rock inclusions, or preferred orientation of elongate feldspar, hornblende, or pyroxene.

Bronzite and plagioclase (andesine to labradorite) are the dominant minerals in the norites, and their relative abundances cause the rock to differ from felsic to mafic although mafic varieties predominate. Augite, hornblende, and biotite can also be important and accessory matrix quartz is commonly present. Where biotite and hornblende are minor they are only present as rims on augite and bronzite. Every gradation in essential mineralogy exists among the norites and gabbros.

In places the norites distinctly cut the diorites, but at other places there appears to be a gradational igneous sequence from norite and gabbro to unlayered diorite.

#### Olivine Norite, Mafic Norite, Peridotite, and Pyroxenite

This is a compositionally variable, mafic to ultramafic, rock unit which typically consists of a very mafic, massive, plagioclase-hornblende-augite-hypersthene norite which may also contain up to 10% olivine (Fo 65-75). Hornblende-bearing peridotite and hornblende-bearing pyroxenite are local ultramafic varieties that can also form distinct minor intrusives too small to show on Figure 2. Plagioclase, which is typically interstitial to the mafic minerals, ranges in composition from labradorite in the norite to accessory anorthite (An <98) in the peridotites. Hornblende can be a major constituent that rims and replaces olivine, augite, and hypersthene.

#### Quartz Monzonite

The quartz monzonite is typically a foliated, medium to coarse-grained, biotite-microcline-quartz-oligoclase to andesine gneiss or porphyritic gneiss with various amounts of 2-5 cm long, euhedral, white to pink, microcline megacrysts within a matrix of subhedral feldspar, quartz, and biotite. Locally the megacrysts may reach 15 cm in length. The quartz monzonite forms bodies several centimeters to hundreds of meters thick. The smaller bodies, which can be characterized as dikes, sills, or irregular patches, are particularly

abundant in the Layered Diorite. To a much lesser extent the quartz monzonite invades the norites, but it does, and that establishes the intrusive sequence.

#### Minor Late Intrusives:

There are several late intrusive rock types which form bodies too small to show on the geologic map.

#### Pyroxenite Dikes

Augite-hypersthene pyroxenite with minor labradorite and various amounts of secondary hornblende forms dikes that cut the quartz norites and olivine norites. These dikes are inferred from their similarity to the hornblendite veins mentioned next to postdate the quartz monzonite.

#### Late Hornblendite

Black, massive, biotite-augite-hornblendite forms several bodies of uncertain shape and size due to the lack of exposure. On the basis of hornblendite and augite-hornblendite veins that cut the quartz monzonite and that branch off the main hornblendite bodies, the main hornblendite bodies are inferred to postdate the quartz monzonite.

#### Late Diorite Dikes

There are numerous simple and composite fine-grained diorite dikes that cut all other igneous rock types of the Complex. One diorite dike crosscuts the quartz norite-Mt. Prospect Formation contact at the northwest corner of the Complex.

### STRUCTURAL GEOLOGY

Truncation of the Precambrian gneisses at the unconformity beneath the Lowerre Quartzite demonstrates that Precambrian deformation occurred (Dana, 1978), though this is not shown on Figure 2. There is evidence for five phases of Paleozoic deformation, although there are no map scale first phase features recognized (Figs. 2, 4 & 5, Table 1).

The only first phase features recognized are minor isoclinal folds and a foliation. An early foliation, but no recognized early minor folds, occurs in the layered diorite. At Stop 6 a first phase isoclinal fold is refolded by a second phase fold which has an axial planar foliation that is the dominant foliation of the region. Commonly two or more early foliations consisting of oriented minerals such as hornblende, biotite, quartz, and feldspar are present in the

TABLE 1  
SUMMARY OF DEFORMATION

PHASE OF DEFORMATION	FIELD TRIP LOCATION	DESCRIPTION
Fifth Phase (Acadian)	Stop 6 Stop 5, Sta. G,H Stop 8	N 10 to 60 W-trending, overturned, open folds developed. Map scale fifth phase folds are responsible for the broad, regional change in foliation strike from NE to E to N to NW across the Mt. Prospect region from the southwest corner to the northeast corner. Smaller fifth phase folds of the Mt. Tom Amphibolite are present at Mt. Tom.
	Stop 6 Stop 8	Open conjugate folds with NE to E trending, fine, axial plane foliation developed during the fifth phase of deformation. No map scale conjugate folds related to the fifth phase are recognized.
Fourth Phase (Acadian)	Stop 8, Sta. G Stop 5, Sta: G,H	Open to tight, NE-trending, fourth phase folds dominate the map pattern. The major change in trend of the axial surfaces of older folds from N-trending in the western part of the Mt. Prospect Complex to E-trending in the southern part of the Complex is due to a fourth phase fold.
Third Phase (Taconian)	Stop 5, Sta. G,H Stop 8	The third phase of deformation is responsible for the refolding of second phase folds in the Mt. Prospect Complex-country rock contacts. The map pattern of the Mt. Prospect Complex is interpreted to be due to the interference of second and third phase isoclinal folds.
Second Phase (Taconian)	Stop 5 Stop 6 Stop 7	The second phase is responsible for the dominant regional foliation, minor folds, and map scale isoclinal folds including folds of the Mt. Prospect Complex-country rock contact. The terminations of the Mt. Tom Amphibolite in the southeastern part of the area are interpreted to be due to a second phase fold hinge line that is repeated by later folding. The Cameron's Line, Waramaug, and Above All thrust faults are considered to be late second phase features.
First Phase (Taconian)	Stop 6 Stop 7	A minor first phase isoclinal fold is isoclinally refolded by a second phase fold at Stop 6. An early foliation that locally occurs in the layered diorite is interpreted to be first phase. No major first phase folds are recognized.

CS-11



layered diorite (Stop 7). Though it is difficult to determine relative ages with certainty, they are first phase and second phase features.

The second phase of deformation is Taconian and includes development of the dominant regional foliation, and both minor and map scale isoclinal folds. The map pattern of the Mt. Tom Amphibolite (Figs. 2 & 5) is interpreted to be a large refolded second phase isoclinal fold. In the western part (largely mapped by Dana, 1978) of the region, second phase isoclinal folds in the Manhattan C section are best explained as being pre-thrusting or syn-thrusting. This includes the second phase isoclinal fold of the Warren Amphibolite which is cut by the Above All Thrust, and the second phase fold in the Garnetiferous Biotite Schist which is cut by the Cameron's Line Thrust. On the other hand, second phase folds deform Cameron's Line (Stop 12). Third phase folds, such as those in the western part of the area and at Stop 12, also fold Cameron's Line. No thrust faults have been observed to cut third phase folds. The major motion on the Cameron's Line, Above All, and Waramaug thrusts is thus considered to have been during the second phase of deformation.

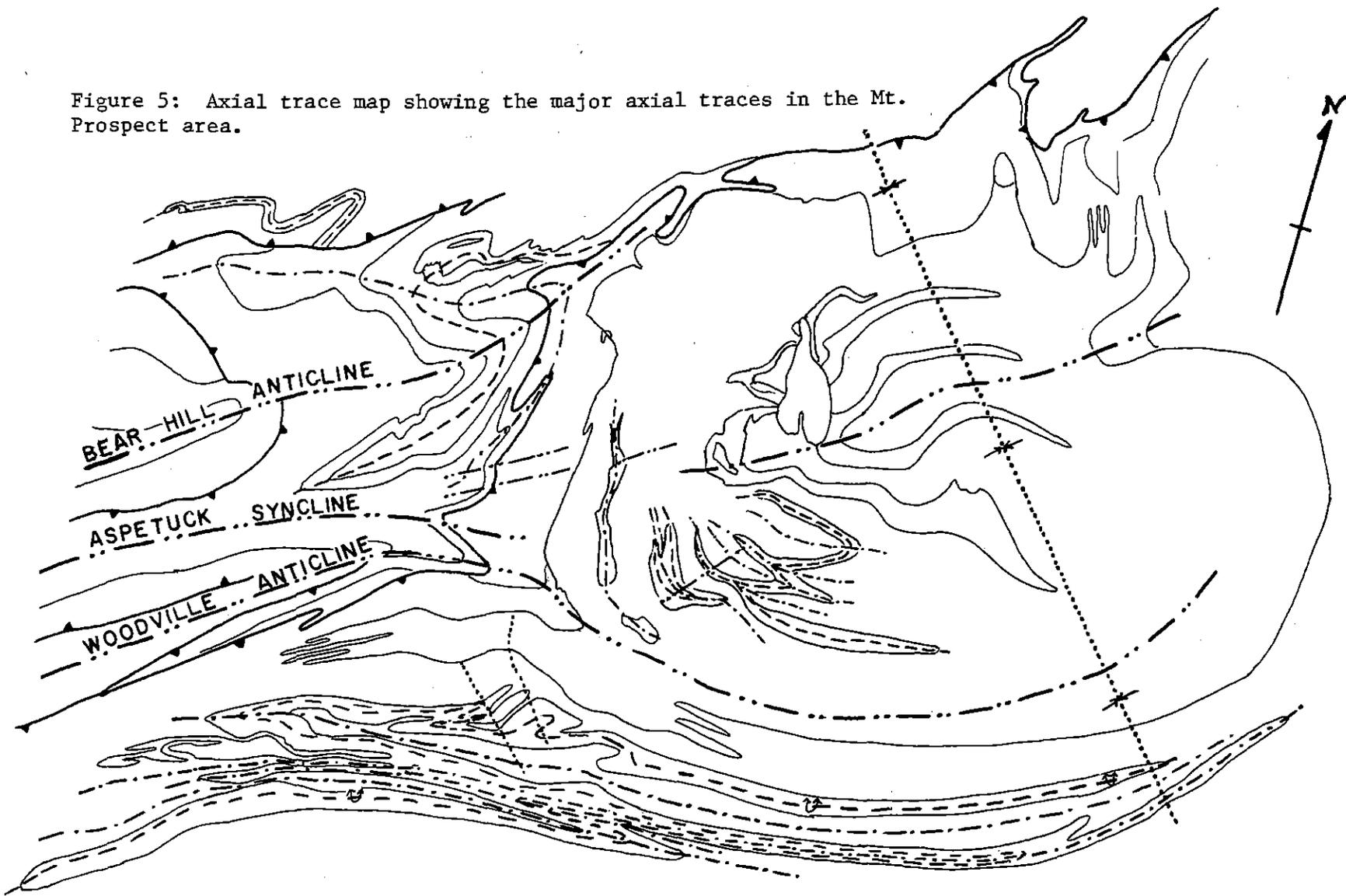
Second phase isoclinal folds have been refolded by tight to isoclinal third phase folds. Within the Mt. Prospect Complex, the interference of second and third phase folds has produced the distorted heart-shaped infolds of Mt. Prospect Formation and zig-zag infolds of the Bee Brook Member of the Hartland (Figs. 2, 4 & 5). The Mt. Tom Amphibolite has also been refolded by east trending third phase folds.

The Mt. Tom Amphibolite-country rock contact is typically concordant being parallel to bedding and the dominant second phase foliation. However the Mt. Tom Amphibolite crosses the Hartland Bee Brook Member-Muscovite Schist Member contact at map scale. Gates (1967) mentions four outcrops where the Mt. Tom Amphibolite apparently crosscuts the bedding and dominant foliation of the Hartland members and argues that the Mt. Tom Amphibolite was intruded during the deformation that caused the dominant foliation. There are two other possibilities. The Mt. Tom Amphibolite may be a lava flow or series of lava flows crossing a time transgressive facies boundary between the Bee Brook and Muscovite Schist Members, or there are early thrust faults cutting the Hartland section. In any of these three interpretations, the Mt. Tom Amphibolite was in place before the end of the second phase of deformation at the very latest.

Critical to all regional tectonic scenarios is the age of intrusion of the Mt. Prospect Igneous Complex. Regional and local evidence imply that all major igneous units of the Complex had intruded the Mt. Prospect Formation and Hartland Formation before

Figure 5: Axial trace map showing the major axial traces in the Mt. Prospect area.

03-14



Axial traces:  
 Second phase - - - - - Third phase - · - · - · Fourth phase - · - · - · Fifth phase ·······

major movement on Cameron's Line. The igneous rocks of the Complex lie completely east of Cameron's Line, do not cut Cameron's Line, and contain inclusions of the Hartland and Mt. Prospect Formations but not Manhattan C. Regionally mafic plutons (Fig. 1) similar to the Mt. Prospect Complex can also be interpreted to lie completely east of Cameron's Line. However Merguerian and Ratcliffe (1977) and Merguerian (1983) interpret one of the mafic plutons, northeast of Mt. Prospect, the Hodges Mafic Complex, as crosscutting Cameron's Line.

The second and third phase folds in the diorite-country rock contact (Stop 8) and the inferred first phase foliation in the diorite (Stop 7) demonstrate that the diorites, which are the oldest major intrusives, intruded before all known deformation. The quartz monzonite, which is the youngest major intrusive is pre-second phase being is folded by major third phase folds (Stop 8) and displaying the dominant second phase foliation as well as minor second phase folds (Stop 5). Its age relation to the first phase of deformation is uncertain.

The generally massive appearance of the norites and related rocks as well as their cross cutting, stock-like, map appearance suggest that they are the youngest intrusives. However in outcrop the quartz monzonite is seen to invade all norite types. These cross-cutting relationships are barely visible on Figure 2. The norites cut the original igneous layering of the diorites, but whether or not they postdate the first phase of deformation is uncertain. A late high angle fault cross-cuts second and third phase folds in the diorite and country rock at the southwest norite contact.

The Acadian fourth phase folds, which dominate the map pattern, are tight to open regional folds with NW-dipping axial surfaces and NE to N20W plunging axes. The major folds have counterclockwise rotation senses with major anticlinal limbs steeply overturned to the southeast. It is important to stress that the regional stratigraphy is consequently overturned (Fig. 4). The fourth phase folds are well defined and the axial surfaces are parallel west of the Mt. Prospect Complex, but they fan out near the Complex. The Bear Hill anticline turns more north and the Woodville Anticline turns to the east along the southern edge of the Complex. The Aspetuck syncline crosses the central part of the Complex and is responsible for the major fold of the Complex. Within the Complex it is difficult to trace a single synclinal axial surface. Instead there seems to be a number of smaller folds.

The dominant N20W to N plunge of the fourth phase folds is the overall plunge of the folded Mt. Prospect Complex sheet. The map pattern is interpreted to be an oblique section through a relatively steep, N-plunging cylinder with a roughly heart-shaped vertical cross

section. The E-W vertical cross section in Figure 4 is another oblique section through that cylinder.

A major NW-trending, Acadian (or Alleghanian) fifth phase synform is responsible for the broad, concave northward, warp in the foliation most easily seen along the south and east parts of the map area. Smaller fifth phase folds are seen in the Mt. Tom Amphibolite near Stop 6.

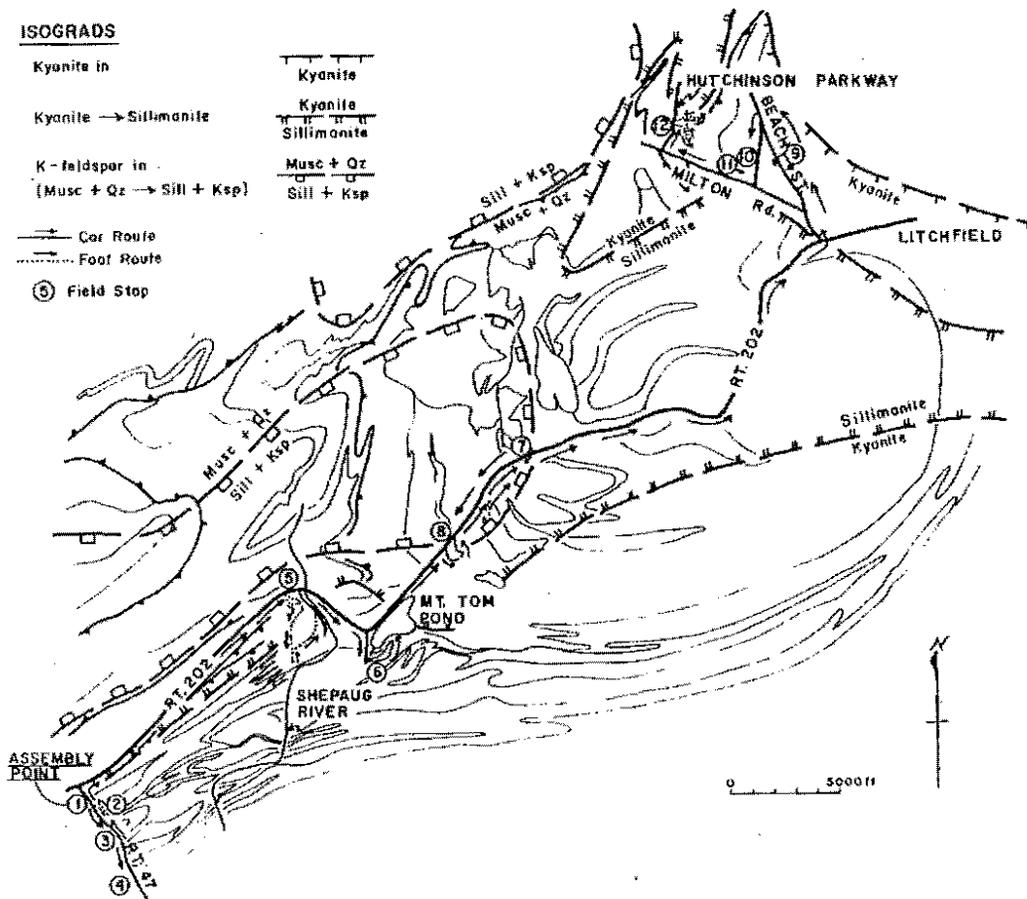


Figure 6: Isograd map of the Mt. Prospect area. Also shown is the field trip route.

## METAMORPHISM

The regional metamorphism ranges from staurolite grade in the northeast part of the map area to sillimanite-K-feldspar grade in the west and northwest (Fig. 6). Textural features indicate that the isograds are not the result of a simple prograde or retrograde metamorphic event. Highest grade metamorphism occurred before or during the third phase of deformation. Most biotite and muscovite flakes are oriented in the dominant second phase foliation. Sillimanite and sillimanite-quartz-orthoclase rods tend to be oriented in the dominant foliation and parallel to third or fourth phase fold axes. Kyanite and staurolite, while parallel to the foliation, commonly do not form strong lineations. Kyanite commonly overgrows sillimanite. Garnets commonly have only sillimanite inclusions where the matrix contains sillimanite and kyanite or only kyanite. Kyanite is thus relatively late and could be Acadian. Garnets typically show a prograde chemical zoning ( $Mg/(Mg+Fe)$ ) increasing toward the rim), but retrograde and complex zoning occur. Large retrograde muscovite plates, commonly containing remnant sillimanite, are oriented at a high angle to the foliation.

Local contact metamorphism is associated with the Mt. Prospect Complex. Garnet-sillimanite-orthoclase-biotite-cordierite and kyanite and/or sillimanite-garnet-biotite-cordierite-gedrite assemblages are typical in these zones. These assemblages pre-date the regional metamorphism and are partly retrograded by it.

Post-diorite norites and hornblendites show metamorphic alterations including fine, metamorphic pigeonite lamellae in augite and compound orthopyroxene-plagioclase-spinel-hornblende rims on olivine.

The maximum temperature and pressure estimates, using various geothermometers and geobarometers, range from 658 C, 6.9 kb in a sillimanite zone sample, to 535 C, 5.2 kb for a staurolite zone sample.

## ACKNOWLEDGMENTS

We wish to thank the Connecticut State Geological and Natural History Survey and the Geological Society of America for generous financial support. We also wish to express our appreciation for the permission, cooperation, and often keen interest of the various landowners whose properties we are visiting on this trip. Among them are the E. O. Phelps and Sons Co., Drillers, Fox Chase Farm, Mt. Tom State Park, James Dibble, John and Audrey White, and the Ferrara, McIntosh, and Andrighetti families. Marie Litterer kindly lettered the figures.

## ROAD LOG

From the New Haven, Connecticut area follow Route 34 west about 9 miles to Route 8 at Derby. Drive north on Route 8 for about 25 miles to the intersection with Route 109 south of the center of Thomaston. Continue west on Route 109 about 14 miles to the intersection with Route 47. Travel north on Route 47 about 3 miles. The assembly point is along the roadside at Stop 1, 0.1 mile south of the intersection with Rt. 202 (Fig. 7). A small white garage is next to the road. A white farmhouse and a barn with a bright red roof are 150 feet west of the road.

THE FIELD TRIP STARTS AT 9:00 AM.

Stops 1 through 4 are roadcuts and outcrops along Rt. 47 in the Bee Brook valley. The range of rock types in Manhattan A, and in the Rusty-weathering Schist, Muscovite Schist, Mt. Tom Amphibolite and Bee Brook Members of the Hartland Formation are seen at these stops (Fig. 7).

Bedding, and second and third phase planar elements have similar trends in the Bee Brook area (Fig. 8). Bedding, the dominant second phase foliation and second phase axial planes average N59E 76NW. The third phase axial planes average N42E 75NW. Second phase lineations consisting of isoclinal fold axes, quartz rods, which are boudinaged quartz layers or detached isoclinal fold noses, and schistosity-bedding intersections, define a great circle with N62E 80NW orientation, essentially parallel to the dominant schistosity, and thus indicating that they all lie in the dominant schistosity. Third phase fold axes have an average plunge of N88W 60.

The two second phase fold axes with clockwise rotation sense are in the Rusty-weathering Schist Member at Stop 2 near the southern contact with the Bee Brook Member. These two axes may be in the southern limb of a second phase map scale fold which crosses the road and closes to the west much like the smaller fold in the Rusty-weathering Schist Member to the north of Stop 2. The two third phase axes are also clockwise and are in the Rusty-weathering Schist near its northwestern contact, southeast of Stop 2. These third phase fold axes are on the southern limb of the third phase fold which refolds the Mt. Tom Amphibolite into a hook pattern about 1000 feet to the northeast.

A very fine, sporadically developed, SW dipping foliation consisting of tiny aligned quartz and biotite grains is fifth phase, but there are no fifth phase folds recognized in the Bee Brook area.

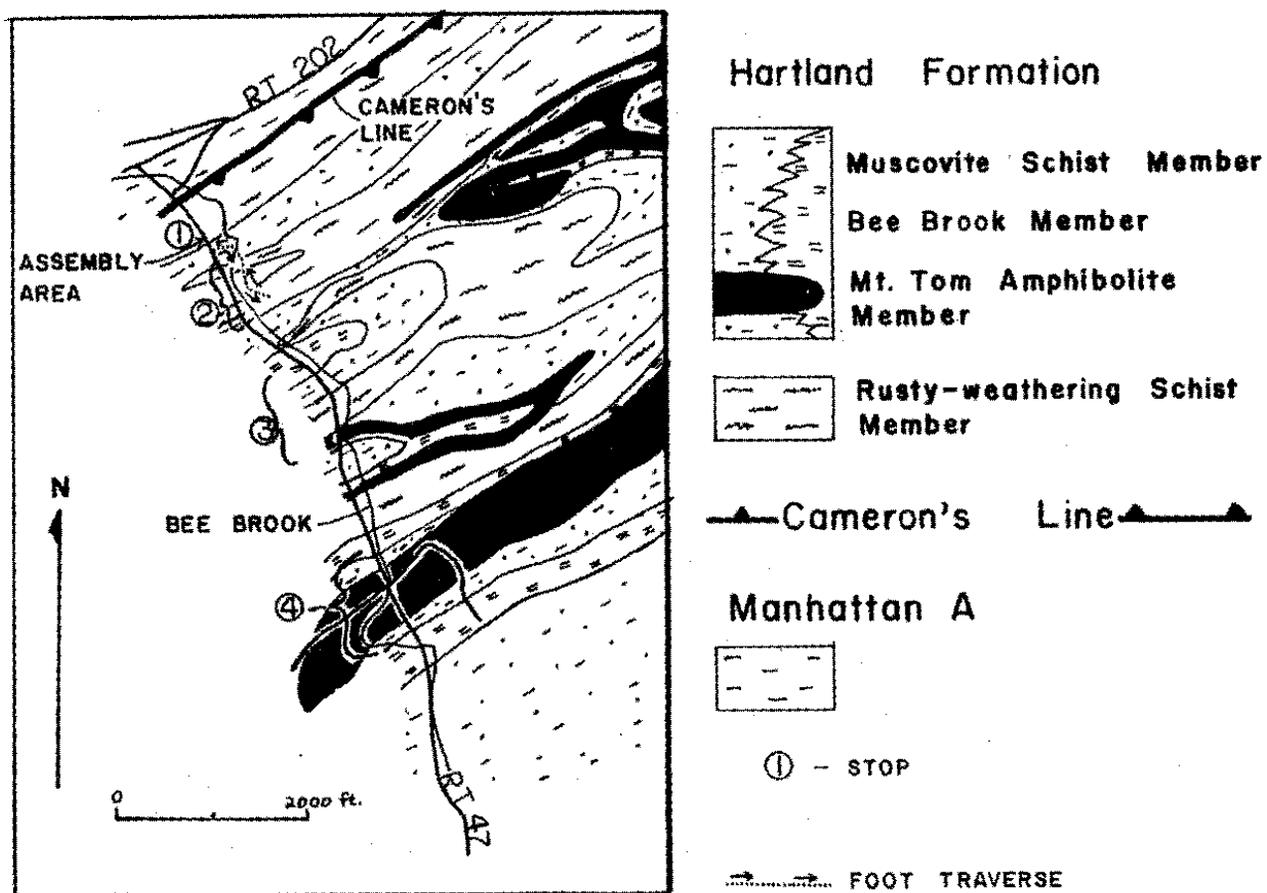


Figure 7: Geologic map of the area around Stops 1-4.

Stop 1: This stop consists of two small roadcuts, one on each side of the road and each about 100 feet north of the small white garage previously mentioned. Thinly-bedded, rusty-to mustard-yellow-weathering, schistose gneisses and schists are present. The dominant rock types include fine-grained, gray-brown staining, well foliated to laminated, slabby, sillimanite-garnet-plagioclase-muscovite-biotite-

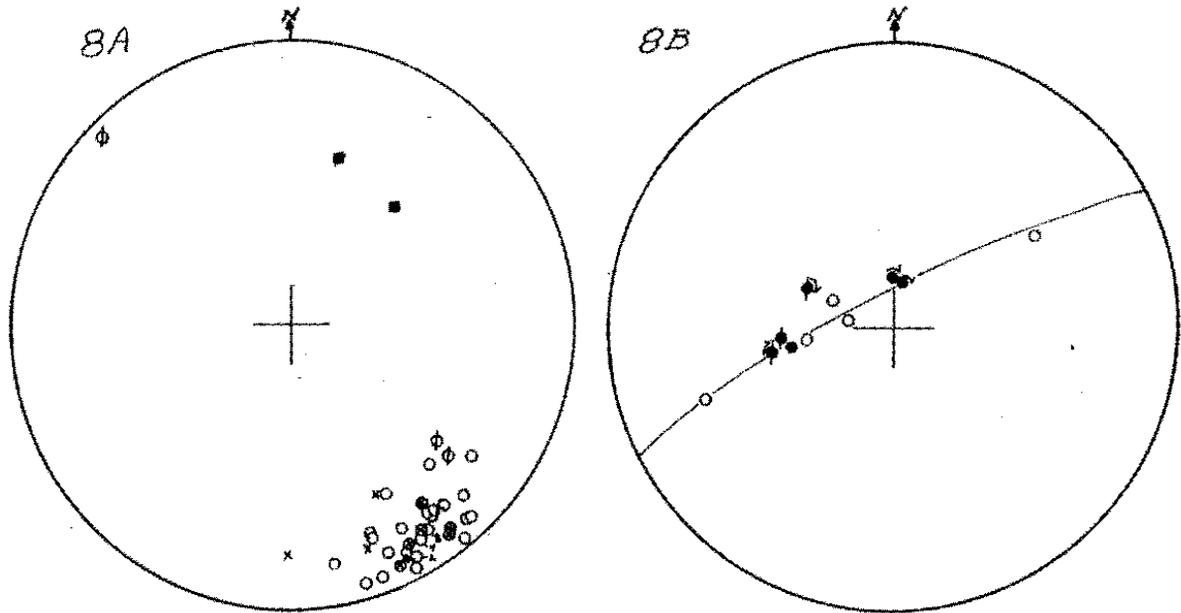


Figure 8A. Foliation data from  
Stops 1-4.

Poles to planar features = 45

- x bedding
- o dominant second phase foliation
- ⊗ axial plane of second phase  
minor fold
- ϕ axial plane of third phase  
minor fold
- fifth phase foliation
- “ number of data that the symbol represents

Figure 8B. Lineation data from Stops 1-4.

Lineations = 11

- o second phase mineral lineation
- ⊗ second phase fold axis
- † third phase fold axis

quartz schistose granulite with mustard-yellow-to rusty-weathering foliation surfaces. Fine-to medium-grained staurolite-garnet-plagioclase-muscovite-biotite-quartz schists and schistose gneisses are also abundant. Subordinately present are siliceous, laminated, garnet-muscovite-biotite-quartz granulites and quartzites; the laminations are most apparent on weathered surfaces. Biotite is commonly more abundant than muscovite, although coarse muscovite may be abundant on

foliation surfaces. Medium-grained, granular, millimeter centimeter scale quartz layers and lenses are distributed throughout the rock. A foot thick, dark, very fine-grained, quartz-plagioclase-hornblende-biotite gneiss layer bordered by 1-2 foot thick layers of staurolite-rich, muscovite schist is present in one outcrop.

Stop 2 consists of a number of outcrops of the Bee Brook and Rusty-weathering Schist Members of the Hartland located along a trail east of Rt. 47. The trail joins Rt. 47 about 300 feet south of the Stop 1 outcrops. Walk E to SE along the trail about 300 feet, ford Bee Brook, and walk about 350 feet SE along the trail to the first major outcrop of the Bee Brook Member. After stopping at this outcrop, we will continue east along the trail. Over the next 400 feet of trail, the Bee Brook and Rusty-weathering Members outcrop along the steep slope to the north. along the trail. Two members of the Hartland Formation, the Bee Brook and Rusty-weathering Schist Members, are exposed along a 600 foot stretch of trail east of Bee Brook.

The Bee Brook Member consists of thinly-bedded (1 cm to 2 m), fine-grained, gray, massive to laminated, quartz granulites and quartzites, fine-to medium-grained, silvery-gray, garnet-chlorite-biotite-quartz-muscovite schists, and subordinate, darker gray, muscovite-biotite schists. The granulites are dominant. All rock types have white, millimeter to centimeter thick, coarse-grained, granular, quartz laminae, wispy layers, and lenses. Salmon pink coticule layers and lenses are locally present.

The Rusty-weathering Schist Member consists of fine-to medium-grained, rusty-brown to rusty-yellow-weathering, laminated (on 1-4 mm scale), garnet-muscovite-biotite-plagioclase-quartz schistose gneiss, fine-grained, black, rusty-weathering, biotite schist, and subordinate beds of fine-grained, gray siliceous granulites. Also present are 1-10 cm thick, light-gray, sandy, quartz layers, and coarser-grained, brick red-to orange-staining, massive, irregular, quartz pods and layers.

#### Mileage

#### Total Interval

0.0	-	Return to the vehicles by way of the trail. Leave Stop 2 and proceed south on Rt. 47 to Stop 3.
0.4	0.4	Park along the west side of the road north of the first major roadcut (Stop 3).

Stop 3: Here (Fig. 7) are several roadcuts in Hartland members extending 550 feet along the roadside and displaying abundant minor folds. Starting at the north end and walking south, the following

rock units are encountered:

The Muscovite Schist Member is 10 to 15 feet of fine-to medium-grained, gray to silver gray, garnet-staurolite-chlorite-biotite-quartz-muscovite schist with subordinate, wispy, centimeter scale layers of a darker garnet-chlorite-muscovite-biotite schist. Also present are fine-grained, gray, siliceous granulite layers. Coarse-grained, massive, quartz lenses, pods, and bulbous fold noses <10 cm wide are abundant. Chlorite-biotite-rich schists with local silky, greenish-gray foliation surfaces are also present. The schistosity is commonly crenulated.

The Bee Brook Member which is exposed over an interval of 350 feet consists of thinly-interbedded quartzite, granulite, and schist similar to those at Stop 2. Bedding is from 1 cm to 2 m thick. Gray to slightly greenish-gray, thinly-bedded (3 cm to 1 m), fine-grained, garnet-biotite-chlorite-muscovite quartzites are interbedded with subordinate, gray and greenish-gray, thinly-bedded, fine-to medium-grained, chlorite-biotite-muscovite siliceous schists, and fine-grained, garnet-muscovite-chlorite-biotite-quartz schists. The quartzites are commonly massive, but may be foliated at minor fold hinges or layered on the millimeter to centimeter scale, the layering being due to slight variation in biotite and garnet. The muscovite schist is more abundant over the southern half of the interval. Here the muscovite schist has abundant chlorite patches, fine-grained sandy layers, and wispy biotite layers. Typically white, massive, medium-to coarse-grained quartz layers < 10 cm thick are present. Atypically the quartz layers are pink due to abundant fine-grained garnets. These coarse, white quartz layers may be veins, judging from the local quartz veinlets that interconnect them. Minor folds within these quartz layers demonstrate that the massive fine-grained quartzites are internally folded. Evidence for a progressive increase in intensity of shearing toward the south is present in the form of numerous fine-grained foliated or laminated siliceous layers <30 cm thick.

The Rusty-weathering Schist Member is exposed over the southern 185 feet of roadside. Over the first 25 feet the rock type is mainly a fine-grained, uniform, medium-to dark-gray, chlorite-muscovite-quartz-biotite schist and schistose gneiss with subordinate biotite-quartz granulite with coticule. The rocks are not rusty at this point and are pin-striped with white, fine-grained, quartz layers and granular quartz layers. The coticule forms aphanitic red-brown layers and isolated ellipsoids <5 cm long. The last 160 feet of exposure consists of a very fine-grained, highly fissile, dark-brown-stained, rusty-to yellow-green-weathering, biotite-rich schist with a few rusty-brown <8 cm thick, irregular, quartz layers and pods.

0.4 - Return to the vehicles and continue south on Rt. 47 to Stop 4.

0.9 0.5 Stop 4: Bee Brook Bridge area (Fig. 7). Park along the road north of the intersection of Rt. 47 and Buffum Road. The bridge over Bee Brook is south of the intersection. Walk to the northern contact of the Mt. Tom Amphibolite in the large roadcut north of Buffum Road. We will walk south along Rt. 47 to the southern contact of the Mt. Tom Amphibolite south of Buffum Road.

A 600 foot cross section of the Mt. Tom Amphibolite Member of the Hartland is exposed in two large roadcuts. The local contacts of the amphibolite with the Bee Brook Member are concordant to the compositional layering in the country rock and the dominant foliation.

The Mt. Tom Amphibolite is a dark-green to black, fine-to medium-grained, biotite-andesine-hornblende gneiss with minor quartz, sphene, epidote, and opaques. Thin epidote-rich veins are present. To the north the amphibolite is in contact with about 15 feet of muscovite schist which is interpreted to be a layer within the Bee Brook Member of the Hartland. The amphibolite is in contact with interbedded quartzite and siliceous granulite of the Bee Brook Member to the south. The Mt. Tom Amphibolite ends about 800 feet to the west of these roadcuts in an apparent early fold nose. Lateral facies changes or minor faulting within the Bee Brook Member may account for the lack of exact stratigraphic repetition across the axial surface of the early isoclinal fold. Two, concordant, meter thick, amphibolite layers occur south of the main contact. It is uncertain whether the Mt. Tom Amphibolite is repeated in isoclinal folds, or minor thrusting, or whether these are separate amphibolite layers.

0.9 - Return to your vehicles and head north on Rt. 47 to the intersection with Rt. 202.

1.9 1.0 Turn right (east) at the intersection onto Rt. 202 and continue to Stop 5. The lowland immediately north of Rt. 202 is underlain by the Inwood Marble which lies in the core of the major fourth phase Woodville Anticline. The hills beyond are underlain by the Manhattan A and C. Manhattan A is also present in the southern limb of the Woodville Anticline immediately south of Rt. 202. Manhattan C is truncated by the Cameron's Line thrust fault in the Bee Brook valley, but in the direction of Stop 5 the Manhattan C section appears and thickens between the Manhattan A and Cameron's Line.

4.9 3.0 The large rusty-weathering Manhattan A roadcut on the south side of Rt. 202 is the first station of Stop 5.

Drive east of the roadcut to the intersection of Rt. 202 with Rt. 341 and Romford Road (not named on the New Preston quadrangle map). Park at the corner of Rt. 202 and Romford Road, and walk back to the roadcut.

-PLEASE WATCH OUT FOR TRAFFIC-

Stop 5: This is a traverse that starts at the Manhattan A roadcut located near the nose of the Woodville Anticline and continues southward (Figs. 2, 9). Proceeding southward one crosses the Waramaug Thrust into Manhattan C and then across Cameron's Line into several members of the Hartland Formation as well as as quartz monzonite prophyry related to the Mt. Prospect Igneous Complex.

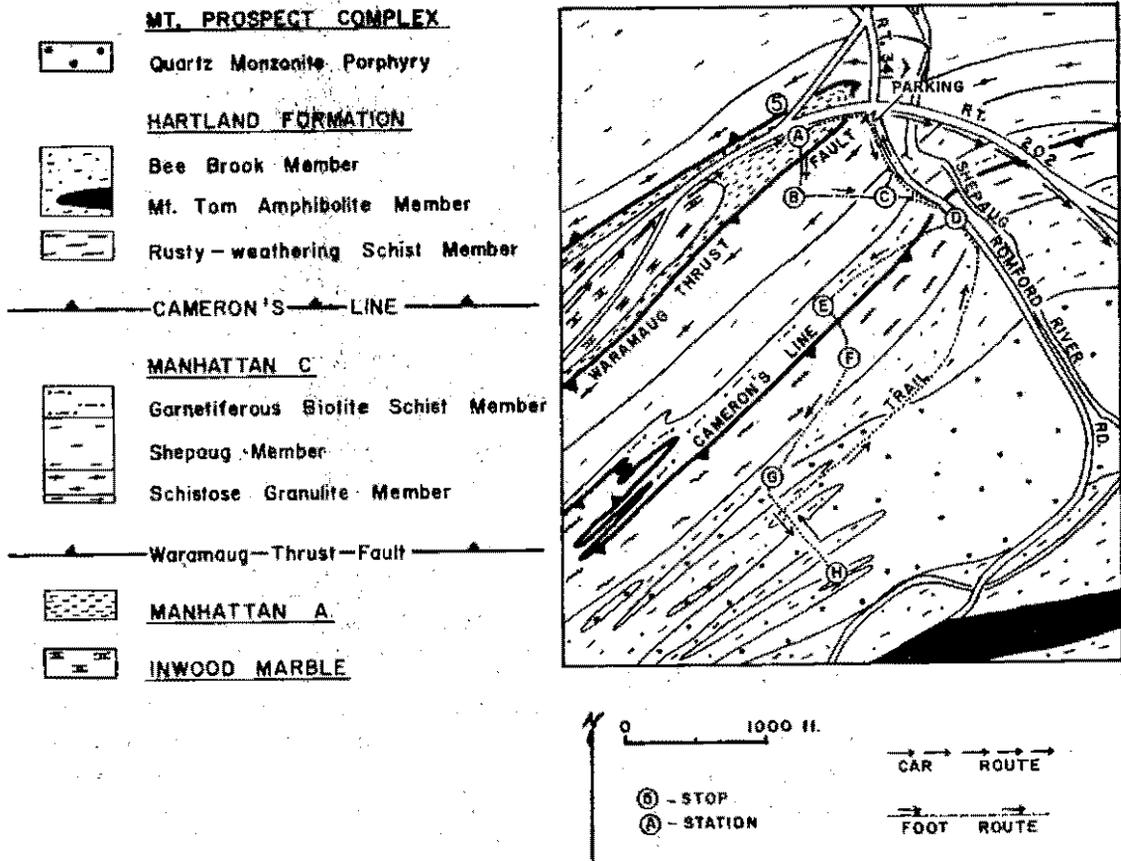


Figure 9: Map of the Stop 5 traverse area located west of the Shepaug River.

The local stratigraphic section is presented in Figure 9. The section, which dips steeply northwest, lies on the overturned, southeast limb of the regional Woodville anticline.

Station A: The Rt. 202 Manhattan A roadcut consists of fine-grained, slabby, rusty-weathering, sulfidic, sillimanite-muscovite-biotite-plagioclase-quartz schistose gneiss and fine-grained, siliceous granulites.

Station B (optional): This is a small outcrop of the Schistose Granulite Member of Manhattan C. It is dominantly a thinly-layered, gritty-surfaced, fine-grained, garnet-muscovite-biotite-plagioclase-siliceous granulite with subordinate schistose gneiss layers. This weathered outcrop has a furrowed, gray-brown surface and a yellow-to burgundy-stained interior. Better exposed outcrops of the Schistose Granulite Member will be seen north of the Mt. Prospect Complex, later in the day.

Station C: Here are pegmatite-rich outcrops of an interbedded, gray, fine-grained, garnet-biotite-plagioclase-quartz schistose granulite and gray, fine-to medium-grained, rough surfaced, garnetiferous, muscovite-biotite-plagioclase-quartz schistose gneiss. White, sillimanite-quartz lenses and <4 mm garnets roughen the surface. Foliation surfaces may be rusty-brown-weathering.

Station D: Proceed southeast along the hillside across Cameron's Line and to two roadside outcrops of the Rusty-weathering Schist Member of the Hartland. Here it consists of thinly-bedded, fine-to medium-grained, yellow brown, to rusty-weathering, garnetiferous, muscovite-garnet-biotite-plagioclase-quartz schist or schistose gneiss. Sillimanite is apparent in coarser varieties.

Station E: Cross Cameron's Line back into the Garnetiferous Biotite Schistose Gneiss Member of the Manhattan C to an exposure of medium-grained, dark-gray, garnetiferous, nubby-surfaced, muscovite-spangled, sillimanite-plagioclase-muscovite-quartz-biotite schistose gneiss. Nubs are sillimanite-quartz pods, lenses, and layers. Garnets may be up to pea size. There are granular, quartz layers <3 cm thick that are boudinaged and lie along the foliation. At most, 5% of the outcrop is poorly foliated, biotite-plagioclase-quartz granulites with patchy mustard-or burgundy-staining, gritty surfaces <25 cm thick.

Station F: Proceed south across Cameron's Line into the Rusty-weathering Schist Member of the Hartland. Here it is a fine-to medium-grained, in places laminated, rusty-brown, sillimanite-staurolite-garnet-muscovite-biotite-quartz-plagioclase schist. Muscovite is concentrated on foliation surfaces. Subordinately present are fine-grained, light-gray, granulites.

Station G: Outcrops of the Bee Brook Member are intruded by sills of quartz monzonite. The dominant country rock types are: thinly-layered, light-gray, fine-grained, siliceous granulite which is massive or laminated with thin white quartz layers, dark-gray, fine-to medium-grained, garnetiferous, sillimanite-garnet-muscovite-feldspar-biotite-quartz schist, and subordinate, fine-grained, white, laminated quartzite. Coarse-grained, massive, quartz layers are also present. A foot thick, concordant, amphibolite layer is present at one place. The main N60E trending Bee Brook Member-Quartz Monzonite contact is approached obliquely as one proceeds 600 feet S45W along the line of outcrops (Fig. 9). The proportion of quartz monzonite increases to the southwest along the exposures. At the northeasternmost of the outcrops, the quartz monzonite is only present as a few <10 cm thick porphyry sills that grade down to thinner medium-grained, wispy, non-porphyrific layers. The southwesternmost outcrop on the traverse, through which the Bee Brook Member-quartz monzonite contact is drawn, is largely quartz monzonite.

Structural Evaluation: The southwesternmost outcrop displays evidence for four phases of deformation in both the country rock and quartz monzonite. The dominant northeast trending foliation, which is the the regional second phase, can be traced from the country rock into the porphyry. In the granulites and schists, the dominant foliation is a muscovite-biotite foliation. The foliation in the quartz monzonite is due to the alignment of biotite and 2-3 mm feldspar grains. Feldspar augen are in the schists adjacent to the igneous contacts. The dominant foliation wraps around the augen. This suggests that the augen and the associated quartz monzonite are older than the formation of the dominant second phase foliation. Since the quartz monzonite is the youngest major intrusive rock of the Mt. Prospect Igneous Complex, then the entire Mt. Prospect Complex was emplaced before or early in the second phase of deformation.

Minor third phase isoclinal folds are present in the country rock. In at least one case a thin non-porphyrific quartz monzonite layer within the schist is folded by third phase folds and a faint foliation is parallel to axial planes of third phase folds.

The ENE-trending regional fourth phase foliation lies at a low angle to the second phase foliation. In the country rock and quartz monzonite this foliation is due to the parallelism of fine-grained quartz and feldspar. Minor fourth phase north plunging "S"-shaped folds are present in the country rock-igneous contacts. The rotation sense of these folds implies a major anticlinal fold to the north, the Woodville anticline.

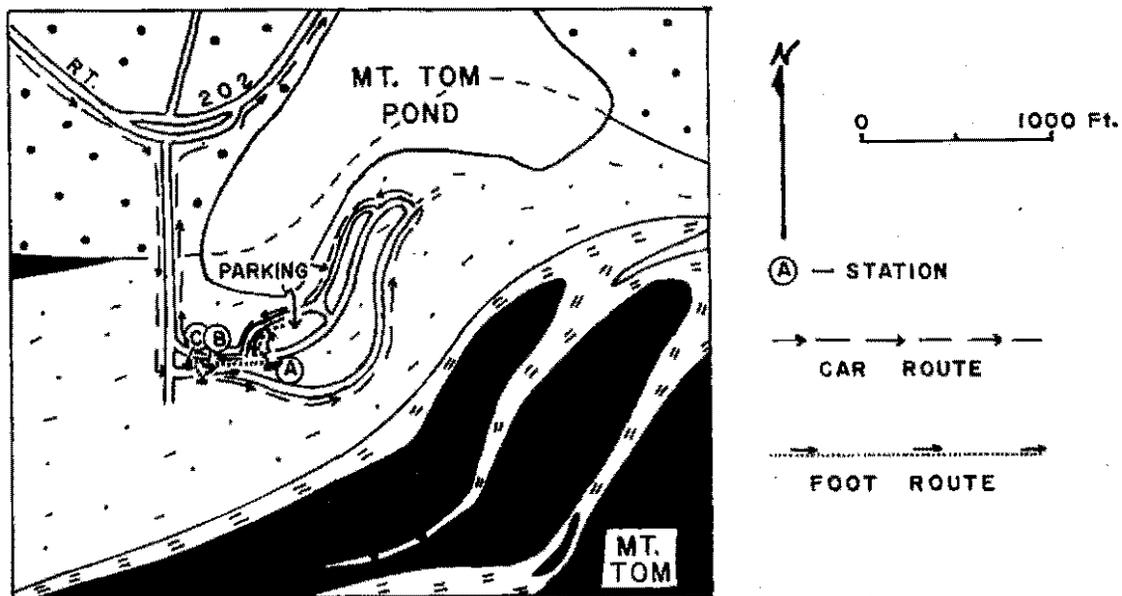
A vague northwest-trending fifth phase foliation is sporadically present. This foliation is a fracture cleavage with recrystallized minerals along it.

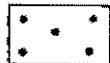
Station H (optional): An outcrop of the Bee Brook Member of the Hartland intruded by quartz monzonite porphyry. The rocks at this outcrop also exhibit structural features that indicate that the quartz monzonite intruded either before or during the development of the dominant foliation. Feldspar porphyroblasts in the schist are probably the product of quartz monzonite contact metasomatism. The dominant foliation wraps around the porphyroblasts and thus probably postdates them. The dominant foliation of the schist continues into the quartz monzonite. The foliation in small Bee Brook Member inclusions in the Quartz Monzonite is continuous with that in the surrounding igneous rock. The inclusions are aligned parallel to this foliation. A wispy non-porphyrific Quartz Monzonite layer in the schist is folded and the dominant second phase foliation is parallel to the axial plane of the fold.

- 4.9 - At the end of the traverse walk SE from Station G or NE from Station H until reaching the path north of the power lines. Follow this path northeast until reaching Romford Road. Walk NW on Romford Road to the vehicles. Drive east on Rt. 202. Manhattan C, Cameron's Line, the Mt. Prospect Formation, and members of the Hartland are crossed in this order over the next 0.9 miles.
- 5.5 0.6 Turn right and follow the signs for Mt. Tom State Park. We cross the folded southwest corner of the Mt. Prospect Complex and enter the Bee Brook Member of the Hartland north of the entrance gate.
- 5.8 0.3 Pass through the entrance and follow the one-way road around a long loop and to the western parking area.
- 6.4 0.6 Park in the western parking area.

-LUNCH-

Picnic tables, fresh water, and toilets are available.



 MT. PROSPECT COMPLEX (undivided)

HARTLAND FORMATION

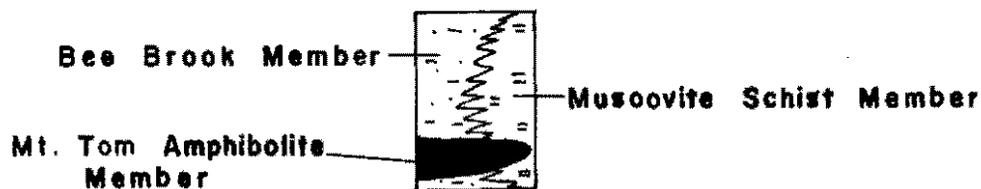


Figure 10: Map of the Stop 6 area located at Mt. Tom State Park.

After lunch walk from the parking lot to the four outcrops of Stop 6.

**Stop 6:** Several roadside outcrops of the Bee Brook Member that display refolded folds involving the first, second, fourth, and fifth phases of deformation are exposed within the Mt. Tom State Park about 400 ft. south of Mt. Tom Pond (Fig. 10). The rock is composed of thinly-bedded, tan-weathering, fine-grained, silvery-gray, staurolite-biotite-muscovite-quartz schist, slabby, gray, biotite-muscovite, siliceous schistose gneiss, and quartzite. Millimeter scale, fine-grained, white to slightly pink, quartz laminae are

present as are coarser-grained, massive, irregular, quartz layers and lenses.

At Station A (Fig. 10) there is evidence in the outcrop for five phases of deformation. A minor first phase isoclinal fold refolded by a second phase fold is exposed on the north facing surface of the outcrop. The dominant schistosity with an average orientation of N 75 E 60 NW (Fig. 11A) is axial planar to the second phase fold. A later foliation has an average orientation of N41E 53SE and consists of fine-grained quartz segregated into paper-thin, millimeter spaced folia. This later foliation cuts across the axial surface of this second phase fold and is axial planar to low amplitude, open folds in the lamination and dominant second phase foliation. This later foliation, which has been observed only south of Rt. 202 (e.g. Stop 8), seems to maintain a SE dip (Figs. 11B, 12, 17), but no known map scale folds are associated with it. This SE dipping foliation is considered to be conjugate to the NW trending fifth phase folds (Fig. 12).

On top of the outcrop fifth phase open folds modify a hook pattern formed by the interference of second and third phase folds.

At Stations B and C, the low angle intersection of very fine compositional layering and the schistosity produces a fine intersection lineation also marked by concentration of biotite and staurolite (Fig. 11A). A closely spaced crinkling of the schistosity seen at B is interpreted to be third phase.

There are two important general features to note among Figures 11, 12, and 13. The poles to bedding and to dominant second phase foliation in the Mt. Tom area including Stop 6 are arrayed in a girdle (Fig. 11C), and the second and third phase lineations have been rotated along great circles to produce a girdle (Fig. 13). These distributions are largely the result of the fourth phase of deformation although the effects of third phase folds, which are interpreted to subparallel the fourth phase folds, are uncertain. A great circle defined by the poles to schistosity around a minor open fourth phase fold at Stop 6 (Fig. 11A) has an orientation similar to the Figure 11C array as demonstrated by the near coincidence of the pole to the schistosity girdle and the axis of the minor fold.

A kink band at B is interpreted to be a fifth phase feature (Fig. 11B). This kink band has a similar orientation to a sporadic, fine-grained NW trending fifth phase foliation seen in the Bee Brook, Muscovite Schist, and Mt. Tom Amphibolite Members of the Hartland Formation (Fig. 12). This foliation is due to thin laminae rich in quartz and feldspar or to the preferred orientation of biotite in the schistose rocks, and to quartz-feldspar laminae or hornblende-quartz-

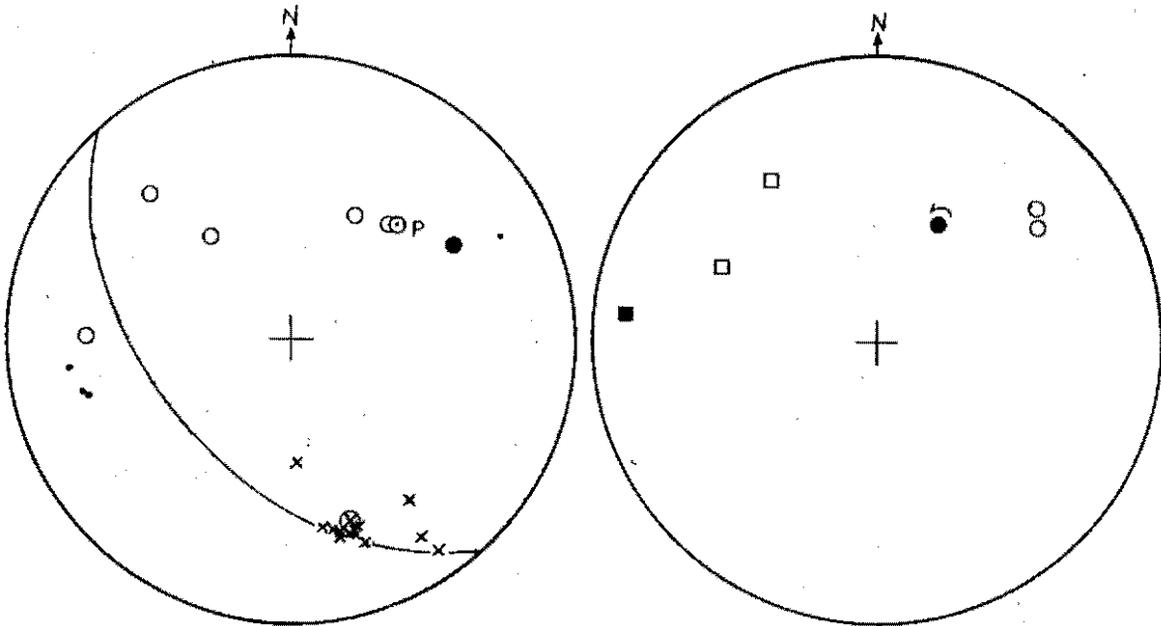


Figure 11A

Figure 11B

Figure 11A: Data from Stop 6, Stations A, B, C.

- Poles to planar features = 14
- × dominant second phase foliation
- ⊗ second phase axial plane
- Lineations = 12
- second phase biotite and staurolite
- lination on the dominant foliation
- minor second phase fold axis
- third phase minor fold or crinkle of the dominant schistosity
- ⊙ axis (N48E 39) to minor fourth phase open fold
- P pole to the girdle of Fig. 11C

Figure 11B: Fifth phase data from Stop 6, Stations A, B.

- Poles to the planar features = 3
- fine, fifth phase, quartz-feldspar foliation
- fifth phase kink band axial plane (N5E 77E)
- Lineations = 3
- intersection of the fifth phase foliation with the second phase schistosity
- fifth phase kink band axis

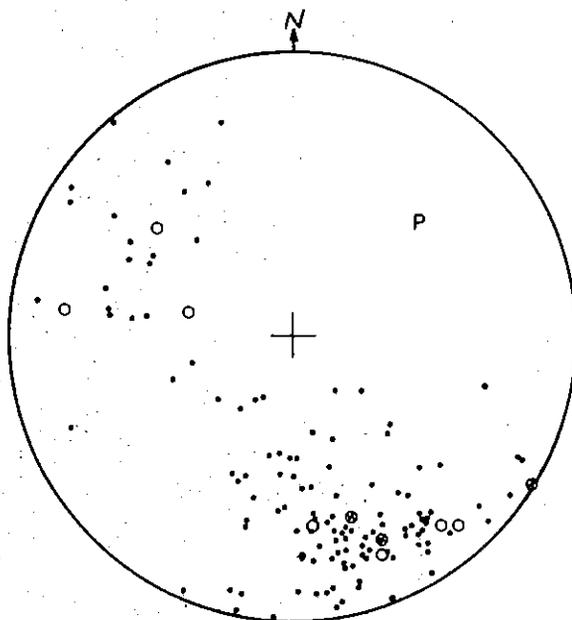


Figure 11C. Foliation data for the Mt. Tom area including Stop 6.

Poles to planar features = 132

○ bedding

• dominant second phase foliation

⊗ axial plane of second phase minor fold

Lineations = 1

P pole (N48E 39) to the girdle defined by the poles to bedding and to the second phase planar features

feldspar laminae in the Mt. Tom Amphibolite. The broad, open, NNW map scale folds in the Mt. Tom Amphibolite at Mt. Tom are fifth phase features (Figs. 2, 10).

- 6.4 - Return to the vehicles, exit the park to the west, and head back to Rt. 202.
- 6.7 0.3 At the "T" intersection before Rt. 202 turn right (east).
- 6.8 0.1 Turn right (east) onto Rt. 202. The road at first sub-parallel the NE trending Mt. Prospect Complex/country

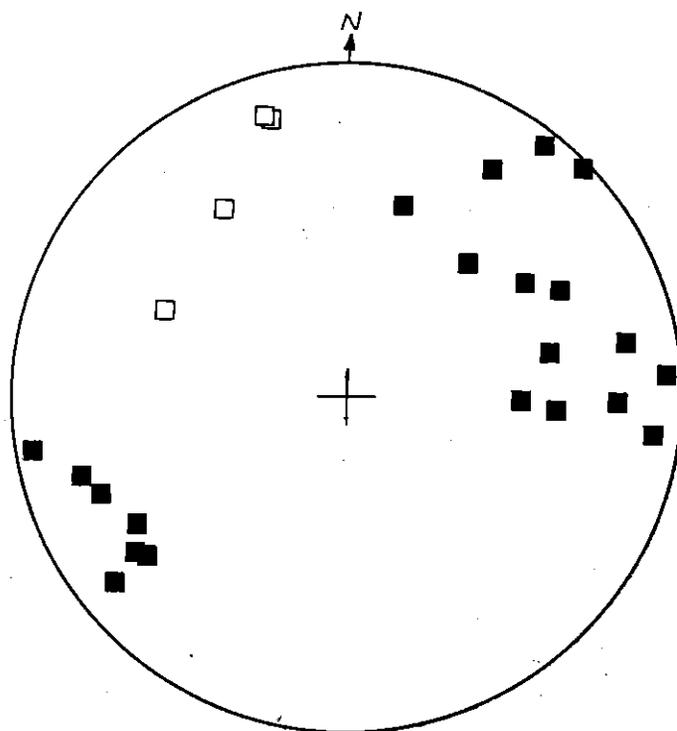


Figure 12 Fifth phase foliation data from the Mt. Tom area

Poles to fifth phase foliation = 25

- foliation of the set with average NW trending strikes
- foliation of the conjugate set

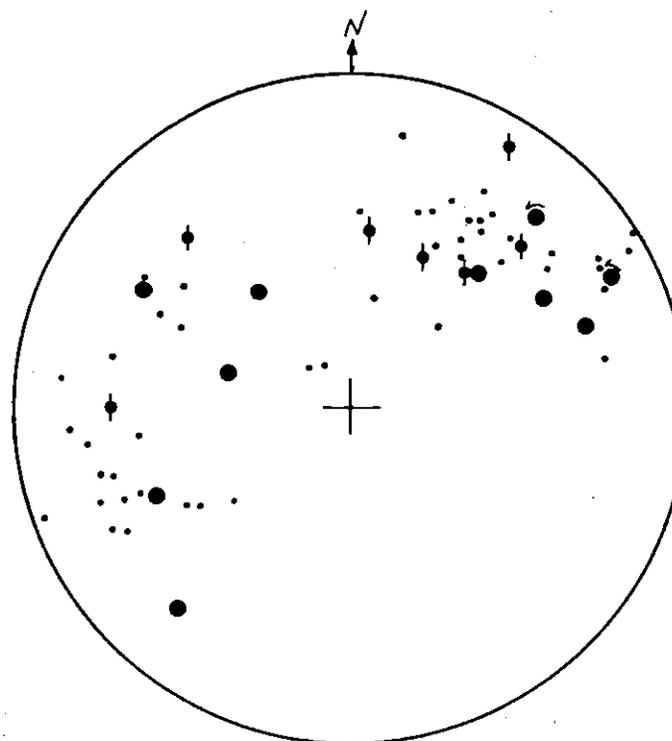


Figure 13 Lineation data from the Mt. Tom area.

Lineations = 35

- second phase mineral lineation or intersection lineation
- second or third phase fold axis
- ◆ third phase crinkle of the dominant second phase schistosity

xenoliths are common in the quartz monzonite. Though not present at this stop, microcline megacrysts are locally present in the diorite, especially near the contacts with the quartz monzonite porphyry. The localization of euhedral megacrysts only near contacts with the late intruding quartz monzonite, and the random orientation of the megacrysts in foliated diorite suggest that the microcline megacrysts in the diorite are porphyroblasts.

A number of foliations are present in this outcrop (Fig. 14). The foliations most prominent are those due to alignment of 1-3 mm anhedral hornblende grains and those due to alignment of 1-3 mm

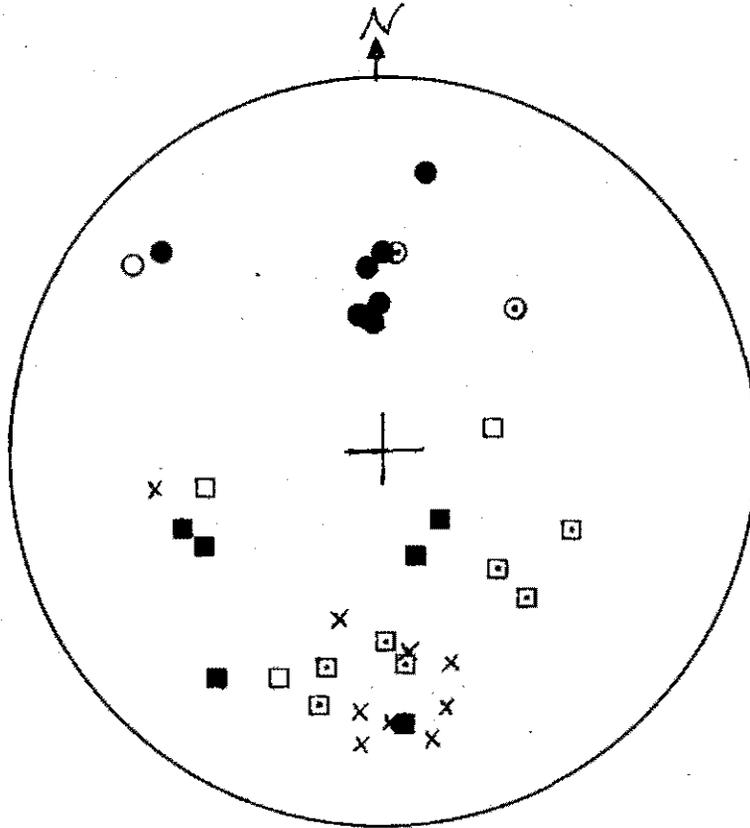


Figure 14: Structural data from Stop 7 and adjacent outcrops.  
 Poles to planar features=25  
 igneous layering  
 hornblende-biotite foliation  
 feldspar foliation  
 cleavage  
 Lineations=10  
 hornblende lineation  
 feldspar lineation  
 intersection of cleavage  
 and igneous layering

rock contact. Primarily diorite and quartz monzonite underlie Mt. Tom Pond to the south, and the Mt. Prospect and Hartland Formations lie to the north. At the NE end of Mt. Tom Pond, the contact trends almost east and crosses the road.

- 7.5      0.7      Continue along a large road cut which consists of Homogeneous Mafic Diorite in contact with the Bee Brook Member of the Hartland. Layered diorite and quartz monzonite are also present. The road recrosses the Mt. Prospect Complex/country rock contact which locally trends NW due to the Woodville Anticline.
- 7.8      0.3      Pass Stop 8 and cross the N trending synform cored by the Mt. Prospect Formation. The low hills to the north and south of the road have diorite on their flanks and the Mt. Prospect Formation at their crests. We will return to Stop 8 after examining the diorite at Stop 7. Continue east along Rt. 202. Pass several diorite outcrops before reaching Stop 7. Mt. Prospect cannot be seen over most of the Rt. 202 route, but it is about 6000 feet north of Stop 7.
- 8.6      0.8      Park in the E. O. Phelps & Sons Co. driveway. Please do not block the access to the buildings. Stop 7 is the large diorite road cut and natural outcrop adjacent to the buildings.

Stop 7: A large, layered diorite gneiss roadcut along Rt. 202. The layered diorite gneiss is the most extensive unit of the Mt. Prospect Igneous Complex. It is composed of typically 1 cm to 5 m thick layers of quartz-plagioclase-hornblende-biotite gneiss locally with minor augite. The layering is due to differences in the relative proportions of the four major minerals, in color of the rocks from light-gray to dark-gray, and in grain size of rocks with this same mineralogy. Layer contacts may be sharp or gradational. Individual layers may wedge out or show low angle cross-cutting relationships with other layers. Mafic xenoliths are common in more felsic hosts but the reverse relationship is rare. Mafic biotite-hornblende lenses are floating in the diorite or locally concentrated along some felsic veins. The former appear to be partially assimilated xenoliths, while the latter may be cumulate patches.

Layered diorite outcrops contain at least minor amounts of intrusive quartz monzonite in the form of veins, irregular, wispy aggregates, discontinuous layers, sills, and rare dikes. Quartz monzonite layers <2 cm thick are typically not porphyritic, but are rich in 2-4 mm, subhedral to euhedral plagioclase grains. Diorite

subhedral to euhedral andesine grains. The hornblende and andesine grains may be randomly distributed in the rock, which is the more general case, or concentrated in wispy lenses. The overall general impression is that these foliations are subparallel to the roughly E-W trending, compositional layers, but in fact low angle intersections among the foliations and compositional layers are common. The layering and foliations together produce the dominant, broadly curved surfaces in the outcrops. Locally the hornblende weathers out to produce either a pitted surface or a crude parting in the rock. Most lineations (Fig. 14) plunge moderately N to NE. This alignment is due to the fourth phase of deformation, and a major fourth phase syncline lies north of Stop 7 (Fig. 2). This lineation orientation roughly parallels the overall plunge of the cylindrical Mt. Prospect Igneous Complex.

There are several, NE-to NW-trending, minor faults in the layered diorite on both sides of the road which locally show an actinolite lineation or slickensides. Displacement is difficult to demonstrate at this outcrop, but these faults postdate the previously mentioned foliations. The faults are in turn cut by randomly oriented, 1 mm wide, felsic veins which may be single and straight or in anastomosing groups.

Foliation that transects the fault consists of small, fine-grained, planar aggregates of feldspar and quartz, and which may be related to the fifth phase of deformation, is sporadically exposed in the large southern outcrop.

- 8.6 - Return to the vehicles and proceed west on Rt. 202 to Stop 8.
- 9.3 0.7 Turn left onto a minor side road which parallels Rt. 202.
- 9.4 0.1 Park along the side of this road just before the road rejoins Rt. 202. We will cross a private yard to get to the Stop 8 outcrops which lie south of Rt. 202. Beside walking through the woods, we will also be in people's back yards and fields so please respect their property and needs.

Stop 8: The purpose of this traverse is to observe a major third phase fold in the diorite-country rock contact.

The map pattern (Fig. 15) shows a north plunging, steep limbed, third phase, isoclinal syncline that is refolded by minor NE-trending, open, fourth phase folds. Poles to compositional layering and second phase foliation in outcrops around the fold hinge define a girdle with a pole plunging at N 1 E 46 (Fig. 16) which is presumed to define the

plunge of the third phase isoclinal syncline. A third phase foliation with an average orientation of N 10 E 82W is axial planar to the major fold. Younger foliations (Fig. 17) have significantly different trends.

Station A: This exposure of the Mt. Prospect Formation consists of interbedded fine-grained, gray, poorly layered, siliceous granulite, gray quartzite and darker-gray, fine-grained, rough-surfaced, garnet-muscovite-plagioclase-biotite-quartz schistose gneiss. Bedding ranges from several centimeters to several meters in thickness. Millimeter scale compositional layering in the schistose gneiss consists of

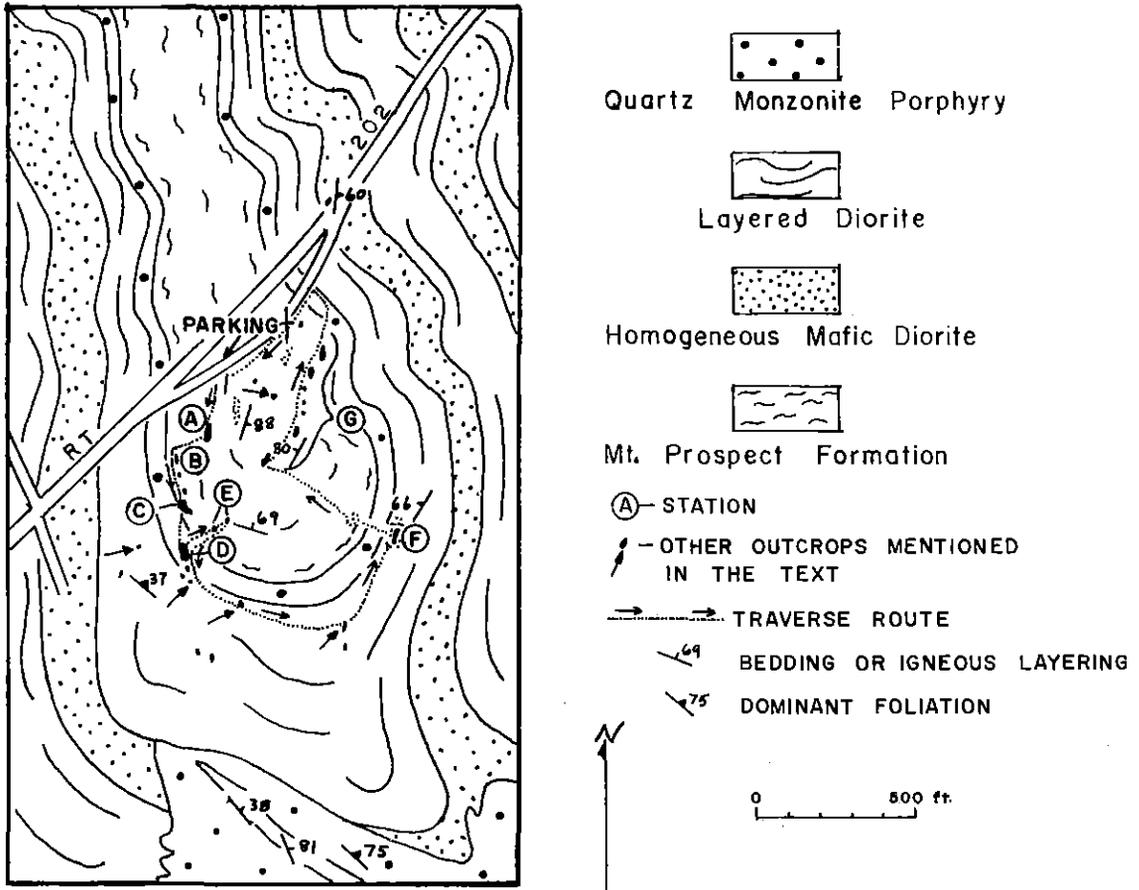


Figure 15: Map of the Stop 8 area showing the third phase fold of the country rock/Mt. Prospect Complex contact.

fine-grained, quartz-rich layers and thinner, biotite-rich layers. The dominant foliation, which is due to the preferred orientation of micas, appears to be parallel to bedding and strikes roughly N-S.

Station B: The Mt. Prospect Formation/diorite contact is exposed here and is subparallel to the NE-trending foliation. The contact and foliation are folded by third phase minor folds.

Station C: This is a fifteen foot high outcrop of fine-grained, well-laminated, gray-white-weathering, plagioclase-garnet-sillimanite-biotite-quartz gneiss. There are two distinct populations of millimeter scale laminations. One consists of smooth, white, fine-grained, quartz layers, and the other is coarser-grained, granular, yellow-staining, irregular, feldspar-quartz layers. The dominant second phase foliation is axial planar to isoclinal folds in the lamination. Also present are 1-10 cm thick, massive, irregular, garnet-quartz layers which may concentrate in 1-2 foot thick zones. The garnet-quartz layers are boudinaged and isoclinally folded and axial plane foliation associated with these folds is due to the preferred orientation of biotite.

Numerous minor third phase isoclinal folds in the layering are obvious and it is possible that several larger isoclinal folds are present in the entire outcrop. Several, late, non-axial planar, faint foliations are sporadically present. One of these, a S-dipping foliation, is probably a fifth phase feature.

Station D: Quartz monzonite porphyry is intrusive into the Mt. Prospect Formation and layered diorite. Foliations are not obvious everywhere in this coarse-grained porphyry but can be identified upon close inspection. The most obvious foliation, due to aligned biotite and small inequant feldspar grains, strikes about N 60 E and dips NW. This is interpreted to be a fourth phase foliation. A weaker, isoclinally folded, second phase foliation is due to aligned biotite. The average N-S trend of this foliation probably represents the trend of the rock units. (Optional): About 250 ft. to the west are small outcrops of layered diorite with NW-trending compositional layering.

Station E: About 100 ft. to the northeast are two small outcrops of the Mt. Prospect Formation with N70W trending laminations. Quartz-feldspar laminae define a fourth phase foliation that trends NE and dips NW. Over the next 1000 ft. of traverse, we will stop at several minor outcrops of diorite in order to demonstrate that the map scale fold closes to the south. We will wind up at Station F.

Station F: Slumped blocks are abundant, but the rock is probably in place in a few places. The diorite layering and a subparallel first phase foliation which is produced by the alignment of feldspar

and hornblende trend N30E and both appear to be cut by a later N-trending second phase foliation. Walking northwest we will cross an area of quartz monzonite and sedimentary rock float, and eventually arrive at Station G.

Station G: A series of outcrops of the Mt. Prospect Formation extend northward along a minor ridge to the road. Along the way the dominant foliation strikes due north on average and has a near vertical dip. Minor fourth phase folds with NE-striking axial planes and NNE-plunging hinge lines are common. These folds have a counterclockwise rotation sense which is related to the major third phase fold that causes the sharp curve to E-W trending folds near the southern border of the Mt. Prospect Complex (Fig. 2).

- |      |     |   |
|------|-----|---|
| 9.4  | -   | Return to the vehicles and turn right (east) onto Rt. 202. Proceed to Stop 9.   |
| 10.2 | 0.8 | Pass Stop 7. Mt. Prospect which is over a mile to the north is largely underlain by olivine norite and quartz norite. Some of the norite bodies are within about 500 feet north of Rt. 202, but none of the late intrusives except for Quartz Monzonite are south of the highway. |
| 11.8 | 1.6 | The long low hill with pastures north of the road is a drumlin. Several drumlins can be seen on both sides of the road over the next two miles. Glacial deposits cover most of the eastern half of the Mt. Prospect Complex. Thus detailed mapping is not possible here.          |
| 14.1 | 2.3 | Turn left at the traffic lights onto Milton Road (which is not named on the 7.5 minute Litchfield quadrangle map). We have crossed the Mt. Prospect Complex over the last 0.1 to 0.3 mile and are now crossing into the Muscovite Schist Member of the Hartland Formation.        |
| 14.4 | 0.3 | Turn right onto Beach Street.   |
| 15.2 | 0.8 | Park the vehicles off the road. This is Stop 9. We will cross the stone fence on the east side of the road and go down the slope where there are abundant outcrops of the Muscovite Schist Member of the Hartland Formation.  |

Stops 9-12: On these traverses we will examine the Muscovite Schist Member of the Hartland, and the Mt. Prospect and Manhattan C Formations north of the Mt. Prospect Complex (Figs. 18, 19). We will cross Cameron's Line several times and demonstrate that it is folded

C3-39

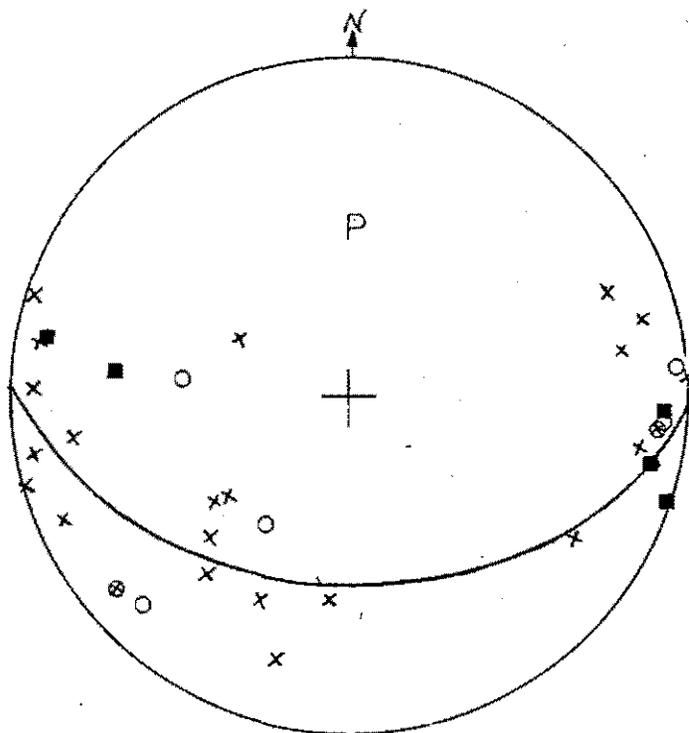


Figure 16. Bedding, second phase, and third phase data from Stop 8 outcrops.

- Poles to planar features = 34
- x bedding
- o second phase foliation
- ⊗ axial plane of second phase fold
- third phase foliation
- Lineation = 1
- P pole (N1E 46) to the plane defined by the poles to bedding and second phase features

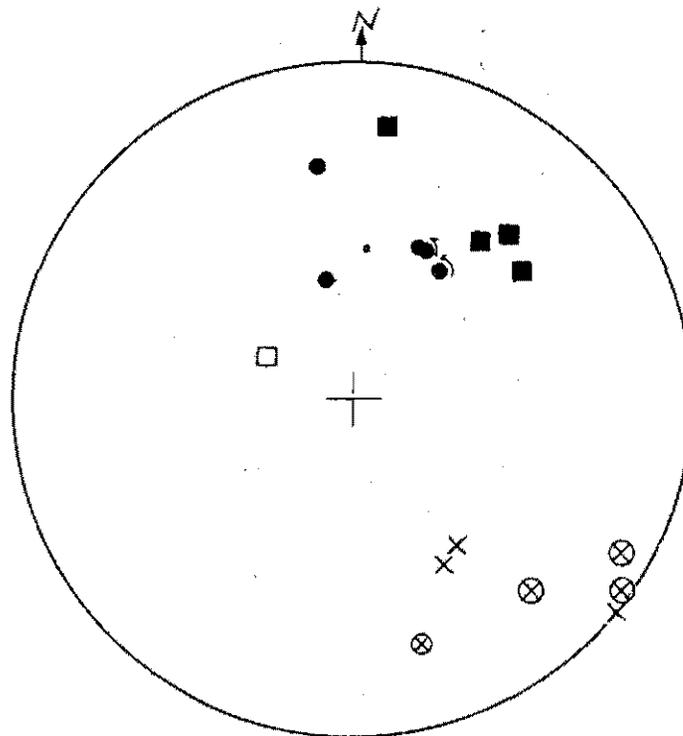


Figure 17. Post-second phase data for Stop 8 outcrops.

- Poles to planar features = 12
- x fourth phase foliation
- ⊗ axial plane of fourth phase fold
- foliation of the fifth phase set with average NW trending strikes
- conjugate fifth phase foliation
- Lineations = 6
- fourth phase fold axis
- intersection of fourth phase foliation and bedding

by second and third phase folds.

Stop 9: Extensive outcrops of the Muscovite Schist Member of the Hartland are on both sides of Beach Street (Fig. 18).

Gray or silver gray, golden-brown-weathering, staurolite-kyanite-plagioclase-chlorite-garnet-biotite-muscovite-quartz schist with millimeter to centimeter scale, fine-grained, sandy, locally laminated, quartz layers, and abundant, coarse, massive, vein-like, quartz layers, lenses, and pods is the most abundant rock. Both types of quartz layers are isoclinally folded. Subordinate, rusty-weathering, dark, biotite-rich schist is present. Rare granitic pegmatites are conformable to the schistosity.

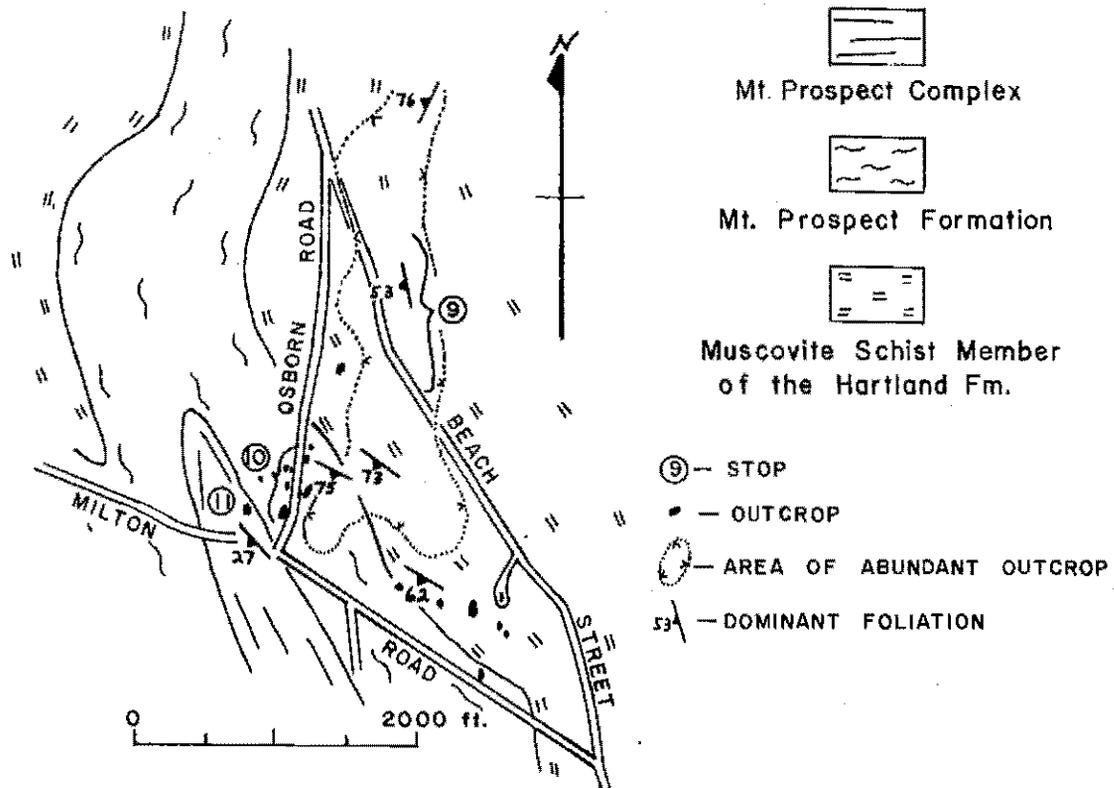


Figure 18: Map of the area around Stops 9, 10, and 11.

A second phase foliation generally trends NW to N. The youngest minor structural features at Stop 9 and 10 are due to the fifth phase of deformation.

Large (<7 cm), biotite-muscovite-garnet-quartz aggregates that are pseudomorphs after staurolite show an imperfect tendency to be aligned in the foliation, and probably are elongate parallel early fold axes. Anhedral, relict staurolite and small euhedral (second generation) staurolites may occur in these pseudomorphs. Quartz rods and fine-grained, elongate, biotite lenses 1-20 cm long are parallel to second phase fold hinge lines and to intersections of compositional layering and second phase schistosity.

One roadside outcrop has abundant muscovite pseudomorphs after andalusite that have rectangular cross sections and are randomly oriented in the foliation.

- 15.2 - Return to the vehicles and drive north on Beach Street.
- 15.4 0.2 Make a sharp (150 degrees) left onto Osborn Road (not named on the 7.5 minute West Torrington quadrangle map and there is no street sign at this intersection) and drive south. We will cross the NW trending Muscovite Schist/Mt. Prospect Formation contact before reaching the Mt. Prospect Formation outcrop at Stop 10.
- 15.9 0.5 Park on either side of the road. The Stop 10 outcrops are along the road over a few hundred feet north of the parking spot. The diorite outcrop of Stop 11 is in the woods about 400 feet west of the parking area. The Muscovite Schist Member is 500 feet east of Stop 10.

Stop 10: At this location the Mt. Prospect Formation consists of medium-grained, dark gray, variably garnetiferous, muscovite-spangled, sillimanite-muscovite-garnet-quartz-biotite schist and various amounts of thin (1-8 cm), light-gray, fine-grained, well laminated, biotite-garnet-quartz granulite layers. The bedding contacts between schist and granulite are nearly parallel to the foliation in the schist, but the granulite layers locally contain wildly contorted laminae. Coarse-grained, granular quartz layers, lenses and pods are isoclinally folded by second phase folds. The quartz layers are down to <1 mm thick to locally give the schistose gneiss a "pin-striped" appearance.

Stop 11 (optional): The most northern known diorite exposure of the Mt. Prospect Complex is a small pavement outcrop (Fig. 18). The outcrop surface is a NW-trending foliation which roughly parallels the

foliation in the Mt. Prospect Formation exposed 250 ft. to the east. This diorite body is interpreted to be in conformable, intrusive contact with the Mt. Prospect Formation and the contact outlines a second phase fold involving the diorite.

- 15.9 - Return to the vehicles and drive south.  
 16.0 0.1 Turn right (west) at the intersection of Osborn Road and Milton Road and continue NW.  
 16.8 0.8 Turn right onto Hutchinson Parkway and head north.  
 17.0 0.2 Park along the road at Stop 12. There are pastures to the west and woods to the east. The first Stop 12 outcrop is a long low outcrop on the east side of the road.

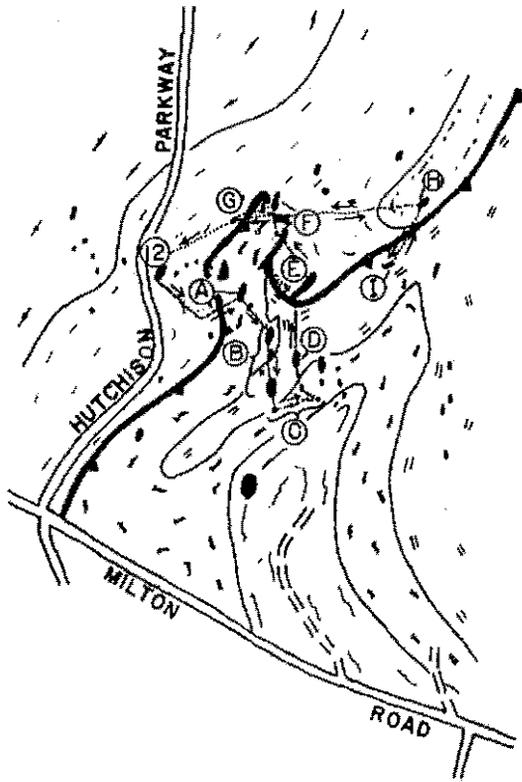
Stop 12: In traveling to Stop 12, we have crossed Cameron's Line. We will now traverse eastward and cross Cameron's Line on foot (Fig. 19). At this locality the map pattern is interpreted to be the result of four phases of deformation. The Cameron's Line thrust juxtaposes truncated Hartland Members against Manhattan C Members. Cameron's Line is subsequently deformed by interfering second, third, and fifth phases of folding. The dominant foliation and associated folds are second phase features. The third phase folds in the foliation also fold Cameron's Line. The open folds are fifth phase features that have an associated SW dipping axial planar crenulation cleavage cleavage.

There is a long, low, roadside outcrop of the Shepaug Member of the Manhattan C at Stop 12. It consists of interbedded, medium-grained, rough-surfaced, garnetiferous, sillimanite-muscovite-garnet-biotite-quartz schistose gneiss, and thinly laminated, fine-to medium-grained, garnet-muscovite-biotite-quartz schistose granulite. The laminae are 1-2 mm siliceous layers and thinner biotite-muscovite-rich layers. Porous clots (<2 cm thick and <15 cm long) in the gneiss are cored by 1 cm diameter garnets or aggregates of small garnets rimmed by sillimanite, quartz, and biotite. The clots are at the noses of minor folds, that together with the SW-dipping crenulation seen here are fifth phase deformational features.

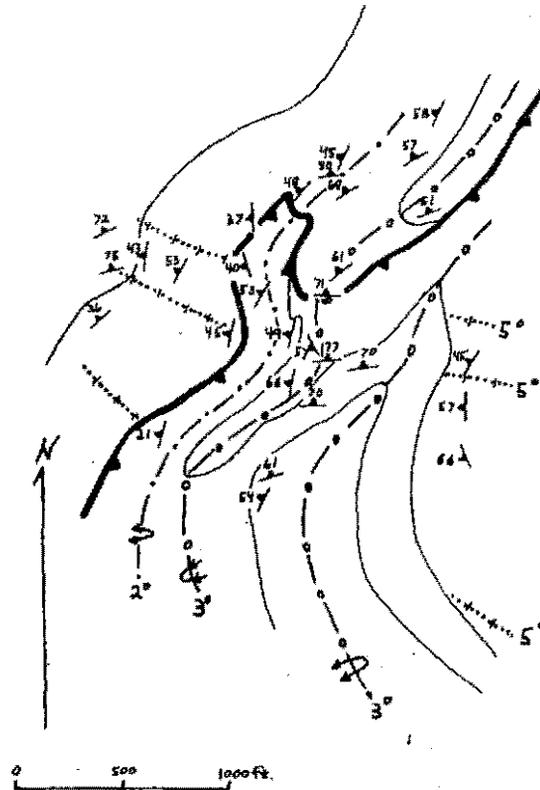
Proceed about 250 ft. S55E along the path to the top of a low hill Station A is about 150 ft. N60E from this point. Cameron's Line, though not exposed, is crossed before reaching Station A.

Station A: An outcrop of the chlorite-biotite-rich schist in the Muscovite Schist Member of the Hartland consists of medium-grained, laminated, garnetiferous, staurolite-garnet-muscovite-biotite-chlorite-quartz schistose gneiss to schist. Aggregates of 1-5 mm

19a. Geologic Map



19b. Axial Trace Map



**Hartland Formation**  
Muscovite Schist Member

-  Muscovite - rich unit
-  Chlorite-Biotite - rich unit
-  Mt. Prospect Formation

 Cameron's Line Thrust Fault 

**Manhattan C**

-  Garnetiferous Biotite Schist Member
-  Shepaug Member
-  Schistose Granulite Member

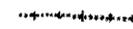
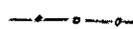
 Outcrop

 Stop

 Station

 Traverse Route

**Trace of the Axial Plane**

-  5° Fifth Phase
-  3° Third Phase
-  2° Second Phase

 50° Dominant Foliation

Figures 19A&B: Geologic and structural data maps of the Stop 12 area.

diameter garnets with quartz and chlorite, and differential weathering of folded foliation may produce a rough outcrop surface.

An attempt has been made, on the geologic map (Fig. 19), to locally divide the Muscovite Schist Member in order to better define the local folds. Thus a biotite-chlorite-rich schist within the Muscovite Schist is distinguished from the normal muscovite-rich schist on Fig. 19A. The outcrop at Station A consists of the biotite-chlorite-rich schist within the Muscovite Schist Member.

Optional Route: Among the features seen at the optional Stations B, C, and D are the Muscovite Schist Member of the Hartland, outcrop size, second phase, isoclinal folds with the axial planar foliation, which is the dominant foliation here, and the axial region of a third phase map scale syncline where the dominant foliation is folded.

Station B is about 200 feet S40E from Station A.

Station B (optional): This is a N-S trending series of four outcrops of the biotite-chlorite-rich schist and the muscovite-rich schist within the Muscovite Schist Member. The average muscovite content increases to the south where the contact between the two subdivisions is crossed.

The large northern outcrop consists of garnet-chlorite-muscovite-biotite-quartz schist to schistose gneiss with 1-20 cm thick beds of light-gray, quartz-rich, schistose granulite and 0.3-1 m thick amphibolite beds. There is an outcrop-scale, southwesterly closing, second phase, isoclinal fold with the dominant foliation parallel to its axial plane and dipping moderately to the west. Plagioclase forms laminae in the amphibolite and these laminae are parallel to the layer contacts. Both surfaces are folded by second phase folds that have an axial plane foliation which is the dominant foliation. The normal muscovite-rich schists of this Member are mapped about 100 ft. to the south. Note that the average foliation in the southernmost, outcrop trends about N20E.

Station C: Several small biotite-chlorite-rich schist outcrops in the Muscovite Schist Member are present in a clearing less than 100 feet south of Station B. Medium-grained, staurolite-garnet-chlorite-muscovite-biotite-quartz schist to schistose gneiss with coarse-grained, quartz-feldspar lenses and layers (<10 cm thick), biotite-amphibolite layers (<3 cm thick) and fine-grained, garnetiferous, laminated, quartz-rich, granulite lenses are present. The granulite lenses have apparently been broken up by shearing along the foliation. The foliated biotite-amphibolite layers are at high angles to both the coarse-grained, quartz-feldspar layers and the younger dominant foliation. The change in strike of the dominant second phase

foliation indicates that the third phase axial trace has been crossed.

Station D is about 150 feet north of Station C.

Station D (optional): This is a large, low, partially slumped outcrop of medium-grained, staurolite-garnet-chlorite-biotite-muscovite schist. The dominant foliation is varied in orientation in the axial region of the southwesterly closing third phase syncline (Fig. 19b) and this and this outcrop is near the axial trace of this third phase syncline. Fifth phase crenulations and minor folds with SW-dipping axial planes are also present.

Station E lies about 250 ft. east of Station A and about 300 ft. north of optional Station D. Cameron's Line is again crossed on the way to Station E from D.

Station E: This is a large outcrop of Manhattan C consisting of light-gray, fine-grained, garnetiferous, sillimanite-chlorite-muscovite-garnet-plagioclase -biotite siliceous schistose granulite. Resistance to weathering by fine-grained, chlorite-sillimanite-biotite-garnet aggregates yields a nubby surface. The orientation change in dominant foliation from W-dipping at Station A to NW-(north end) or N (south end)-dipping at Station E is caused by the third phase, map-scale syncline. Station E lies near the axial trace of this syncline.

We will now traverse about 250 ft. N30W into a pasture to demonstrate the closure of the second phase fold in Cameron's Line.

Station F: This is a group of three outcrops through which Cameron's Line passes (Fig. 19). The southern outcrop consists of medium-grained, staurolite-garnet-muscovite-chlorite-biotite-quartz schist with a 50 cm thick layer of fine-grained, siliceous granulite. The western outcrop consists of muscovite-biotite-quartz schistose gneiss. These two outcrops are the biotite-chlorite-rich schist in the Muscovite Schist Member of the Hartland (Fig. 19). The large northeastern outcrop represents Manhattan C and is a garnetiferous, rough-surfaced, muscovite-biotite-garnet-plagioclase-quartz schistose granulite with subordinate <10 cm thick layers of biotite schist and garnet-hornblende-biotite gneiss.

Cameron's Line trends NW on average through the outcrops while the dominant axial plane second phase foliation trends N30E on average which is parallel to the axial surface of the fold (Fig. 19). Thus this relation between second phase foliation and the axial surface of the fold indicates that this map scale fold is a second phase anticline and that these exposures are in its axial region.

Station G is located about 150 feet N70W from Station F. This outcrop of the Shepaug Member of the Manhattan C consists of light-gray, muscovite-spangled, garnetiferous, chlorite-muscovite-garnet-biotite-quartz schistose gneiss. Individual large garnets or garnet aggregates are surrounded by extensive biotite-chlorite-muscovite-quartz haloes. The dominant foliation dips west. This outcrop further constrains the location of Cameron's Line to the folded shape shown by the second phase fold in Figure 19.

The optional stop in the Garnetiferous Schistose Gneiss Member of Manhattan C at Station H is located about 900 feet east of G. This Member is locally truncated by Cameron's Line.

Station H (optional): The Garnetiferous Schistose Gneiss Member of Manhattan C exposed here is a medium-grained, nubby-surfaced, garnetiferous, sillimanite-muscovite-garnet-biotite-quartz schistose gneiss. "Nubs" are resistant garnets and small sillimanite-quartz lenses. This distinctive, uniform, dark-gray rock, which has relatively few fine-grained granulite layers, is equivalent to that at Stop 5.

Exposures of the Muscovite Schist Member of the Hartland are a few hundred feet south along a minor brook.

Station I (optional): This outcrop consists of coarse-grained, staurolite-garnet-chlorite--biotite-muscovite schist, rusty-weathering schistose gneiss in the stream bed and dark green to black, amphibolite layers. Foliation dips moderately northwest.

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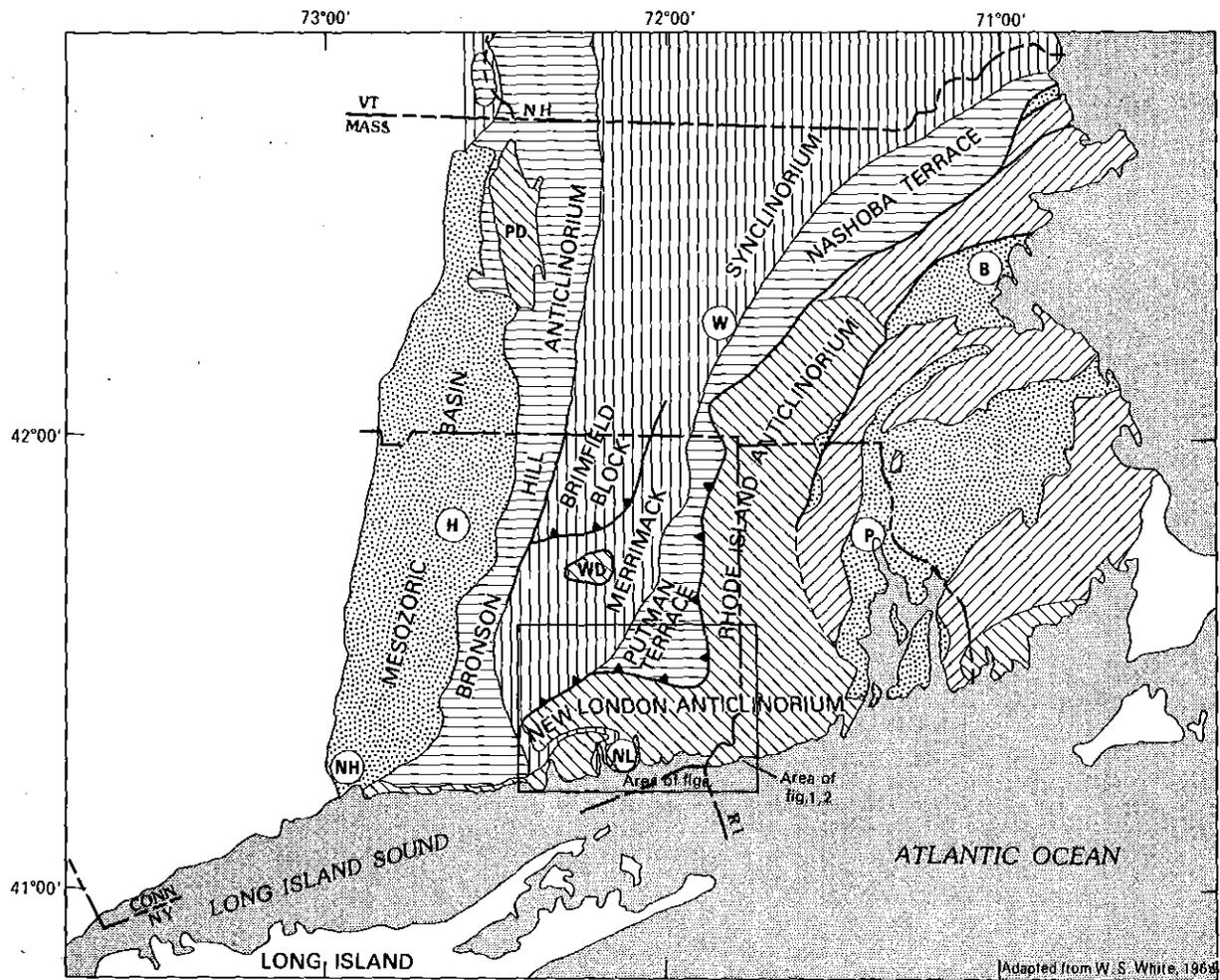
Trip C-4  
HONEY HILL FAULT AND HUNTS BROOK SYNCLINE  
Richard Goldsmith  
U.S. Geological Survey  
Reston, VA 22092

Introduction

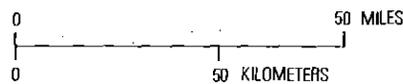
This field trip constitutes a traverse across the New London anticlinorium (fig. 1) from the east flank of the Lyme Dome and the Hunts Brook syncline across the Montville dome to the Honey Hill fault (figs. 2 and 3). The purpose of the trip is two-fold: (1) to observe the transition in metamorphism and deformation from the south limb of the anticlinorium to the north limb where the regional metamorphic fabric seems to pass into the slab of ductily deformed rock which is the Honey Hill fault zone, and (2) to get a partial look at the schist in the Hunts Brook syncline and the repetition of stratigraphic units on either side of it in order to better consider the questions regarding the stratigraphic and structural position of the Schist (from here on referred to as the Hunts Brook schist).

The New London anticlinorium is an east-west trending segment of the Hope Valley terrane of Gromet and O'Hara (1984), part of the Late Proterozoic Avalon platform of southeastern New England. The Honey Hill fault is a major ductile north- to northwest-dipping shear zone that separates the rocks of the New London anticlinorium from the Late Proterozoic and/or lower Paleozoic Tatnic Hill and Quinebaug Formations that form the Putnam terrane. The Honey Hill fault truncates rock units and isograds in the Putnam terrane in the sector between the Preston Gabbro and the Canterbury Gneiss (figs. 2 and 3), but to the west, rock units in the Putnam terrane and Merrimack synclinorium parallel the fault trace. Rock units and isograds in the north limb of the New London anticlinorium are sub-parallel to the fault trace throughout its length.

Rocks in the New London anticlinorium are intrusive and stratified rocks (table 1; fig. 2) containing sillimanite-muscovite and sillimanite-orthoclase metamorphic mineral assemblages. These rocks have undergone at least two stages of deformation (fig. 3). The first produced the pervasive foliation (gneissosity, schistosity) in the rocks which is inferred to be axial planar to large-scale recumbent folds, evidence for which is deduced from the map pattern of units and from rare mesoscopic isoclinal folds that have an axial plane foliation in outcrop. Later folding about steeper axial surfaces deformed the earlier foliation and produced the interference pattern of domes and basins now present (dome stage of deformation). The dome stage, as I have defined it, consists of several subsidiary stages of deformation that had different orientations. In the early part of the dome stage, vergence was toward the south, so that the normal limb of the New London anticlinorium lies to the north and the steep limb to the south. Later folds distorted the steep limb (fig. 3). In the southern part of the anticlinorium, on the overturned limb, a later foliation has formed locally that is axial planar to minor folds that fold the the older foliation. Coarse-grained textures, partial-melting phenomena, movement of blocks of rock past each other along narrow zones without cataclasis or diaphoresis ("s" on fig. 3), and the high-grade mineral assemblages that characterize the dome stage deformation in the southern part of the anticlinorium (Lundgren, 1966; Goldsmith, 1985) indicate that the rocks were deformed at fairly high temperatures. In the northern limb of the



Adapted from W. S. White, 1968



**EXPLANATION**

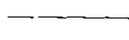
- |   |   |   |                                |
|---|---|---|--------------------------------|
|  | Quaternary  |  | Contact, dashed where inferred |
|  | Mesozoic and Paleozoic sedimentary and volcanic rocks |  | Fault                          |
|  | Metamorphosed Middle Paleozoic and older rocks        |  | Thrust, teeth on upper plate   |
|  | Metamorphosed Lower Paleozoic and older rocks         |   |                                |
|  | Late Proterozoic rocks: gneissic non-gneissic         | <b>B</b>  | - Boston                       |
|   | Granitic rocks not shown separately                   | <b>H</b>  | - Hartford                     |
|   |   | <b>NH</b>   | - New Haven                    |
|   |   | <b>NL</b>   | - New London                   |
|   |   | <b>P</b>  | - Providence                   |
|   |   | <b>PD</b>   | - Pelham dome                  |
|   |   | <b>WD</b>   | - Willimantic dome             |
|   |   | <b>W</b>  | - Worcester                    |

Figure 1. Map of major structural features of southeastern New England, and showing area of figures 2 and 3.

Table 1. Description of rock units in the New London anticlinorium.

Permian	Westerly Granite-- Gray, fine-grained, equigranular granite consisting of quartz, oligoclase, microcline and biotite
	Narragansett Pier Granite-- Light-gray to pinkish-gray, medium-grained to sub-porphyritic granite composed of microcline, calcic oligoclase, and quartz, and 4 to 7 percent biotite. Minor magnetite and allanite.
	Pegmatite-- Orange-pink, white, and gray, zoned, cross-cutting pegmatite.
Ordoevician or Pennsylvanian	Joshua Rock Granite-- Gray, weakly foliated, even-grained, medium-grained, aegerine-augite-bearing quartz-albite-microperthite granite. Contains a rare-earth-bearing sphene and has cherry-red spots of hematite on weathered surface.
	Hunts Brook schist (informal name)-- Garnetiferous sillimanite and orthoclase-bearing quartz-plagioclase-biotite gneiss, and quartz-feldspar gneiss. Locally rusty weathering. Rare layers of calc-silicate rock and amphibolite. Thin white quartzite locally at base.
Late Proterozoic	unconformity?
Late Proterozoic	Alaskite gneiss (equivalent to Hope Valley Alaskite Gneiss)-- Pale orange-pink, light-gray to white, quartz-microcline-albite granite characterized by 1 to 2 percent magnetite or magnetite and biotite. Foliation marked by alternate flat lenses of feldspar and of quartz. Locally may contain muscovite.
	Biotite granite gneiss (equivalent to Potter Hill Granite Gneiss)-- Gray, streaked, biotite-quartz-oligoclase-microcline gneiss. Near margins may contain garnet, sillimanite, and muscovite. Biotite both concentrated in streaks and disseminated. Contains local inequigranular phases and biotite-poor phases approaching alaskite.
	Waterford Group
	Rope Ferry Gneiss-- Lenticularly layered to massive (intrusive phase), mostly even-grained, gray hornblende-biotite-quartz-plagioclase gneiss and, in places, biotite-quartz-microcline-plagioclase gneiss containing scattered lenses of amphibolite. Mafic minerals tend to be concentrated in small clots and streaks.
	New London Gneiss-- Layered facies: alternate layers of amphibolite and light-colored biotite-quartz-plagioclase gneiss, granodioritic in composition. Layers are less than 30 cm to several meters thick. Massive facies: gneissic granodiorite, non-layered, uniform in texture and color. Characterized by shiny black biotite flakes and conspicuous

magnetite grains. Locally contains elongate inclusions and rotated blocks of amphibolite.

Mamacoke Formation-- Upper Cohanzie member (informal name): Light-colored, sugary textured, biotite-quartz-feldspar gneiss containing quartz-sillimanite nodules and greenish-gray calc-silicate gneiss and schist containing diopside and epidote; dark-gray, even-textured biotite-plagioclase gneiss, much of which contains abundant small red garnets and sillimanite; coarse-grained, inequigranular, amphibolite tending to form blocks and lenses in a granitoid matrix; a few thin white layers of quartzite locally at the top. Lower member: indistinctly to distinctly layered, gray biotite-quartz-plagioclase gneiss and subordinate biotite-quartz-microcline-plagioclase gneiss. Magnetite is prominent on weathered surfaces. Contains minor layers of amphibolite and hornblende-bearing gneiss and rare thin beds of quartzite.

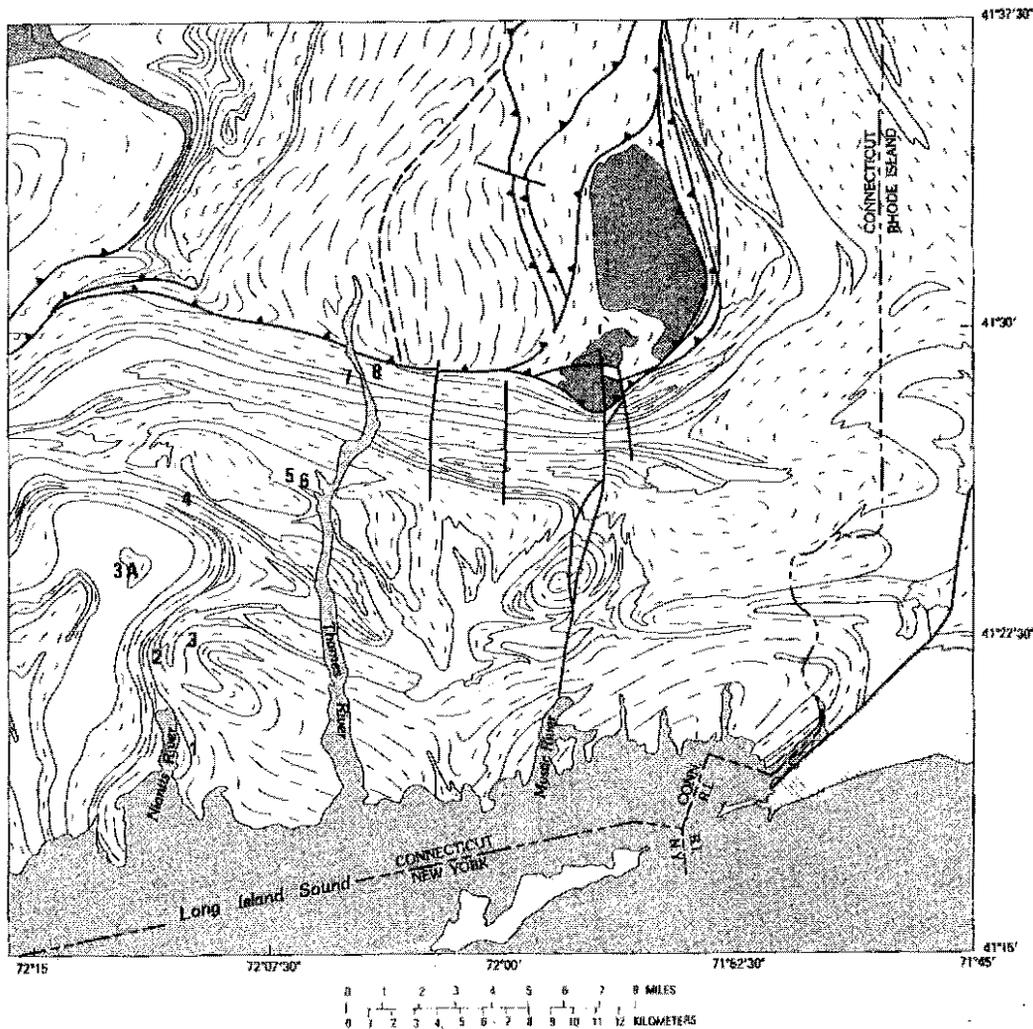
#### Plainfield Formation

Upper member-- In upper part, thin-bedded gray quartzite containing micaceous partings, interlayered subordinate biotite-feldspar-quartz schist and biotite-quartz-feldspar gneiss containing a little muscovite and sillimanite. Beds are mostly 1 to 7 cm thick but as much as 30 cm thick. In the Old Mystic area, grades upward into a gray biotite-quartz-feldspar gneiss containing knots of hornblende that mark the boundary with the Mamacoke Formation. In lower part consists of light-gray quartzite in layers and elongate lenses 60 cm to 1 m thick, a few layers of white to light-green diopside-bearing calc-silicate quartzite.

Middle member-- Gray biotite-quartz-feldspar gneiss containing hornblende and diopside, calc-silicate quartzite and gneiss, amphibolite, garnetiferous schist, sugary-textured sillimanite-bearing biotite-quartz-feldspar gneiss, and thin beds of white to gray quartzite.

Lower member-- Light-gray quartzite in beds 15 cm to 1 m thick that contain sillimanite and biotite partings and are interbedded with sillimanitic and non-sillimanitic mica schist and mica gneiss. In Lyme dome, consists of thick sequences of micaceous to feldspathic quartzite that contains quartz-sillimanite nodules and thick sequences of pelitic schist and gneiss as above.

Late Proterozoic



EXPLANATION

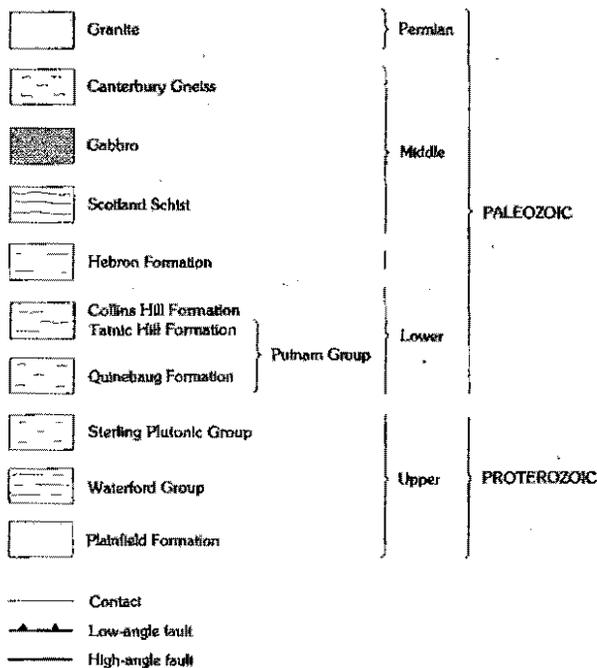
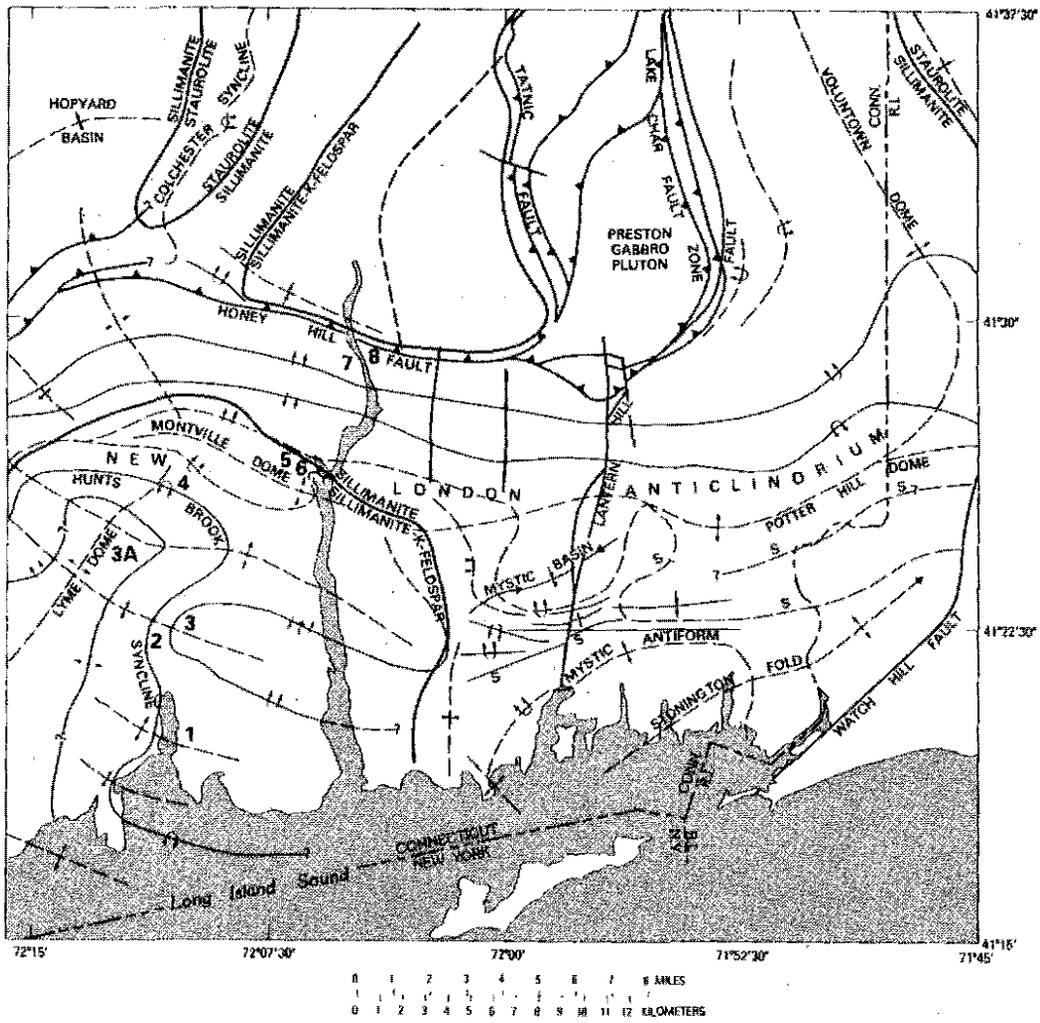


Figure 2. Geologic sketch map of southeastern Connecticut and adjacent Rhode Island and New York, showing numbered locations of field trip stops. Adapted from Goldsmith (1985).



EXPLANATION

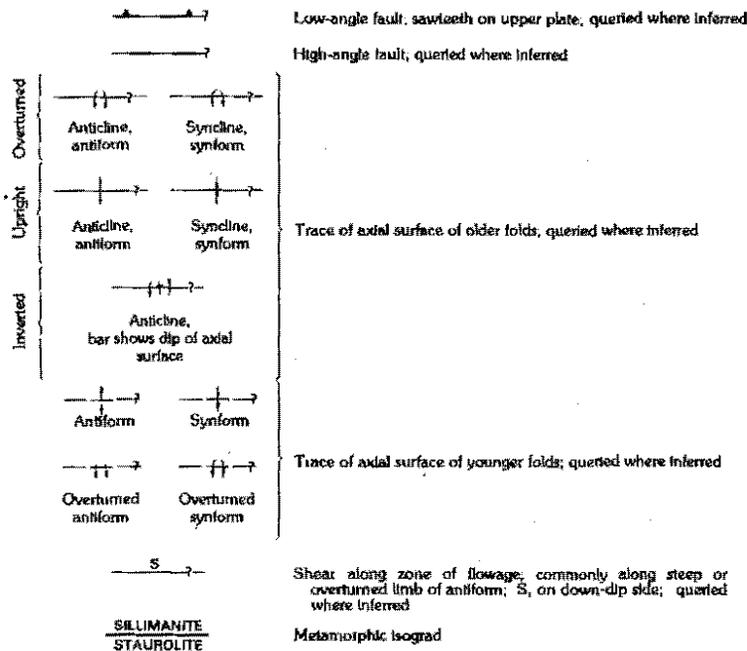


Figure 3. Map of structural features of southeastern Connecticut and adjacent Rhode Island, showing generations of fold structures and numbered locations of field trip stops. Adapted from Goldsmith (1985).

anticlinorium, the rocks are at amphibolite facies but are finer-grained and appear to have blastomylonitic textures. In the Honey Hill fault zone, the former high-grade mineral assemblages predominantly those in rocks of the Tatnic terrane, have retrogressed to greenschist facies assemblages (Lundgren and others, 1958; Snyder, 1964; Lundgren and Ebblin, 1972; O'Hara and Gromet, 1983).

Several workers have studied the timing of these events. The retrogressive greenschist facies metamorphism along the Honey Hill fault has been dated as Permian (O'Hara and Gromet, 1983). The Joshua Rock Granite (table 1) which has been dated as Late Pennsylvanian or early Permian by R. E. Zartman (oral communication, 1981), is involved in the doming-stage folding and the flattening in the thinned section between the Lyme Dome and the Montville Dome along Hunts Brook (Stop 4, fig. 3). The dikes of Westerly Granite, dated as Early Permian (Kocis and others, 1978), clearly cut all earlier fabrics. Pink, zoned pegmatite believed to be related to the Westerly and Narragansett Pier Granites cuts blastomylonitic rocks in the Honey Hill fault zone. These observations constrain the events between the dome-stage folding and the final movement on the Honey Hill fault to a very narrow interval. Furthermore, they indicate that rapid uplift and northward tilting of the New London anticlinorium occurred in the Permian. However, the time of pre-dome stage deformation, isoclinal folding, and formation of the foliation is uncertain. All also may have to be Permian. South of the Honey Hill fault, there is no evidence of a metamorphic event between the Late Proterozoic and the Permian (M. H. Pease, Jr., written communication, 1982). In the Putnam terrane above the Honey Hill fault, there are indications of metamorphism and pegmatite intrusion that are Middle Ordovician or older, and Acadian (Devonian) metamorphic and plutonic events have occurred in the Putnam terrane and the terrane of the Merrimack synclinorium to the west (Zartman and others, 1965). We do not know to what extent, if at all, the Hope Valley terrane that is exposed now in the Willimantic dome was involved in the Acadian metamorphism. As mentioned above, one of the aims of this trip is to observe the change in textures and style of deformation across the New London anticlinorium and to see if it is correct to say that the Honey Hill fault zone represents the localized and final stage of movement at a shallow level of a more pervasive and deeper-seated deformation that began earlier.

The other topic of this trip is the nature of the Hunts Brook syncline (or synform?) and the contact between the Hunts Brook schist and the underlying Waterford Group. I originally mapped the Hunts Brook schist as Brimfield Schist because this pelitic unit could be traced into the Montville and Niantic quadrangles from the Essex area where it was mapped by Lundgren (1962) as Brimfield Schist (fig. 4). This unit was at the time equated with the Tatnic Hill Formation and we believed the Tatnic Hill could be traced continuously into and around the Chester syncline into the Brimfield Schist on the flanks of the Bronson Hill anticlinorium (Rodgers, 1985). Recent mapping by Wintsch (1979; Wintsch and Kodidek, 1981) however, has brought into question the continuity of the Brimfield with the Tatnic Hill around the syncline. Furthermore, the plagioclase gneisses that lie below the Brimfield on the flanks of the Bronson Hill west of the Chester syncline appear to be Ordovician in age (Zartman and Naylor, 1984), whereas the plagioclase gneisses south of the Tatnic Hill and the Honey Hill fault are Late Proterozoic in age (Pignolet and others, 1980). The two sets of plagioclase gneisses differ in compositions according to Wintsch, 1980). Nevertheless, I believe that the

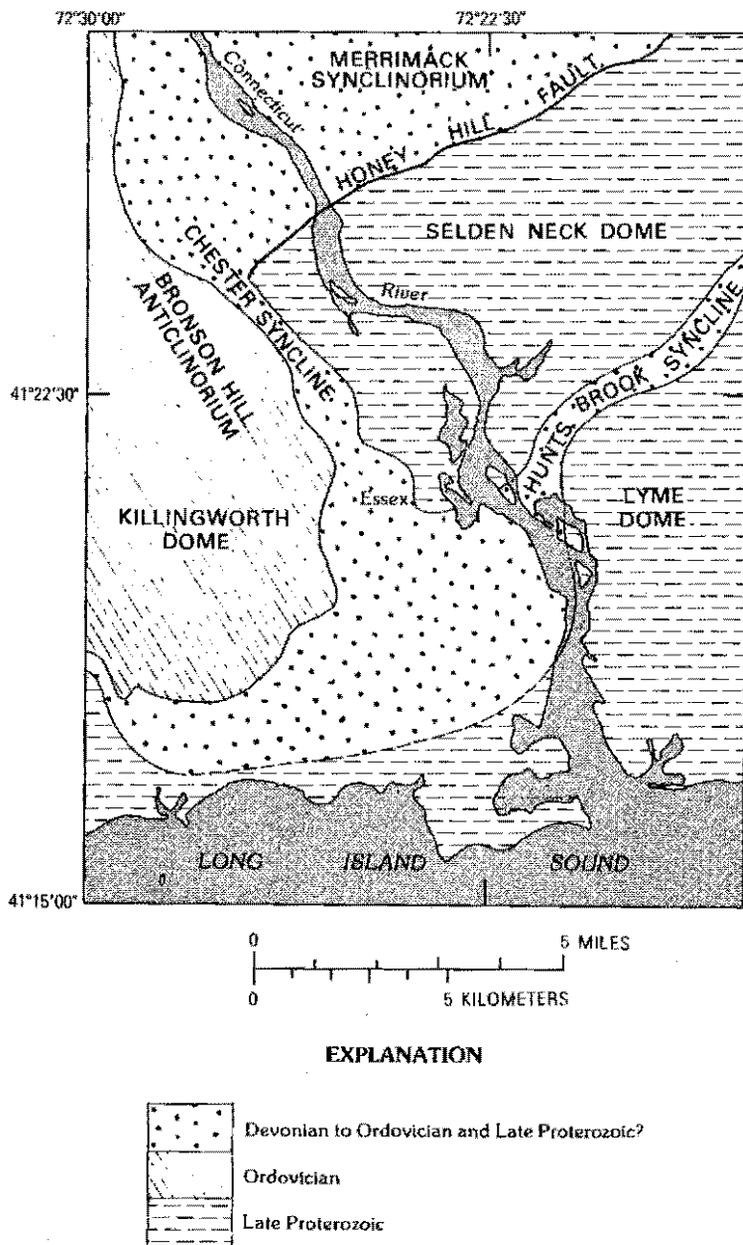


Figure 4. Major geologic features in the area of the Chester syncline, southeastern Connecticut.

Schist in Hunts Brook is in the same structural position and probably stratigraphic position above the Waterford Group as is the Tatnic Hill. In this case, the contact of the Hunts Brook schist with the Waterford Group in the Hunts Brook syncline should represent the same horizon as the Honey Hill fault on the north limb of the anticlinorium. However, one could argue that the Honey Hill fault has cut up section above the Tatnic Hill and that this contact was before metamorphism either a normal stratigraphic contact in a continuous sequence of Late Proterozoic rocks, or was an unconformity. In the latter case, the Hunts Brook schist could be equivalent to the Tatnic Hill or the Brimfield, or both. If both, a major break must lie beneath the schist because it would rest on both Bronson Hill Ordovician plagioclase gneiss to the west and Late Proterozoic Waterford Group plagioclase gneiss to the east. If not, probably there are two schists in two different tectonic blocks, in which case the Hunts Brook could be part of a Late Proterozoic sequence. If so, the Hunts Brook schist is of different composition than the pelitic rocks of the Plainfield Formation which tend to be relatively more quartzose than the Hunts Brook and to contain tourmaline. There is a schist in the Essex area of the Chester syncline (R. P. Wintsch, this guidebook) which lies above Late Proterozoic plagioclase gneiss that greatly resembles the Rope Ferry Gneiss, and this schist must be continuous with the Hunts Brook Schist as originally mapped. On this trip we can not resolve the problem as to the identity and correlation of the Hunts Brook schist, but we can look at the Hunts Brook and its contact and see if it can tell us anything. I am inclined to favor the position that it is a metamorphosed unconformity because of the fairly abrupt change in rock type, because of rare lenses of quartzite that lie along the contact, and because of lack of evidence for faulting.

#### ROAD LOG AND STOP DESCRIPTIONS

The trip area is covered by the Uncasville, Montville, Niantic and New London 7-1/2 minute quadrangle maps. The geology of these quadrangles is published in U. S. Geological Survey Geologic Quadrangle Maps GQ's 574, 575, 576, and 609 (see references).

The assembly area for the trip is the parking lot of the Connecticut Yankee Travelodge located on Route 161, 50 yards south of its intersection with I-95 at Flanders. From I-95 take exit 74 south towards Niantic. The Travelodge is on the right. The trip may be run either from north to south or south to north depending on the time of high tide that day. The log will be presented as from south to north. If the trip is run from north to south, the first stop can be reached via I-95 to I-395 to Route 2A south of Norwich and thence as directed from Stop 7 to Stop 8.

Miles		
cum.	int.	
0	0	From Travelodge parking lot, turn right (south) on Route 161.
1.7	1.7	Bear right at second traffic light onto Pataguansett Road.
2.6	0.9	Intersection of Pataguansett Road and Route 156. Turn left (east).
2.7	0.1	Turn right (south) on McCook Point Road.
3.1	0.4	McCook Point. Turn left into parking lot.

STOP 1 (30 minutes) Hunts Brook schist (informal name) at McCook Point.  
Outcrops at east end of the beach are Hunts Brook Schist in the trough of the

Hunts Brook syncline. The Hunts Brook consists of quartz, plagioclase, biotite, sillimanite, garnet, and minor amounts of potassium feldspar. Note the garnet sand on the beach. No muscovite is present. The rock is physically a gneiss and is within the second sillimanite zone of metamorphism. The protolith of the Hunts Brook is considered to be a sequence of layered pelite and semi-pelite. In places, but not here, a thin quartzite is exposed at its base. The layering is considered to be transposed bedding accentuated by metamorphic differentiation. Unfortunately I have few modes and no chemical data for the Hunts Brook schist. The syncline has been folded along a series of east-west trending late folds (fig. 3). Minor folds plunge to the east. The major structure is presumed to plunge north like the Montville fold, although the axis of the Hunts Brook synform has most likely been rotated by the later folding. The minor folds in these outcrops are of the doming stage and fold the foliation. Traces of axial surfaces vary from N65E to N80E, and plunges of axes are mostly steep to the east. Short limbs of some folds have been sheared off. Note also boudinaged layers that have steep axes. The pegmatite dikes that cut the deformed gneissic fabric are abundant in the coastal area and are considered to be Permian in age and contemporaneous with the Narragansett Pier and Westerly Granites.

- Return to cars and proceed north (right) from parking lot.
- |     |     |  |
|-----|-----|--|
| 3.3 | 0.2 | RR overpass, must turn right, then left.         |
| 3.5 | 0.2 | Intersection Route 156. Turn left (west).        |
| 3.7 | 0.2 | Turn right at light onto Pataguansett Road.      |
| 4.7 | 1.0 | Turn left at traffic light onto Route 161.       |
| 6.5 | 1.8 | Enter I-95 northbound ramp.                      |
| 7.0 | 0.5 | Exit Route US 1.                                 |
| 7.3 | 0.3 | Cross I-95                                       |
| 7.7 | 0.4 | Turn sharp onto River Road before second bridge. |
| 7.8 | 0.1 | Go to end of road and park in pull-off.          |

STOP 2. (1 hour) Contact of Hunts Brook schist and Ropes Ferry Gneiss on east limb of the Lyme Dome. - Climb bank next to last house up to trail and proceed south (about 15 minutes) along trail past intermittent outcrops of hornblende-biotite gneiss (Ropes Ferry Gneiss) considered to be a meta-plutonic rock. Large ledges to the right are Ropes Ferry Gneiss, New London Gneiss, and transition to Mamacoke Formation cut by a dike of Westerly Granite. Near the large ledges, veer left, where marked, to beach and shore of the Niantic River. Follow along the shore (difficult if tide is high) and pass into layered rocks forming the contact zone of Ropes Ferry Gneiss and Hunts Brook Schist. The contact zone contains thin layers of dark- and light-colored rock, garnet-biotite gneiss, and amphibolite pods. Ropes Ferry Gneiss is exposed in the bank at one place. At the furthest available outcrops are discontinuous quartzite layers. Similar rocks, except for the quartzite, are exposed on the east side of the synform across the Niantic River. Axial surfaces of early dome-stage minor folds trend northerly and are inclined less steeply to the east than is the layering. Axes plunge to the north, a different orientation than the minor folds at Stop 1.

The options for the pre-metamorphic nature of the contact as discussed above are: (1) a ductile shear zone, (2) an unconformity of non-conformity, or (3) a conformable contact in a Late Proterozoic sequence. I see no evidence for (1) and for the reasons given above, I prefer option (2). In this option, the Hunts Brook could be Late Proterozoic or early Paleozoic in age through

correlation with schist in the Chester syncline area to the west or the schists in the terrane north of the Honey Hill fault.

Return via trail to cars.

- 7.9 0.1 Return to US 1, turn right (east).
- 8.0 0.1 Cross bridge and turn left (north) onto Oil Mill Road.
- 8.5 0.5 Turn right (east) just beyond I-95 overpass onto Parkway North.
- 9.1 0.6 Stop at road cuts at top of hill.

STOP 3. (20 Minutes) Alaskite Gneiss, Westerly Granite, and Waterford Group - Plainfield Formation transition zone. - Outcrops at road level are alaskite gneiss equivalent to Hope Valley Alaskite Gneiss. On the south side of the road, the alaskite contains inclusions of biotite gneiss of the Mamacoke Formation, and at one point, the alaskite is cut by a characteristically gently dipping dike of Permian Westerly Granite. On the west side of the outcrop on the bank towards the Interstate, the contact of the alaskite with quartzite in the transition zone between Plainfield Formation and the Waterford Group (Mamacoke Formation) is exposed. We are on the east (and north) side of the Hunts Brook synform near the nose of a pre-dome stage anticline (fig. 3). The anticline is a tightly refolded anticline containing biotite gneiss in its core flanked by Plainfield Formation, alaskite gneiss, and further outward by rocks of the Waterford Group. By climbing down the roadway--be careful--one can see that the alaskite clearly transects the stratified rocks and provides a minimum age for the Plainfield and Waterford. The alaskite gneiss is correlated with the Hope Valley Alaskite Gneiss in Rhode Island which has been dated by U-Pb isotopes in zircon in northeastern Connecticut and southeastern Massachusetts at 630 m.y. (Zartman and Naylor, 1984). The gneissic fabric (foliation) in the alaskite is co-planar with the regional foliation which has been folded during the doming stages. The Early Permian Westerly Granite lacks this gneissic fabric and cuts folds of the dome stage, thus giving a minimum age for the doming.

- 9.7 0.6 Return to Oil Mill Road. Turn left (south).
- 10.2 0.5 Intersection of US 1, turn right (west).
- 10.7 0.5 Cross I-95. Light colored outcrops to right are alaskite gneiss. Continue on US 1 toward Flanders.
- 11.1 0.4 Outcrops to right are biotite granite gneiss of the Sterling Plutonic Suite equivalent to Potter Hill Granite Gneiss in Rhode Island. Foliation is nearly vertical. The biotite granite gneiss and the alaskite gneiss are on the west limb of the Hunts Brook synform (east limb of the Lyme dome). The valley to the west of the ledges is occupied by sillimanite-bearing mica gneiss of the Plainfield Formation. The granite gneiss contains rare prisms of sillimanite on foliation surfaces which I believe are the result of assimilation of wall rock material by the magma during emplacement and slow cooling.
- 11.3 0.2 Intersection with Route 161. Turn right (north).
- 11.6 0.3 Outcrops of biotite granite gneiss.
- 13.9 2.3 Outcrops of gneissic biotite granite well within the Lyme dome.
- 15.7 1.8 Turn right (east) onto Westchester Drive. Bear right to circle.

STOP 3A (15 minutes) Quartz-sillimanite nodules in feldspathic quartzite of the Plainfield Formation. - The lower member of the Plainfield Formation in the Lyme dome consists of two primary assemblages: a thick pelitic gneiss which outcrops to the south of this locality, and this unit of feldspathic

quartzite containing nodules of quartz and sillimanite. The protolith was probably a clayey sandstone, perhaps somewhat feldspathic. At lower metamorphic grade, this rock would probably be a muscovitic quartzite. We are in the sillimanite-potassium feldspar zone. I believe a metamorphic, apparently concretionary process at this high grade of metamorphism has produced these lenses and nodules. Elsewhere in the area, the nodules have formed in a granitoid matrix (see Stop 6) so that the presence of these nodules in this rock is somewhat unusual. Quartz-sillimanite and quartz-sillimanite-muscovite nodules are common features of high-grade metamorphic terranes in which rocks of pelitic and psammitic composition are present. They have been reported in Ireland, Norway, Finland, India, Canada, Connecticut, New York, and probably elsewhere. J. A. W. Bugge (1943) gives a fairly complete discussion of the nodules and ascribes them to migmatization (the vogue in those days) in a zone of aluminous rocks. The nodules in the New London area usually are ellipsoidal, flattened in the foliation, and are elongated down dip in the direction parallel to the plunge of mineral lineation and fold axes in the area. At one place, however, the a-b plane of the ellipsoids dipped more steeply than the foliation.

Return to Route 161.

- 16.3 0.3 Turn right (north) on Route 161. Ledges in this area are all nodular quartzite of the Plainfield Formation.
- 17.6 1.3 Intersection of Route 85 at traffic light. Turn sharp right (south) on Route 85.
- 17.9 0.3 Turn left (east) on Turner Road.
- 20.0 2.1 Intersection with East Pond Road. Turn left (north).
- 20.8 0.8 Outcrops on right are Ropes Ferry Gneiss on south limb of Hunts Brook synform in the Lyme Dome, the same rock seen at Stop 2.
- 20.9 0.1 Turn right (east) on Fire Street.
- 21.1 0.2 Park near entrance to gravel pit.

STOP 4. (2 hours) Section across Waterford Group on the north side of Hunts Brook Syncline, Montville dome. - At this stop we will traverse from the top of the Waterford Group almost to the Plainfield Formation. The Hunts Brook schist is not exposed but is presumed to occupy the valley in which the brook flows.

From the cars, take the dirt road leading to the gravel pits, bear right towards the end of the first open pit, and climb the hill following the flags along a chain of outcrops. The ledges up the slope and to the top are Ropes Ferry Gneiss. The lowest and first outcrop is layered and is in the contact zone between the Ropes Ferry and Hunts Brook schist as at Stop 2. Climb over or walk around a large continuous ledge of Rope Ferry Gneiss. Climb to the top of the ledge at its end and bear N40-50E to a ravine. At the top of the ravine and on the slope down are interlayered amphibolite and light-colored rocks of granodioritic composition (New London Gneiss). On the north side of the ravine, bearing N50E, is a ledge of gneissic aegerine-augite granite (Joshua Rock Granite). The Joshua Rock is present only on the north side of the Hunts Brook syncline. It has not been identified in the Lyme dome. As the Joshua Rock is Pennsylvanian or Permian, whereas the adjacent Waterford Group is Late Proterozoic, the Joshua Rock is not part of the Waterford Group as I had originally proposed (Goldsmith, 1976). As mentioned above (Stop 3), the Joshua Rock is involved in the dome-stage folding and has a weak gneissic fabric. It is thus an intrusive rock older than the Westerly Granite. It forms a sill following a particular horizon between the New London Gneiss and

the Mamacoke Formation. It does not appear to transect other units. A somewhat similar alkalic rock lies south of the Honey Hill fault in the Fitchville quadrangle (Snyder, 1964).

Bear right around the ledge of Joshua Rock and thence over the top, bearing left towards the power line along a ledge of Mamacoke Formation. The rock here is a compositionally varied member of the Mamacoke Formation, informally called the Cohanzie member (table 1). First is a rock containing amphibolite blocks and lenses in a granitoid matrix. North of this rock is a distinctive dark-gray, garnet-rich, sillimanite-biotite-quartz-feldspar gneiss. North across the trail along the power line are two other distinctive marker units. One is a calc-silicate layer that contains epidote, diopside, and calcic plagioclase; the other is a granitoid rock containing quartz sillimanite lenses. Angle right to the trail along the power line and follow it east into gray biotite gneiss typical of the bulk of the Mamacoke Formation. Follow the trail to light colored ledges of alaskite gneiss (Hope Valley Alaskite Gneiss equivalent). The ledges to the northeast across the wide valley are also alaskite gneiss. Thin slivers of quartzite of the Plainfield Formation are exposed in places on both sides of the alaskite, but none are seen on the ridge we are on. The ledges immediately north of the power line are all alaskite, which is about 200 feet thick here.

Return west-southwest along the power line past the point at which we intersected it, into a woods road. A ledge west of the first intersection with another woods road is again the aegerine-augite granite gneiss. Follow the woods road back to the power line and bear left (south) down hill back to gravel pit and cars.

- Turn around and proceed west on Fire Street. The road passes down through the section just traversed into the Plainfield Formation.
- |      |     |   |
|------|-----|---|
| 23.2 | 2.1 | Four corners. Turn right (east) on Chesterfield Road.   |
| 23.7 | 0.5 | Sillimanite-biotite gneiss of the Plainfield Formation. This gneiss forms a partial mantle around the Gay Hill pluton of granite gneiss.  |
| 23.9 | 0.2 | Quartzite of the Plainfield Formation and pegmatite.  |
| 24.3 | 0.4 | Granite of the Gay Hill pluton. This granite has a peripheral zone of sillimanite and tourmaline-bearing granite where in contact with the sillimanite-bearing gneiss and schist of the Plainfield Formation. |
| 25.2 | 0.9 | Cross Old Colchester Road at Fair Oaks School.  |
| 25.6 | 0.4 | Turn right (east) on Route 163.   |
| 26.2 | 0.6 | Typical outcrops of quartzite of the Plainfield Formation.  |
| 26.6 | 0.4 | Cross Oxoboxo Brook.  |
| 27.4 | 0.8 | Montville Post Office.  |
| 28.4 | 1.0 | Ramps at I-395. Proceed under I-395 and park in abandoned service station on left.  |

STOP 5. (1 hour) Plainfield Formation in center of Montville dome. - Be alert for cars exiting from I-395. Cuts here are in the lower member of the Plainfield Formation on the north side of the Montville dome, about at the general horizon of the nodular quartzite and schist in the Stop 3A area. Pelitic gneiss, semi-pelitic gneiss, quartzite of the Plainfield Formation, and sills of biotite granite gneiss are exposed in the northbound entrance to I-395. We are still in the sillimanite-potassium feldspar zone of

metamorphism. Pelitic rocks contain sillimanite, garnet, and tourmaline as well as biotite, quartz, and feldspar. Tourmaline is a characteristic accessory in pelites of the Plainfield Formation. Note the somewhat laminar aspect of the pelitic gneiss. At entrance to the cuts on the west side is coarse sillimanite in a pegmatitic segregation vein. A sill of inhomogeneous biotite granite gneiss in the first part of the cuts consists of subequal amounts of microcline, calcic oligoclase, quartz, and minor biotite. The granite gneiss is considered to be intrusive. Cross-cutting contacts can be discerned on the east side of the cuts. Semi-pelitic gneiss containing flat lenses of quartz is common in the Plainfield. Loughlin (1912) called the Plainfield the "Plainfield quartz schist" because of the abundance of quartz in the unit. Quartzite layers tend to boudinage; their orientation indicates east-west extension. The amount of quartzite increases to the north end of the cuts, where some quartzite layers contain diopside. The hill to the north is composed of thin- to thick-bedded quartzite and minor interbedded meta-pelite, but the quartzite is not well exposed on I-395 which cuts across the hill, probably because the highway follows a zone of closely spaced joints. An amphibolite layer, probably a metamorphosed mafic dike, is exposed at the curve of the road on the east side. Late, north- to northwest-trending high-angle faults containing gouge and breccia cut the schist, gneiss, and quartzite in places. Folds of foliation have steeply-dipping axial surfaces oriented about N55W and dipping 70-75° N; axes plunge S75E 30°. Intersection lineation ranges from N20W 30° to N40W 55°.

Proceed east on Route 163 to Uncasville.

28.4 0.4 Turn right (south) on Route 32 at traffic light. Turn right again immediately at second traffic light and park behind stores.

STOP 6. (20 minutes) Quartz-sillimanite nodules. - Ledges across street from parking lot are granite gneiss containing approximately equal amounts of quartz, microcline, albite to sodic oligoclase, two percent biotite, and minor garnet and tourmaline. These outcrops are part of the marginal phase of the Gay Hill pluton. Outcrops behind the gas station on the east side of Route 32 contain nodules of quartz and sillimanite which have weathered from the rock leaving pits. A belt of pelitic schist of the Plainfield lies to the north of this outcrop. The composition of the nodules is similar to that in the nodules of the feldspathic quartzite except for the presence of tourmaline. The granite matrix in this rock is richer in albite and poorer in microcline than the main mass of the pluton. The Gay Hill pluton is almost entirely surrounded by pelitic gneiss and quartzite of the Plainfield Formation. I consider the garnet, tourmaline, and quartz-sillimanite nodules found in the outer part of the pluton, I consider to be the result of interaction of granite magma and pelitic wall rock. The composition of the granite is close to that of the granite-melting minimum.

Quartz-sillimanite nodules are fairly common in southeastern Connecticut in the Plainfield and Mamacoke Formations as well as in small granite masses and border zones of plutons. The mechanism for their formation and their presence in both quartzite and granite would make an interesting study. The matrix of the host rock, where granitoid, is generally rich in potassium feldspar and the texture is saccharoidal. I believe the nodules to be of metamorphic origin and related to muscovite decomposition reactions and consequent production of feldspar and quartz in the second sillimanite zone of metamorphism. This process however does not fully explain the presence of

nodules in a granitic matrix.

- Return to Route 32, turn left (north) on Route 32.
- 28.6 0.2 Outcrops of Plainfield schist and granite gneiss.
- 32.3 3.7 Fort Shantok Road. Turn right (east).
- 33.1 0.8 Turn left up hill on road just before Route 2A overpass.
- 33.5 0.4 Park in wide area overlooking Route 2A.

STOP 7. (45 minutes) Waterford Group and alaskite south of the Honey Hill fault. - The purpose of this stop is to observe the reduction in grain size and the laminar foliation related to strong penetrative deformation during development of the Honey Hill fault. In thin section the textures are not cataclastic but are metamorphic, and the foliation is conformable with the pattern of regional foliation in the lower plate.

The rocks are alaskite gneiss equivalent to Hope Valley Alaskite Gneiss and undifferentiated rocks of the Waterford Group, probably the Mamacoke Formation. Plainfield Formation outcrops across the valley to the south. Amphibolite layers typical of the layered New London Gneiss lie to the north close to the fault. The outcrop where the cars are parked is alaskite at the base of a sill-like mass about 2,700 feet thick; this mass forms the ridge we are on. The sill is entirely in the Waterford Group. Note the inclusions of mafic rock. This alaskite contains slightly more biotite (1 to 3 percent) than the sills of alaskite to the south that cut the Plainfield Formation where the mafic mineral is mostly magnetite. Drop down the embankment to the east end of the cuts along Route 2A. To the west of the alaskite is layered biotite-quartz-plagioclase gneiss. Layers differ in amounts of biotite, and some layers also contain hornblende. Note the fine grain size and laminar nature of the foliation. Contacts of layers appear sharp but are actually fuzzy in detail. The foliation is parallel to the trend and dip of the Honey Hill fault to the north. A streaking lineation is N20W 40°. A few folds toward the west end of the cut have axial surfaces that dip more steeply than the layers. Several steeply dipping zones of younger closely spaced joints and faults toward the west end of the cut that are younger than the Honey Hill fault strike N35E. Others trend about east-west. Pyrite is evident along some of the faults and joints. Zoned cross-cutting pegmatites are probably Permian in age. The undeformed pegmatites indicate that the pervasive dynamic metamorphism was earlier than pegmatite emplacement, and they represent an argument for pre-Permian age for the primary high-grade dynamothermal metamorphism in the lower plate and for pre-Permian movement on the Honey Hill fault.

- Return down hill to Fort Shantok Road.
- 33.9 0.4 Turn right (north) to Route 32.
- 34.7 0.8 Turn left (south) on Route 32.
- 35.0 0.3 Take ramp eastbound onto Route 2A. Cross Thames River.
- 36.8 1.8 Intersection with Route 12. Turn left (north). Continue on Rte 12 passed intersection of Rte 2A East.
- 37.7 0.9 Turn left past hospital onto a private road. Proceed to house at end of road.

STOP 8. (1 hour) Honey Hill fault. - Outcrops on the river bank across the railroad tracks are in the Honey Hill fault zone. Ledges on the hill to the north and along the railroad tracks are Tatnic Hill Formation in the upper plate. The outcrop at water's edge may actually contain the contact between

upper plate and lower plate rocks. Rocks are mylonite and mylonite gneiss and have a typical anastomosing fabric. The rocks were metamorphosed originally at sillimanite-orthoclase grade in the upper plate and sillimanite-muscovite grade in the lower plate, but they have been recrystallized to greenschist facies assemblages of green biotite, muscovite, and epidote. Garnet has been largely altered to green biotite, and sillimanite to sericite. This alteration has been dated by O'Hara and Gromet (1983) as Alleghanian. It also dates the latest movement on the fault and indicates the pressure-temperature conditions under which this deformation occurred. Intersection lineations plunge to the northwest as at the previous stop. Less deformed Tatnic Hill Formation along the railroad tracks to the north has a slightly different orientation than that of the mylonites. The parallelism of foliation in the lower plate seen in these rocks and at the last stop does not continue far into the upper plate. It is gone in a short distance, somewhere between the "ropy" gneiss in the upper part of the outcrop and the cuts along the railroad to the north. In this stretch of the Honey Hill fault, the units in the upper plate are discordant to the fault trace, except right along it, in contrast to the situation further west where the units have shallower dips and the trends of rock units parallel the fault trace (fig. 2)

End of trip. Return to I-395 via Route 2A. Travel south on I-395 for I-95 and New Haven.

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## PEGMATITES OF THE MIDDLETOWN DISTRICT, CONNECTICUT

by

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## GEOLOGY

Granitic pegmatites are widely distributed throughout the metamorphic rocks of Connecticut, but they are especially abundant in the vicinity of Middletown. The pegmatite fields of the Middletown district extend from Glastonbury south to Middletown and southeast to Haddam (Fig. 1). In general, the pegmatites have concordant north-south axial trends in the northern portion of the district but become increasingly discordant to the southeast with east-west axial trends in the vicinity of Haddam. The majority of pegmatites exposed in the Middletown district are hosted by metapelites (Collins Hill Formation of Rodgers, 1986), but many pegmatite bodies also are emplaced in gneisses (e.g., the Maromas gneiss) (Cameron et al., 1954; Stugard, 1958). Although the physical properties of host rocks may have influenced pegmatite shape and size, there is no indication at present that host rock lithology had a significant effect on pegmatite distribution or composition. Methot and Brookins (1973) have demonstrated that Rb/Sr systematics of several pegmatites in the district are distinctly different from those of their enclosing rocks, and thus that the pegmatites originated from some other protoliths (that are not exposed) and were emplaced into their hosts with very little chemical interaction. Most studies of oxygen and hydrogen isotopes in similar pegmatite districts also point to limited but intense pegmatite-wall rock interaction, the result of low water/rock volumetric ratios (e.g., Taylor et al., 1979; Taylor and Friedrichsen, 1983; Walker, 1985). The major and minor element compositions of the Middletown pegmatites vary widely throughout the district, but these chemical differences stem from fractionation processes that operated in pegmatite fields elsewhere (e.g., Cerny, 1982). The principal chemical variations are in modal plagioclase/microcline and in the proportions of minor or accessory minerals that contain high concentrations of the incompatible lithophile elements Li, Cs, Rb, Be, Nb, Ta, P, and B. Internal textures also are varied, but these too are controlled by pegmatitic processes that are typical of pegmatite fields elsewhere. In general, graphic quartz-feldspar intergrowths and aplites are associated with unzoned or weakly zoned pegmatites. With increasing chemical fractionation as

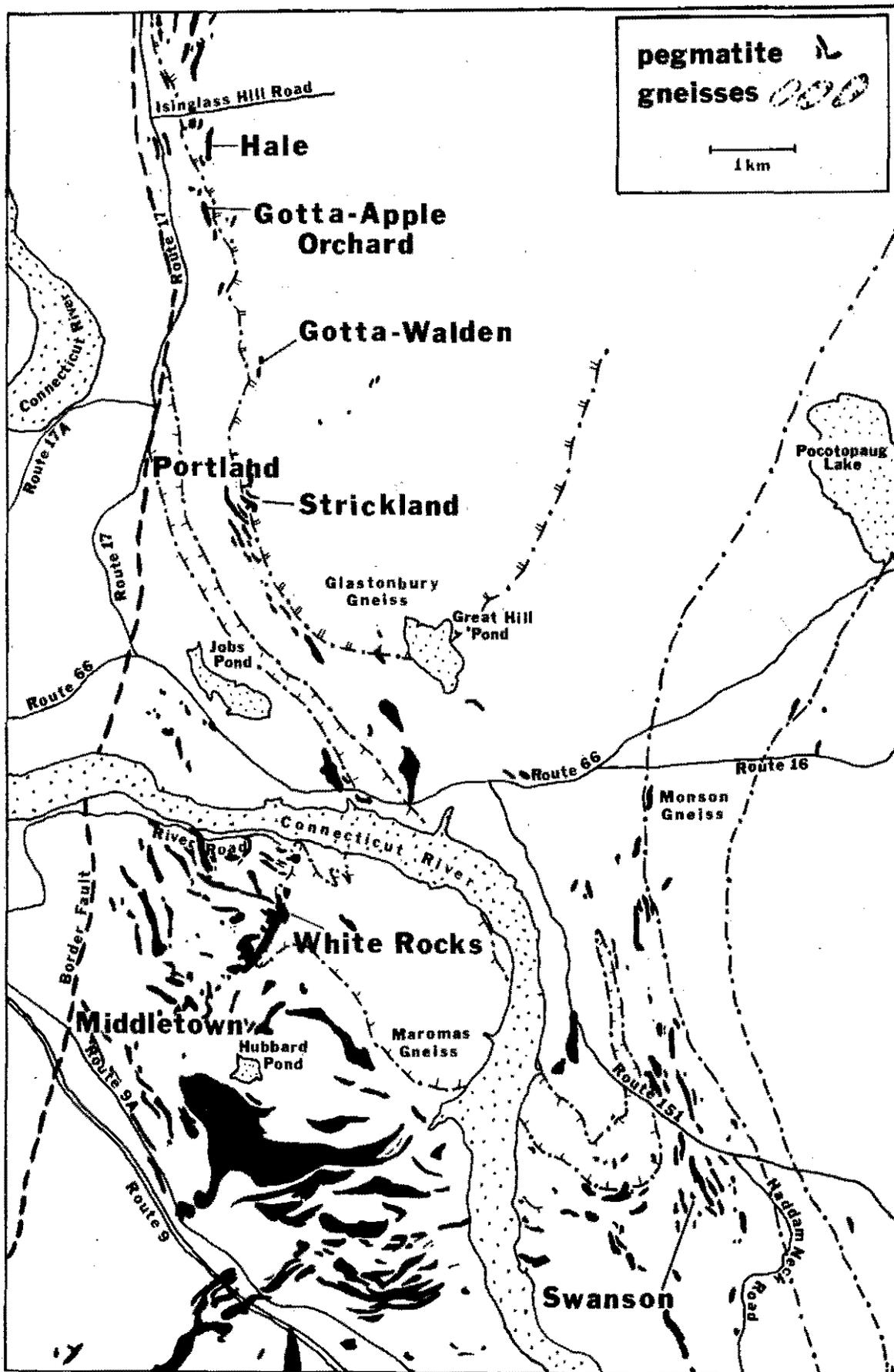


Figure 1. Distribution of pegmatite bodies in the Middletown area. Hachured lines delimit the boundaries of the major gneiss formations. Data are from Herz (1955), Stugard (1958), and Eaton and Rosenfeld (1972).

reflected by abundances of rare-element-rich minerals, the pegmatites display increasing mineralogical diversity and zonation. By comparison to other pegmatite districts, the regional chemical and textural zonation of pegmatites in the Middletown district roughly define the directions of pressure-temperature gradients that may have existed at the time of emplacement, or may reflect the folding or doming of isothermal surfaces to yield concentric zonation patterns in present exposures (Figure 2). In this model, compositionally simple pegmatites with graphic and aplitic textures are developed nearest to the parental granite or anatectic source; discretely zoned pegmatites that contain very-coarse-grained microcline, abundant beryl, and quartz-rich cores are further from the magma or thermal center; and complexly zoned, cleavelandite-bearing pegmatites with high concentrations of lithium and other incompatible elements are furthest from source (e.g., see Cerny, 1982). The P-T gradients that are proposed in Figure 2 reflect the fact that with increasing fractionation, pegmatitic magmas contain higher concentrations of more incompatible elements, especially fluxing components such as Li, B, and F. As a result, the most fractionated pegmatites have lower liquid and solidi, and thus these magmas can migrate down gradients to lower P-T conditions.

Although a variety of pegmatitic rocks are exposed in the vicinity of Middletown (e.g., Erslev, 1976), the characteristic pegmatites of the Middletown district are generally non-foliated, slightly to strongly discordant, and hence were emplaced in a largely atectonic or extensional structural environment after development of the pervasive regional foliation and attainment of prograde metamorphic assemblages. Radiometric isotope systematics yield ages of  $260 \pm 3$  m.y. (U-Pb, Th-Pb),  $258 \pm 1$  m.y. (Rb-Sr), and  $249 \pm 8$  m.y. (K-Ar) (see Brookins et al., 1969; Brookins, 1970). The dates derived from U-Pb, U-Th, and Rb-Sr systematics may be regarded as absolute ages of crystallization. The K-Ar date is interesting because it signifies that the pegmatites cooled from magmatic liquidus temperatures of  $600^{\circ}$ - $700^{\circ}$ C to the K-Ar closure temperature of muscovite (approximately  $250^{\circ}$ C) in roughly 10 m.y. (the relatively large uncertainty in the K-Ar age stems from degrees of retention or loss of radiogenic argon by different host phases, and probably from real variations in cooling ages that are controlled by pegmatite size and the heat contents of pegmatites and their host rocks). Between Middle Haddam and East Haddam, pegmatites that are typical of those of the Middletown group strike east-west and are sharply discordant to the north-south fabric of their metamorphic hosts. In the Moodus area, a north-northwest secondary schistosity developed as feldspathic blastomylonite at medium to high (amphibolite) metamorphic grade cuts the pegmatites, and the pegmatite field is truncated against this zone of blastomylonite, which defines the trace of a ductile shear zone termed the Cremation Hill fault (London, 1984d, 1985c). If the pegmatites involved in this deformation were comagmatic with other pegmatites that have been dated, then two important implications

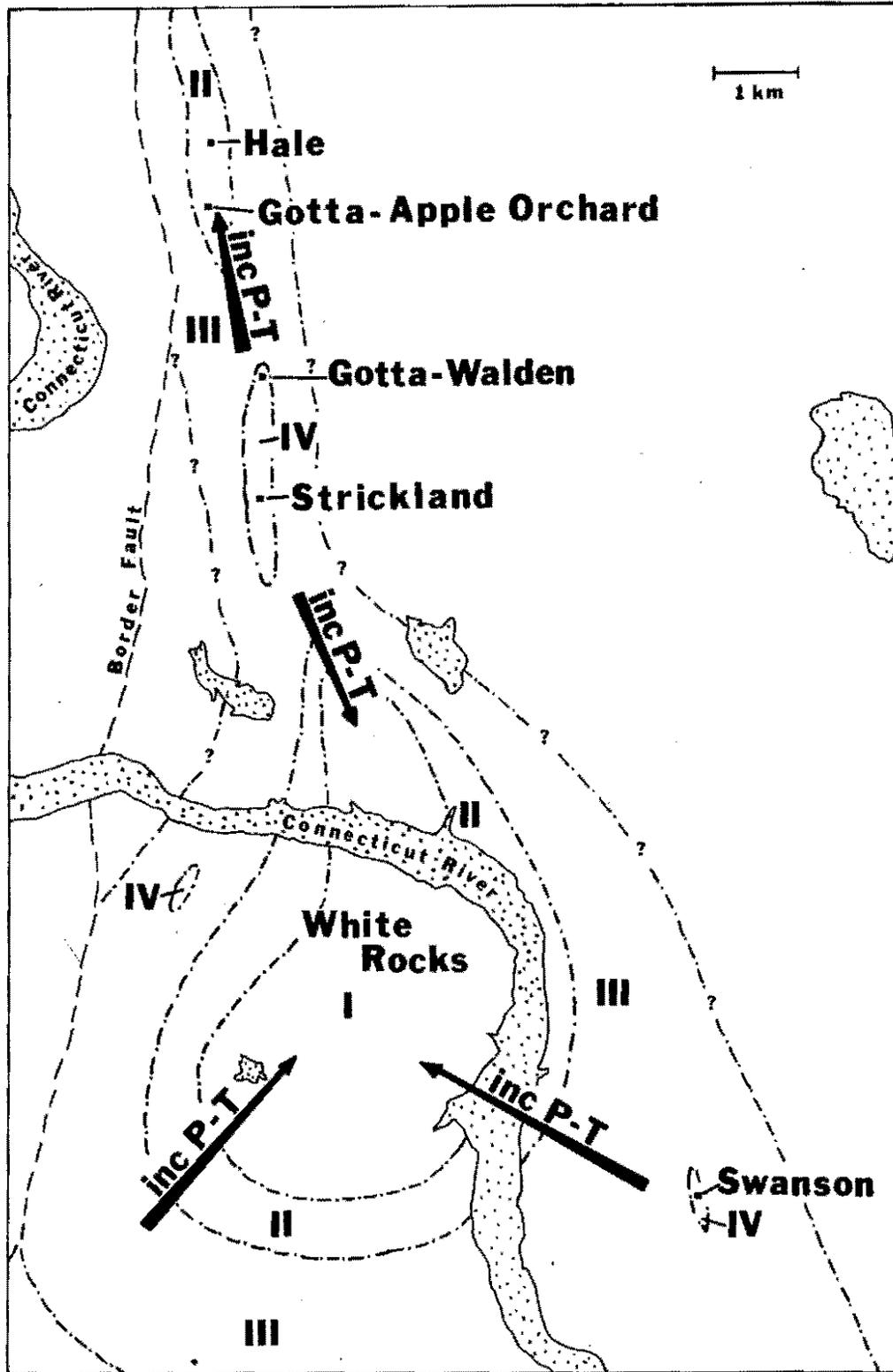


Figure 2. Geochemical zonation of pegmatites in the Middletown district. Pegmatite zonal designations are: I: unzoned microcline-plagioclase-quartz pegmatitic granite; II: graphic microcline-quartz pegmatite; III: zoned microcline-quartz-beryl pegmatite; IV: complexly zoned lithium-rich, rare-element pegmatite. Fractionation increases from I to IV. Note that each successively more fractionated pegmatite type usually contains some or all of the previous, less-fractionated types as outer zones (i.e., lithium-rich pegmatites usually have borders of pegmatitic granite and inner zones of quartz-microcline-beryl pegmatite).

are (1) that regional tectonism (other than Jurassic-Triassic rifting) continued after 260 m.y., and (2) that at the time of pegmatite emplacement, temperatures of hosts at the eastern boundary of the pegmatite field were significantly higher than in rocks to the west. If this last statement is true, then deformed and undeformed pegmatites near and west of the Cremation Hill fault can be expected to give K-Ar ages that are significantly younger than 249 m.y. Future comparisons of absolute and cooling ages within pegmatites may serve to define the thermal regimes at the time of pegmatite emplacement, and differential cooling rates as a function of variable rates of uplift.

There are few means by which the P-T conditions of crystallization can be ascertained in granites or pegmatites. Fluid inclusion analyses and stable isotope systematics usually give unreliable results because of pervasive postentrapment modification of inclusions and extensive retrograde reequilibration of isotopic distributions. Stability relations among the lithium aluminosilicates, however, provide a means by which approximate, sometimes precise P-T conditions can be unequivocally ascertained (London, 1984a, 1986). In the Middletown district, spodumene is the only lithium aluminosilicate mineral (found at the Gotta-Walden prospect and the Strickland pegmatite). If pegmatite liquidus crystallization spanned the approximate range of 600°-700°C, then minimum pressures for spodumene-bearing pegmatites would have been in the vicinity of 3000-4000 bars, with host rock assemblages above the aluminosilicate invariant point and in the sillimanite field (Holdaway, 1971). With respect to regional P-T gradients in the metamorphic hosts as discussed above, it should be noted that rocks on the eastern boundary of the Middletown district record assemblages in the second sillimanite zone, decreasing to first sillimanite zone in the vicinity of Haddam Neck and Higganum, and kyanite zone along the western boundary of the district in Portland.

#### CHEMISTRY

All of the Middletown pegmatites, even the most fractionated ones, consist principally of quartz, plagioclase, and microcline. Exotic Li-minerals such as spodumene, lepidolite, and amblygonite-montebbrasite, Cs-rich beryl and pollucite, and Nb-Ta-Sn oxides are common accessories in some pegmatites but are volumetrically insignificant. The most fractionated pegmatites show only minor amounts of fluorine (mostly in micas and amblygonite-montebbrasite) and phosphorus (as apatite, lithiophilite-triphyllite, and amblygonite-montebbrasite) but significant enrichment in boron (as reflected by an abundance of tourmaline).

Most studies of pegmatite petrogenesis ultimately focus on the properties or components that serve to distinguish the unusual textures and mineralogies of pegmatites from ordinary granites. The components H<sub>2</sub>O, the halides Cl and F, and

incompatible cations such as Li are believed to play major roles in controlling crystallization trends, element partitioning, and fluid properties (e.g., Jahns 1982). The components boron, cesium, rubidium beryllium, and phosphorus may also be important, although their behavior in hydrous granitic melts is not well known.

Most petrologists agree that H<sub>2</sub>O plays a major role in the generation of pegmatites and their associated ore deposits. By analogy to experimental results, Jahns and his colleagues have attributed the development of coarse-grained textures, idiomorphic crystal forms, and much of the mineralogical zonation in pegmatites to the separation of an aqueous fluid phase from H<sub>2</sub>O-saturated granitic magma early in the crystallization of pegmatite (Jahns and Burnham, 1969; Jahns, 1982). An evaluation of this hypothesis is beyond the scope and intent of this paper; however, there is now evidence that saturation in H<sub>2</sub>O is not requisite for the generation of rapidly grown, coarse-grained, idiomorphic crystals in synthetic systems that are analogous to those of natural, highly fractionated pegmatite fluids (London, 1983, 1984b, 1986). Intuitive reasoning suggests that water should be a more effective flux if it is retained in the silicate melt than if it is exsolved, and especially if it dissolves in the silicate liquid in proportions greater than that of albite-H<sub>2</sub>O (London, 1986; cf. Burnham, 1979). Thus, large quantities of H<sub>2</sub>O should not be necessary to generate pegmatites, and an increasing body of data from pegmatite textures (e.g., Norton, 1983), wall-rock alteration (e.g., Shearer et al., 1984) and stable isotope systematics (e.g., Walker, 1985) are consistent with very low water/rock volumetric ratios in the formation of all types of pegmatites. The development of miarolitic pockets is not contingent on large amounts of exsolved aqueous fluid or on shallow pegmatite emplacement (London, 1983, 1984b, 1985, 1986). Indeed, spodumene, the high-pressure lithium clinopyroxene, is present in miarolitic gem pegmatites in Connecticut, as well as other districts in North America (Maine, California) and around the world (e.g., Afghanistan). In summary, the commonly held notion that much of pegmatite formation results from "crystallization from an aqueous fluid" is probably one of the most important misunderstandings in igneous petrology (Jahns and Burnham, 1969). Indeed, Dick Jahns recognized the confusion that arose from this concept and attempted to clarify matters in his last major paper on pegmatite geology (Jahns, 1982). Pegmatites crystallize from dense silicate melts that contain relatively little water in comparison to the rock volumes that are produced (and in comparison to other types of felsic magmatic-hydrothermal deposits, such as base metal porphyries). At best, pegmatites may crystallize from silicate melt in the presence of an aqueous fluid, whose effects on promoting element partitioning and crystallization rates are still largely unknown.

In the Jahns-Burnham pegmatite model (Jahns and Burnham, 1969), the halogens, principally chlorine and fluorine, have been

regarded as crucial to the dissolution and transfer of mineral components and to extreme element partitioning that leads to pegmatite zonation. Experimental studies of metal solubilities and complexation in aqueous chloride vs. pure H<sub>2</sub>O systems generally indicate that Cl does enhance mineral solubilities, especially of alkali components and silica (e.g., Anderson and Burnham, 1967; Flynn and Burnham, 1978; Carron and Lagache, 1980; Webster, 1981; Fournier, 1983). Some recent studies of fluid inclusion compositions in lithium rich pegmatites, however, reveal that primary pegmatitic aqueous fluids have low salinities, generally < 4 wt. % equivalent NaCl (e.g., Foord, 1976; Taylor et al., 1979; London, 1982a, 1983, 1984b, 1985a, 1985b, 1986; cf. Cameron et al., 1953). In spodumene, beryl, quartz, and tourmaline, from miarolitic pockets of the Kulam deposit, Mawi district, Afghanistan, salinities of included aqueous fluids are as low as 0.5 equivalent wt % NaCl, and most are less than 2 eq. wt. % NaCl (London, 1985b, and unpublished data). If Cl contents are low, especially during the primary (vs. subsolidus) stages of crystallization, then the solubilities of most mineral components would be diminished, and fluid/melt partition coefficients would be decreased.

Fluorine has been shown to be effective at enhancing silicate melt-H<sub>2</sub>O miscibility and lowering liquidus and especially solidus temperatures of hydrous feldspar and granite melts (e.g., Wyllie and Tuttle, 1961; Koster van Groos and Wyllie, 1968; Manning, 1981); its relevance to pegmatite geology has been recognized in recent years (Jahns, 1982; Cerny et al., 1985). Local concentrations of pink micas (usually called lepidolite) have often been taken as an indication of relatively high fluorine activities in pegmatites; in most cases, however, the bulk of these pink micas are rose-colored muscovites (with Mn chromophore), and are not as F-rich as true lepidolites (e.g., see Rimal, 1962; Munoz, 1971; Rinaldi et al., 1972). The fact that lithian micas of the Connecticut pegmatites do predominate over spodumene is an indication of how low fluorine activities must be in spodumene-bearing pegmatites (London, 1982b). The compositions of minerals in the amblygonite-montebbrasite series also can be used to monitor activities of F as HF. Although these are typically referred to as amblygonite (the F-rich member of the solid solution), they usually turn out to be montebbrasites with intermediate to low fluorine contents (e.g., Cerna et al., 1972; London and Burt, 1982). Using the  $d_{131}$  vs. F x-ray determinative method of Cerna et al. (1973), primary amblygonite-montebbrasite from the Strickland and Gotta-Walden pegmatites in Connecticut contain 40 and 45 mole % amblygonite, respectively, and thus the phases should be called montebbrasite. Primary montebbrasite from Strickland is altered along fractures to a low-fluorine montebbrasite (20 mole % amblygonite), as at other localities that have been well studied (e.g., Cerna et al., 1973; Loh and Wise, 1976; London and Burt, 1982). From the equilibrium constant for F=OH exchange (Loh and Wise, 1976) and an assumption of ideal mixing of F=OH (albeit a poor one, because of proton-

proton avoidance), the fluorine activity (as HF) of a fluid phase in equilibrium with a phase of 50 mole % amblygonite at 600°C and 2000 bars is three orders of magnitude below the activity of HF at the minimum field boundary (600°C, 2000 bars) of lepidolite (Munoz, 1971). The magmas from which the Middletown pegmatites crystallized, therefore, contained relatively low fluorine concentrations.

The Middletown pegmatites are not notably phosphate-rich, although manganapatite is abundant in cleavelandite units of zoned pegmatites. Montebrasite and lithiophilite have been found in small quantities at Gotta-Walden and Strickland. In experimental melts, phosphorus behaves in part as a network former, in that it robs the melt of cations and silica tetrahedra of non-bridging oxygens, and as a network modifier, in that its monomers and polymers cut silicate melt structures (e.g., Ryerson and Hess, 1980; Mysen et al., 1981). At the Gotta-Apple Orchard, Gotta-Walden, Strickland, and Swanson mines, apatite-garnet and apatite-tourmaline intergrowths occur in cleavelandite units. The concentration of phosphorus and boron (discussed below) with comparatively alkali-rich rocks (albitites) may be a direct result of complexing between P, B, and alkalis and alkaline earths, and may possibly reflect liquid immiscibility between alkali-phosphosilicate or alkali-borosilicate fluids and silica-rich fluid.

Boron incorporated as tourmaline is abundant in the pegmatites and country rock contacts in the Middletown district, and this is typical of virtually all pegmatite districts around the world (e.g., Page et al., 1953; Cameron et al., 1954; Beus, 1968; Ovchinnikov, 1976). Other than quartz and feldspars, tourmaline is one of few minerals that is present at compositionally simple to complex pegmatites (i.e., it is ubiquitous). Tourmaline is assuming major significance in pegmatite petrogenesis as the role of boron becomes increasingly well understood, and as a growing number of studies demonstrate that the concentrations of the halogens Cl and F may be negligible in typical pegmatite systems (e.g., Chorlton and Martin, 1978; Pichavant, 1981, 1983; London, 1983, 1984a, 1984b, 1985a, 1985b, 1986). In the Middletown pegmatites, tourmaline is locally abundant in altered wall rocks at pegmatite contacts, it is a common constituent of pegmatite border and wall zones, but it is particularly abundant in albitites: both fine-grain aplites and coarse-grained cleavelandite units. In hydrous granitic liquids, boron probably behaves much like phosphorus, but with a distinct tendency to act as a melt network modifier. Experiments with alkali borosilicate systems demonstrate that melts are significantly depolymerized with exceedingly low viscosities (approximately  $10^2$  poise); crystallization rates are rapid, leading to very-coarse-grained, idiomorphic phenocrysts; silicate liquid-H<sub>2</sub>O solubilities are enhanced to the point that complete miscibility may be attained at geologically feasible P and T; and the solubilities of incompatible lithophile cations are greatly increased, probably because of the synergistic

effects of boron and H<sub>2</sub>O on melt depolymerization (most of these observations are from London, 1984b, 1986, and also Chorlton and Martin, 1978; Bonniaud et al., 1978; and Pichavant, 1981, 1983). It is important to note that experiments of London (1984b, 1986) produced euhedral albite, quartz, and petalite phenocrysts up to 0.5 mm in 48-72 hour runs at vapor-undersaturated conditions (i.e., of complete silicate-H<sub>2</sub>O miscibility). This can be interpreted to signify that coarse-grained, euhedral textures are not unequivocal evidence of the presence of an exsolved aqueous fluid phase. Although complete silicate fluid-H<sub>2</sub>O miscibility may exist in borosilicate fluids at moderate pressures, alkali borate-silica immiscibility is well documented in anhydrous experiments (e.g., Pichavant, 1983; Hervig and Navrotsky, 1985); this is because of the strong tendency for boron, like phosphorus, to form polymers that may coalesce to a separate fluid, and to form strong B-O-M<sup>+</sup> bonds. This leads not only to liquid immiscibility but also expands the quartz saturation field by increasing the silica content and silica tetrahedral polymerization in the fluid that coexists with the alkali borate-rich fluid. In pegmatites, the evidence of such liquid immiscibility may be quartz-rich cores with razor-sharp and smooth contacts with tourmaline- and phosphate-rich albitites.

#### ECONOMIC GEOLOGY

Much of the pegmatite mining in the Middletown district early in this century was for mica, feldspar, beryl, and especially gemstones (e.g., at the Strickland, Gillette, and Swanson mines); many of the pegmatites produced beryl and mica during the 1940's (Cameron et al., 1954); at present, the only commercial mining of pegmatites in the district is for ceramic-grade microcline, and this is the only economic commodity of the Middletown pegmatites in the foreseeable future. Bulk mining of mica from weathered schists and granites and beryllium minerals from zeolitized tuffaceous rhyolites has largely replaced pegmatitic sources for these commodities. Pegmatites continue to be important sources of Nb-Ta-Sn oxides, but these phases occur only in trace amounts in the Middletown pegmatites. If such rare-element enriched pegmatites ever existed in the Middletown district, they would have been structurally highest and farthest from source; thus, they have been eroded, or buried by sediment in the graben of the Connecticut Valley to the west. Some of the Middletown pegmatites (e.g., Gotta-Walden, Strickland, Gillette, and Swanson) still have potential to produce gemstones and mineral specimens for collectors, but these operations are labor-intensive and generally cannot sustain gem mining as a sole economic activity.

#### DESCRIPTIONS OF STOPS

The Hale, Gotta-Apple Orchard, and White Rocks pegmatites are actively quarried, so it is not possible to describe precisely what the exposures will look like at the time of the trip. The Gotta-Walden and Swanson mines are inactive; the

Strickland pegmatite is shown on Figure 1 because it is one of the larger and better-known exotic pegmatites of the district, but we will not visit it on this trip.

All of the localities are on private land, and permission to enter the properties must be obtained from the owners or operators. In addition, all participants will be required to sign waivers that release property owners from all liability for personal injury and loss or damage to personal property. In addition, MSHA regulations require that all participants wear hardhats, safety glasses or goggles, and reinforced boots into active quarries. Participants must provide their own safety equipment.

#### ROAD LOG

From New Haven, proceed north on I-91 to the interchange for Rt. 9 southbound (approximately 25 miles from downtown New Haven), and take Rt. 9 south for 3.5 miles to exit 3 at West Street in Cromwell. Take exit 3 south, and we will convene the trip at the north end of the nearby K-Mart parking lot. Allow about 45 minutes for the drive from New Haven, and be prepared to leave the K-Mart lot by 9:00 AM. If you have not brought a lunch, we will have several opportunities to stop for food.

- 0.0        The trip log begins at the K-Mart parking lot. Return to Rt. 9 south to Middletown.
- 2.9        First traffic signal on Rt. 9, turn right under the railroad overpass and proceed uphill to the rotary in front of St. John's church.
- 3.1        Turn left at the rotary onto Main Street in Middletown. Proceed through Middletown on Main Street, past the Middletown Green and straight on the Main Street Extension.
- 4.1        Turn left off of Main Street Extension onto Cooley Avenue; follow the signs to the Middletown Post Office.
- 4.2        Turn right off of Cooley Avenue onto East Main Street, still following signs for the Middletown Post Office.
- 4.3        At the traffic signal, turn left onto Silver Street.
- 4.7        Cross over Rt. 9 on Silver Street, continue straight on Silver Street uphill and past the Connecticut Valley Hospital.
- 5.4        Bear right off of Silver Street onto River Road.
- 5.6        Pass WCNX radio station.

6.6

Turn right off of River Road uphill into the White Rocks quarries of the Feldspar Corporation.

The White Rocks quarries (Figure 1) are located near the apparent center of the pegmatite district. As a whole, they are primitive pegmatites in that they possess simple compositions and generally poorly developed zonation. Contacts with wall rocks are generally sharp without conspicuous metasomatic alteration. Negligible tourmalinization and coarsening of quartz-feldspar-biotite schists are developed locally. There are many pegmatite bodies in the vicinity of White Rocks. As in other districts, these pegmatites probably formed a stockwork through their host rocks. At early stages of crystallization, the fluid communication through the system was good, and compositional gradients established by liquid-state diffusion or zone refining were transmitted through the entire system. As the degree of crystallization increased, the pegmatites may have become sealed off from each other and evolved along separate crystallization trends. The pegmatites at White Rocks were almost certainly interconnected, as evidenced by uniform zonation and fractionation trends through many individual bodies. The present quarry exposures are in homogeneous and poorly zoned pegmatite. Toward the west in exposures that are not inaccessible because of mining and backfill, the homogeneous and graphic pegmatites evolved to well zoned bodies with almost monomineralic microcline "blocks" sharply bounded against essentially pure quartz pods or "cores"; beryl was abundant but only occurred at the microcline-quartz contacts. One small pegmatite in the westernmost cuts contained a small pod of cleavelandite albitite with lithian mica, polychrome tourmaline, cesian beryl, and accessory microlite. These observations, collected over the past ten years through numerous visits (London, unpublished data), reflect a sharp zonation pattern with increasing fractionation from the center of the White Rocks hill to the west (Figure 2).

In the active quarries, the pegmatites consist predominantly of two zonal assemblages: (1) homogeneous, medium-grained (2-5 cm), microcline-plagioclase phenocrysts with interstitial quartz and accessory garnet, biotite, muscovite, and tourmaline, and (2) graphic quartz-microcline intergrowths. In the first assemblage, scaly muscovite is pseudomorphous after coarse biotite sheets. The contacts between these two zones appear to be gradational through an intermediate texture in which microcline becomes coarsely porphyritic and graphic. Single microcline crystals with graphic quartz intergrowths attain dimensions of approximately 1 meter. Not all coarse-

grained microcline is graphic; some large, quartz-free crystals mark the incipient development of what are termed block microcline zones, which denote the abrupt transition to pegmatitic zoning as reflected by almost quantitative separation of quartz and feldspars. Columbite is a common accessory of the incipient block microcline zones at White Rocks.

Some exposures display intersecting tabular pegmatite dikes that cut the graphic pegmatite. On close inspection, however, the cross-cutting pegmatite dikes are not separate intrusives that are discordant to and significantly later than the graphic pegmatite. The graphic pegmatite grades continuously into the pegmatite dikes; in fact, the pegmatite dikes are developed within single graphic microcline crystals, such that microcline and quartz within the dikes and in the graphic pegmatite on either side are crystallographically continuous. The pegmatite body as a whole apparently was able to sustain shear during crystallization; one possible interpretation is that rapid deformation of partially crystallized pegmatite produced shears in the viscous aluminosilicate liquid, with resultant diffusion of  $H_2O$  down pressure gradients toward the shear zones leading to enhanced crystal growth rates and thus coarser grain size in the residual liquid of the pegmatite "dikes".

The origin of graphic microcline-quartz textures was much debated in the early part of this century but now is generally acknowledged to represent simultaneous crystallization of quartz and feldspar. The explanation advanced by Fenn (1979) is that feldspar growth rate exceeds component diffusion rates through the crystal boundary liquid. Crystallization of feldspar on topographic "highs" (e.g., by step growth along edge dislocations) produces liquid that is depleted in feldspar components but enriched in silica in the "valleys" between highs; this leads to saturation in quartz in the valleys. In addition, for graphic textures to be produced instead of granitic textures, crystal nucleation rate and nucleation density must remain low.

- 6.6 Leave the White Rocks quarries, return to River Road, turn left, proceed back to Silver Street, turn left onto Silver Street, retrace the route back to the Post Office.
- 8.9 At the traffic signal at the intersection of Silver Street and East Main, continue straight on Silver Street.

- 9.0 Turn right onto Main Street Extension. Continue straight on Main Street Extension to downtown Middletown.
- 9.5 Turn right off of Main Street onto Monitor, proceed 0.1 miles to a parking lot behind Burger King, which is where we will buy lunch. Rather than eat at the Burger King, we will take our food to the Hale quarry.
- 9.7 At Monitor and Main, turn right onto Main Street and continue through Middletown.
- 9.7 Bear left at the rotary in front of St. John's church, follow the signs for Portland and Rts. 66E and 17N.
- Cross the Connecticut River into Portland.
- 11.3 The first traffic signal upon entering Portland is at the junctions of Rt. 17A and Rt. 17N/66E. Proceed straight through the traffic signal on Rt. 17A through Portland.
- 13.1 Pass the Portland Fire Department.
- 14.1 Stop sign at the intersection of Rts. 17A and 17. Turn left (north) onto Rt. 17 towards Glastonbury.
- 15.1 Pass Gotta's Farm Store.
- 16.2 Turn right onto Isinglass Hill Road.
- 16.4 Turn right into the Hale quarry off of Isinglass Hill Road.

The Hale, Gotta-Apple Orchard, and Gotta-Walden bodies appear to be portions of one large pegmatite system that strikes north-south with a moderate westerly dip. The principal features to see at the Hale and its extension to the Gotta-Apple Orchard quarry are: (1) banded albite-quartz-(garnet)-(tourmaline)-(muscovite)-(apatite) aplites, (2) graphic quartz textures in individual very-coarse-grained microcline crystals, (3) block microcline-beryl-quartz pods, and (4) the spatial distributions of aplite-pegmatite zonation and the location of abundant tourmaline. The Gotta-Walden prospect exposes (1) a cleavelandite unit with abundant garnet and accessory tourmaline and apatite, and (2) a very-coarse-grained, non-graphic zone of block microcline with abundant muscovite at the apex or "hood" of the pegmatite. Although the textures are quite different, the strike and dip of the dike and the distributions of internal zones (by composition) are identical to those at Hale and Gotta-Apple Orchard.

If conditions are dry, we can drive from the Hale quarry past the powder magazines east of the cut to the Gotta-Apple Orchard extension. Alternatively, we will double back on Isinglass Hill road to Rt. 17S to the entrance to the Apple Orchard quarry, as follows:

- 16.6 To get to the Gotta-Apple Orchard quarry, leave the Hale quarry, return to Isinglass Hill Road, left on Isinglass Hill Road to Rt. 17, left (south) on Rt. 17.
- 17.2 Turn left off of Rt. 17S onto the asphalt road that leads uphill through the apple orchard to the quarries. The entrance to the apple orchard is marked by a silver metal gate. The quarries are about 1 mile up the road.

The Hale and Gotta-Apple Orchard quarries are discussed together, because they represent portions of the same pegmatite dike. In both quarries, aplites consist principally of albite and quartz with minor garnet and tourmaline and accessory muscovite and apatite. Fine banding defined by alternating concentrations of quartz, albite, and garnet form smooth layers and crenulated masses. In general, the layers are flat and parallel, except in the vicinity of large crystals of microcline or massive quartz; at these intersections, the banding in the aplites commonly deflects around the large crystal. Thus, the crenulations appear to result in part from unidirectional growth and deflection around small perturbations on the inwardly directed solidification surface. The growth of large microcline crystals may also have depleted the boundary fluid in feldspar components and thus led to slowed aplite formation in their vicinity. In most igneous rocks, unidirectional solidification produces comb layering, in which non-equidimensional crystals have their fast growth directions oriented perpendicular to the solidification front. The crystals of microcline within the aplites are oriented perpendicular to the aplite bands and flare inward toward the center of the pegmatite; beryl and tourmaline in most pegmatites near Middletown have a similar comb-textured orientation with their long axes perpendicular to pegmatite borders and flaring into the pegmatite. The tourmaline in aplites at Hale and Gotta-Apple Orchard, however, is generally foliated parallel to layering, or perpendicular to the presumed growth front. This observation suggests that at least some of the layering in the aplites could have resulted from flow segregation of phenocrysts. At the Gotta-Apple Orchard quarry, there is no simple relation between the orientation of aplite bands and the pegmatite contacts (i.e., the aplite banding is not parallel to pegmatite contacts). Aplite layers cut pegmatite, but more commonly pegmatite dikes appear to cut aplite.

Lineated and foliated quartz-muscovite intergrowths that are "axial planar" to the aplite crenulations project into coarse-grained quartz-microcline pegmatite. The microcline crystals, especially in the Hale quarry, usually but not always contain graphic quartz. Single microcline crystals contain graphic and non-graphic portions, suggestive of continuously or episodically changing conditions (perhaps changes in activity of H<sub>2</sub>O) during crystallization. As at White Rocks, beryl at the Hale quarry is found only where coarse microcline crystals project into small pods of massive quartz.

Coarse tourmaline is abundant only on the western margin of the pegmatite in coarse-grained microcline-quartz pegmatite (in both quarries). This tourmaline displays the inwardly expanding or flaring habit that is typical of border zone tourmalines at pegmatites throughout the world. The only conspicuous wall-rock alteration at either the Hale or Gotta-Apple Orchard quarries is along these tourmaline-rich borders, where foliated hornblende gneisses are altered pseudomorphously to tourmaline + quartz, with detailed preservation of layering and foliation. Coarse tourmalines of the wall and border zones and within the microcline-quartz pegmatite possess different pleochroic zones that correspond to different tourmaline compositions. The nature of the compositional variations are not known at this time. It is interesting to note, however, that tourmalines in the aplites are generally unzoned, and their pleochroic schemes correspond to those of the outer overgrowths of tourmaline in the coarse pegmatite. This zonation could signify that tourmalines in the pegmatitic zones started to crystallize before those of the aplite, and hence that aplite crystallization followed that of pegmatite.

The border-to-border sequence of aplite → graphic microcline → microcline + quartz pegmatite → tourmalinized wall rock is a classical zonation in shallowly dipping pegmatites. The sequence of layered albitic aplites on the lower or footwall, grading upward into coarse-grained microcline pegmatite, and wall-rock alteration along the upper or hanging wall only, defines the zoning sequence in layered aplite-pegmatite intrusives of San Diego county, California; these pegmatites served as the basis for an important model for pegmatite crystallization that was proposed by Dick Jahns (Jahns and Tuttle, 1963). Jahns' explanation for such non-concentric layered sequences involved the upward migration under gravitational gradients of exsolving aqueous fluid; partitioning of components between H<sub>2</sub>O-saturated melt and aqueous fluid

provided the means of producing compositional gradients in fluids that resulted in heterogenous, zoned rocks. This model is receiving considerable scrutiny in current pegmatite research. At the Hale and Gotta-Apple Orchard pegmatites, however, the albitic aplites are on the hanging wall, and the coarse microcline pegmatite plus tourmalinized wall rock are on the foot wall; thus, the pegmatite zonation is upside down and could not have been produced by gravitationally induced melt-vapor separation, unless the pegmatites have been overturned more than  $90^{\circ}$  by rotational deformation since their crystallization. There is no evidence of internal deformation in the pegmatites, nor any specific model for such regional deformation in the metamorphic rocks.

- 19.1 Return to Rt. 17, turn left (south).
- 19.4 Turn left at the fork off of Rt. 17 onto Cotton Hill Road.
- 19.6 Continue straight on Cotton Hill Road to the parking lot for Walden's Rock Shop; park next to the shop.

The Gotta-Walden prospect is reknown for gem beryl and rarer phases such as spodumene and pollucite. The pegmatite, however, is compositionally and zonally very similar to the Hale and Gotta-Apple Orchard: cleavelandite albitites with garnet, tourmaline, and apatite on the western hanging wall grade eastward into coarse microcline pegmatite on the foot wall. The exotic Li minerals and beryl, however, are concentrated in the albitite. Of special interest here are nodular garnets in the albitite that are rimmed by manganapatite and tourmaline. Garnet-tourmaline-plagioclase and garnet-manganapatite-plagioclase equilibria are capable of buffering activities of boron and phosphorus, respectively. The relative modal proportions of these phases at Hale, Gotta-Apple Orchard, and Gotta-Walden imply that the albititic liquid components of these pegmatites were not rich in B and P, either because these components were not enriched in the bulk pegmatite fluid, or because they diffused or were partitioned into some other liquid component of the pegmatite system, or were lost to the wall rocks (e.g., London, 1984c).

Exposures at the Hale, Gotta-Apple Orchard, and Gotta-Walden pegmatites display several important characteristics of pegmatitic mineralogical zonation: contacts between feldspar-rich zones are abrupt but gradational, whereas feldspar-quartz zonal contacts are razor sharp. Pegmatite masses can be divided into two fundamental components: one that is feldspar-rich, and

the other quartz-rich; the feldspar component is either sodic or potassic; the two feldspars rarely occur together in nearly equal proportions. Aplite-pegmatite contacts at these Connecticut pegmatites are typical of feldspar (albite)-feldspar (microcline) zonal contacts in that they are abrupt but not as sharp as feldspar-quartz contacts. Boron- and phosphorus-rich minerals are concentrated in feldspathic zones, especially in the albitites (aprites and cleavelandite units), and beryl is concentrated at feldspar (albite or microcline)-quartz zonal contacts. As noted above, the chemical segregation of components in pegmatite systems has been previously ascribed to the effects of element partitioning during exsolution and gravitational ascent of a comparatively saline aqueous fluid from silicate melt (Jahns and Burnham, 1969). The Cl and F contents of pegmatitic fluids, however, are usually quite low, so that aqueous fluids should not be capable of dissolving and transporting large quantities of material. There are a number of other arguments against the involvement of large quantities of saline aqueous fluid in the chemical segregation of pegmatitic magmas (e.g., London, 1985a, 1985b, 1986). One explanation that does not require silicate-H<sub>2</sub>O immiscibility is that of liquid-state segregation, termed thermogravitational diffusion, of components under thermal and gravitational gradients. Thermal gradients in pegmatites probably are small but may be of the order of 50°-100°C; over several million years, such gradients would certainly cause major redistribution of components (e.g., see Orville, 1963). Gravitational gradients may be especially effective in pegmatitic magmas if they are highly depolymerized. Thermally and gravitationally induced chemical segregation is a precursor to liquid immiscibility, and silicate-silicate liquid immiscibility is a real possibility for pegmatitic systems. Such immiscibility would be promoted by the relatively high concentrations of boron and phosphorus in pegmatites, which form such stable complexes with alkalis that these components are significantly removed from the residual silicate liquid. In simple synthetic systems, immiscibility is prevalent, and the immiscible fluids are alkali borate or phosphate liquid that coexists with essentially pure silica. Evidence of such immiscibility in pegmatites might be manifest as the tourmaline- and phosphate-rich albitites that have sharp, smooth, sometimes convoluted contacts with essentially pure massive quartz. Silicate liquid-H<sub>2</sub>O immiscibility probably does occur at early stages in most pegmatites, but if the exsolved aqueous fluid is relatively pure H<sub>2</sub>O, then it may have a comparatively minor effect on bulk redistribution of elements (n.b. that an assumption of Burnham's (1979) albite-H<sub>2</sub>O model is that the activity of albite

component in aqueous fluid that coexists with H<sub>2</sub>O-saturated albite melt is essentially zero). In hydrous silicate liquids that are fluxed by boron, phosphorus, and fluorine, H<sub>2</sub>O will tend to redissolve or homogenize back into the melt as the contents of hyperfusible components are increased by fractional crystallization of quartz and feldspars, and as the compositional differences between melt-aqueous fluid solvus pairs are minimized. Significant amounts of H<sub>2</sub>O may be released only during the late-stage crystallization of these fluids, so that most exsolved aqueous fluid probably exists late in the history of pegmatite consolidation during subsolidus rather than primary magmatic conditions.

- 19.8 Leave Walden's Rock Shop, proceed south on Cotton Hill Road back to Rt. 17, turn left (south) onto Rt. 17.
- 20.9 Stop sign at the intersections of Rts. 17 and 17A. Continue straight through the intersection on Rt. 17S.
- 22.8 Stop sign at the junction of Rts. 17 and 66. Turn left onto Rt. 66E. There is a Dairy Queen at this intersection for those who need refreshment.
- 25.0 Pass by roadcuts of massive, unzoned pegmatitic granite.
- 25.8 Traffic signal at the intersection of Rts. 66 and 151 in Cobalt. Turn right onto Rt. 151. Proceed through Middle Haddam.
- 28.2 Bear left at the traffic signal at the entrance to Hurd State Park. Continue on Rt. 151, which is now called Moodus Road.
- 29.0 Turn right off of Rt. 151 onto Haddam Neck Road. The turn is marked by a sign for the Connecticut Yankee Information Center.
- 29.7 Continue on Haddam Neck Road past a series of red barns.
- 29.8 Cross under powerlines.
- 29.9 Turn left at the Haddam Neck Fire Department and park in the southwest corner of the lot.

Cross the road and hike approximately 0.3 miles along a path that leads to the Swanson mine under the powerlines on the north side of an electrified livestock fence. As you near the Swanson mine, the path forks in front of a large hill of outcrop; bear right to the Swanson mine dumps.

The Swanson mine was prospected and successfully operated for gem beryl during the 1940's (Cameron et al., 1954). Although very little of the pegmatite is now exposed, the dumps still provide samples of the exotic mineralogy of this highly fractionated pegmatite. The prospect is developed in a small portion of a much larger and compositionally ordinary microcline-rich pegmatite; however, microcline is scarce at the Swanson. The two zonal constituents of the Swanson pegmatite are massive quartz (core), and cleavelandite albitite that contains abundant beryl, polychrome tourmaline, manganapatite, lepidolite, garnet, cassiterite, microlite, and triplite (a Mn-fluorophosphate); the presence of triplite at the Swanson may signify that the chemical activity of fluorine species was higher here than at other pegmatites of the district; however, the fluorine species would have been largely halide salts (leading to the production of micas), rather than HF (there is no topaz) (London, 1982b). The relative abundance of Nb-Ta-Sn oxide minerals here probably is a direct consequence of the depolymerizing effect of the boron and phosphorus that were obviously important constituents of the albitic liquid. With increasing depolymerization, the melt structure is sufficiently "open" and disordered that comparatively high quantities of incompatible elements may be retained in solution. When the fluxes are removed from such fluids, as by the crystallization of tourmaline and phosphates, the albititic and exotic ore components of the fluid are also dumped from solution.

End of trip. Return to Rt. 66 via Haddam Neck Road and Rt. 151.

Mineralogical and textural variations such as those of the Middletown pegmatites are what have made pegmatite geology so alluring to so many petrologists. Complex as they are, these exposures present only a few of the many pegmatitic features that hold a wealth of untapped data on basic inorganic periodic chemistry, partition coefficients, fluid properties, crystallization processes, and cooling histories. The P-T conditions of many pegmatites can now be evaluated by combinations of lithium aluminosilicate stability relations (London, 1984a), fluid inclusions (e.g., London, 1986), and stable isotope geothermometry (e.g., Taylor and Friedrichsen, 1983). In the future, further insight into pegmatite petrogenesis will require an improved understanding of the properties of the pegmatitic fluid phases from emplacement through subsolidus conditions, and of the causes of chemical and hence mineralogical zonation.

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**LATE QUATERNARY DEPOSITS OF THE SOUTHERN QUINNIPIAC-FARMINGTON  
LOWLAND AND LONG ISLAND SOUND BASIN: THEIR PLACE IN A  
REGIONAL STRATIGRAPHIC FRAMEWORK**

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**INTRODUCTION**

The glacial and postglacial deposits and related erosional features of the southern part of the Quinnipiac-Farmington lowland, CT (fig. 1, fig. 2, fig. 3), have been studied in detail from the late 1800's to the 1960's. They are here placed into a modern regional stratigraphic framework. This synthesis results from: compilation of the glacial geology of Connecticut for two State maps (Stone, J. R., Schafer, J. P., London, E. H., and Thompson, W. B., in press; Schafer, J. P., Stone, J. R., London, E. H., and Thompson, W. B., unpublished map); recent cooperative work between the Connecticut Geological and Natural History Survey and the U.S. Geological Survey at Woods Hole in mapping the distribution of glacial and Holocene deposits in Long Island and Block Island Sounds (Lewis and Needell, in press; Needell and Lewis, 1984; 1985); and a regional radiocarbon chronology for New England and Long Island (Stone and Borns, in press).

The Quinnipiac-Farmington lowland contains ideal examples of deltaic and related lacustrine deposits (fig. 2) of successive glacial lakes that formed during northward retreat of the last ice sheet. Distal glaciofluvial deposits of the Quinnipiac valley terrace (fig. 2, unit qt) are unique in southern New England because they extend through 47 km (29 mi) of the lowland and are distinct in color and mineralogy from the ice-marginal deltaic deposits that they overlie and entrench. Till in the lowland is the typical red-brown variety derived from the Mesozoic rocks that underlie the Central Lowland of Connecticut. On the western border of the Lowland, red-brown lowland till overlies gray, crystalline-derived upland-type till, and southwesterly striations cross southeasterly ones (fig. 3). Divergent directions of ice movement during glacial advance and retreat are indicated.

The thick and extensive body of lake clay beneath Long Island Sound (fig. 4) was laid down in a major glacial lake which occupied that basin. The New Haven and East Haven deltas (fig. 2, units Cn, Ce) were also deposited in this lake, which is here called glacial Lake Connecticut. The postglacially tilted waterplane of the lake slopes southward from an altitude of about 30 ft (48.3 m) at the New Haven delta to about

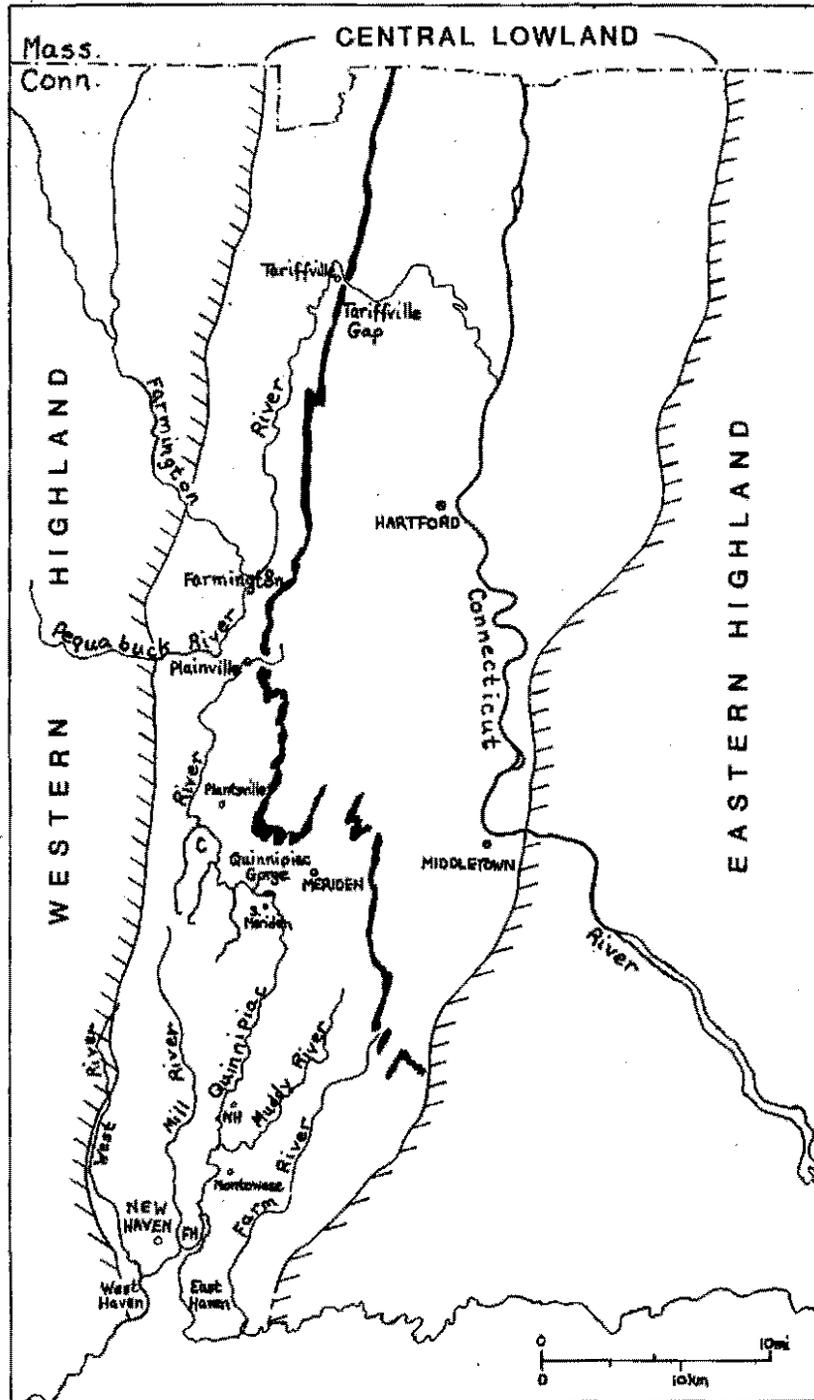


Figure 1.—Physiography of the Central Lowland of Connecticut. The Quinnipiac-Farmington lowland lies west and south of the basalt and diabase ridges, shown in heavy black; The Connecticut lowland lies east and north of the ridges. Location of towns are shown: C, Cheshire; NH, North Haven; FH, Fair Haven.

present sea level along an isobase that extends from Fairfield to Clinton. The waterplane projects to the north shore of Long Island, south-southeast of New Haven, at an altitude of about 72 ft (22 m) below sea level (-22 m MSL).

The absolute ages of glacial and postglacial deposits in the Long Island Sound-lower Quinnipiac-Farmington lowland area are estimated from a regional array of correlated ice-marginal deposits and radiocarbon dates (Stone and Borns, in press), and dated materials in marine sediments that are related to the late Pleistocene-Holocene marine transgression in the region. These dates indicate that the glacial-lake sediments in Long Island Sound basin probably are older than 18,000 radiocarbon years B.P. Sediments of glacial Lakes Quinnipiac and Southington and fluvial sediments of the Quinnipiac valley terrace probably are older than 17,000 radiocarbon years B.P. Marine transgression into eastern Long Island Sound probably occurred between 12,000 and 13,000 years B.P.; transgression into central Long Island Sound took place later, but before 10,200 radiocarbon years B.P. At 6810 years B.P. alluvial deposition in the modern floodplain of the lower Quinnipiac valley was still taking place at an altitude of at least as low as -6 m MSL.

## SETTING

### Quinnipiac-Farmington Lowland

Connecticut consists of three physiographic provinces: the Eastern and Western Highlands, and the Central Lowland, which extends north to south through the State (fig. 1, fig. 3). The bedrock of the Western Highland in the area of this trip consists of Paleozoic crystalline rocks including gneisses, schists, and metavolcanics (Rodgers, 1985). The bedrock of the Central Lowland consists principally of red or reddish brown Mesozoic arkosic sandstones and siltstones, which generally dip eastward to southeastward. The sedimentary rock units are interrupted by basalt and diabase which form narrow linear ridges within the Lowland. These traprock ridges trend generally north-south, except at Meriden where the Hanging Hills turn sharply eastward (fig. 1). The ridges are nearly as high as the Highlands to the east and west and they divide the Central Lowland into two parts. The wide Connecticut lowland in the east and north is drained by the Connecticut River, which flows out of the basin at Middletown. The Quinnipiac-Farmington lowland forms the relatively narrow western and southern parts of the Central Lowland.

Today the drainage of the Quinnipiac-Farmington lowland is separated at Plainville by a low divide on glacial deposits. North of the divide the Farmington River, which enters the lowland from the northwest, is joined by the Pequabuck River, which enters from the west. The Farmington flows northward about 30 km (18.6 mi) and enters the Connecticut lowland through a gap in Talcott Mountain at Tariffville (fig. 1). South of Plainville the Quinnipiac River flows southward to Cheshire. There it turns east-southeastward and passes through a gorge cut in the soft, and there highly faulted, New Haven Arkose to South Meriden. From thence it flows southward and, along with the lesser Mill and West Rivers to the west and the Farm River to the east, drains the southern portion of the lowland.

Except for their respective courses through the Tariffville Gap and Quinnipiac Gorge, the Farmington and Quinnipiac Rivers flow entirely on thick glacial deposits. The bedrock floor of the lowland north of the Quinnipiac Gorge does not have a systematic slope either north or south. It lies generally at about sea level, but contains several deep glacially scoured closed depressions. These depressions are 100-200 ft (30.5-61 m) deep and one at Plainville is -300 ft (-91.5 m) in altitude at its bottom (Haeni, 1975;

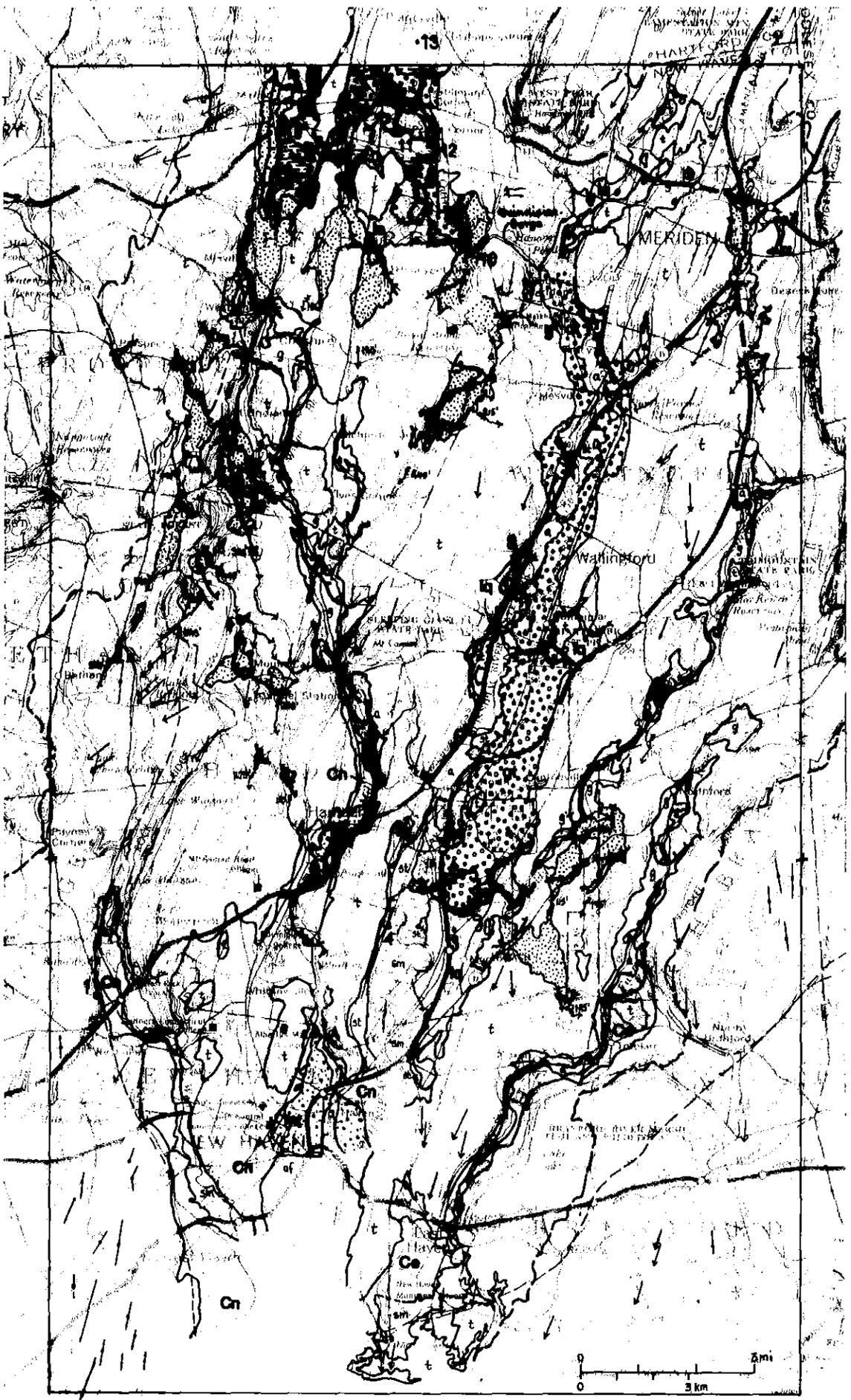


Figure 2.—Quaternary geologic map of the southern part of the Quinnipiac-Farmington lowland, CT. Modified from Schafer, J. P., Stone, J. R., London, E. H., and Thompson, W. B. (unpub. map).

### EXPLANATION

#### Postglacial deposits (late Wisconsinan, Holocene)

af	Artificial fill
a	Alluvium
sw	Swamp deposits
sm	Salt marsh deposits
st	Stream terrace deposits

#### Glacial meltwater deposits (late Wisconsinan)

Deposits graded to and deposited in major glacial lakes in lowlands

Cn	New Haven delta of Lake Connecticut
Ce	East Haven delta of Lake Connecticut
lq	Lake Quinnipiac
ls	Lake Southington
	lake-bottom deposits

Deposits of ice-marginal ponds in north-draining valleys tributary to the Mill and Quinnipiac Rivers

	undivided deposits
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Deposits of lacustrine and fluvial-lacustrine systems in the upper Mill, Muddy and Farm River valleys

g	undivided deposits
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Deposits of glaciofluvial systems

	Mill River terrace
	Quinnipiac valley terrace

#### Glacial deposits (late Wisconsinan)

t	Till, includes areas of bedrock outcrops
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#### Map symbols

	Striation, groove, or crag and tail
	Drumlin axis
	Glacial lake or pond spillway, altitude in feet
	Ice-margin position, ticks point toward ice
	Melt-water-carved channel in till
	Drainage divide of Quinnipiac River basin
	Contacts between Mesozoic Rocks and crystalline rocks
.2	Field-trip stop location
.A	Assembly point (East Rock Park)

Mazzaferro, 1975; Handman, 1974, 1975; Handman and Meade, 1975; Handman and Ryder, 1973). In the Quinnipiac Gorge, bedrock lies at just below 100 ft (30.5 m) in altitude. South of Cheshire and the Quinnipiac Gorge, the bedrock valleys of the Quinnipiac, Mill, West and Farm Rivers slope southward to the Sound. The bedrock floor of the Quinnipiac valley south of the Gorge (Haeni, 1974a; 1974b; 1975; Haeni and Sanders, 1974; Brown, 1974) has a glacially scoured depression 150 to 200 ft (45.7 m to 61 m) below sea level from Yalesville to North Haven. From North Haven to Fair Haven, where the bedrock channel joins that of the Mill River valley, it slopes from -100 to -250 ft (-30.5 to 76.2 m) in altitude. Beneath the southern part of New Haven, West Haven and New Haven Harbor the bedrock valleys of the West, Mill, and Quinnipiac Rivers converge at about -250 ft (-76.2 m) in altitude. A bedrock trench beneath the Sound slopes southwestward to depths greater than -228 m MSL.

### Long Island Sound Basin

Sediments of late Jurassic, Cretaceous and Tertiary age cover the western margin (paleoshelf) of the Atlantic Ocean basin and form the seaward-thinning sedimentary wedge of the Atlantic Continental Shelf of the United States. From Florida to New York City the wedge of Coastal Plain sediments extends many miles inland to where the Coastal Plain strata pinch out against the eastern flank of the Appalachian Mountains. From New York City eastward to the Scotian shelf the Coastal Plain strata are mostly below sea level. Waterbodies such as Long Island Sound, Block Island Sound (fig. 4) and Rhode Island Sound and the Gulf of Maine occupy a string of lowland basins at the inner margin of a dissected Coastal Plain wedge. The seaward flank of this inner lowland is marked by a north-facing submerged cuesta scarp that is cut in Coastal Plain strata. The position of this cuesta scarp in many places coincides with or is close to the maximum extent of the late Wisconsinan ice sheet.

Long Island Sound occupies the westernmost basin of the inner lowland. The Long Island Sound basin is bounded seaward by a highly dissected, north-facing escarpment of Cretaceous Coastal Plain strata (fig. 5). The south-dipping crystalline bedrock surface under Long Island Sound is a seaward extension of the bedrock that underlies coastal Connecticut. Submerged extensions of the Connecticut and Thames River bedrock valleys slope under the Coastal Plain cuesta, and the general morphology and slope of the submerged bedrock surface are very similar to those of the glacially modified bedrock surface on land.

From about the middle of Long Island Sound basin southward, the crystalline bedrock is overlain by Coastal Plain strata (fig. 5). The north-facing cuesta scarp rises an average of 90-100 m above the bedrock/Coastal Plain contact, but the cuesta is very irregular. Several large valleys, as much as 140 m deep, incise the Coastal Plain strata. The morphology and distribution of these valleys indicate that they are glacially modified fluvial features. Opportunities for extensive fluvial erosion of the Coastal Plain strata occurred during major sea level lowerings in the Oligocene, late Pliocene and early Pleistocene. Because the fluvially carved inner lowland and Coastal Plain cuesta existed prior to glaciation, their positions influenced the configuration of terminal positions of the Wisconsinan ice sheets. Two major moraine lines cap the Coastal Plain cuesta.

### **PREVIOUS INVESTIGATIONS IN THE QUINNIPIAC-FARMINGTON LOWLAND**

The earliest detailed studies of the Quaternary features of the area were those of James D. Dana, who published intermittently on the subject between 1870 and 1895. Dana (1883-1884) recognized the southwest direction of ice movement within the Central Lowland and attributed it to basal movement of the ice in the Lowland while the upper part of the ice sheet moved southeastward across the State. Most valuable are Dana's

detailed descriptions of sedimentary features in the stratified deposits of the New Haven plain. Dana (1870;1875, p. 168-183; 1884, p. 113-130) described from many places an upper stratum of cross-bedded gravel and sand, 10 to 20 ft thick and showing predominantly southward current directions, which unconformably overlies a lower stratum of finer grained sand. This unconformity today is interpreted as a topset-foreset contact of the New Haven delta (as suggested by Lougee, 1938).

In the **Manual of the geology of Connecticut** (Rice and Gregory, 1906) Gregory briefly described the glacial geology of the Quinnipiac-Farmington lowland and accounted for the present courses of the Farmington and Quinnipiac Rivers as results of glacial drainage diversions caused by deposition of stratified drift in former channels during retreat of the last ice sheet. He recognized the existence of a glacial lake between Cheshire and Southington and its role in the development of the Quinnipiac Gorge.

**The Quaternary geology of the New Haven region, Connecticut**, by Freeman Ward was published in 1920. It included the first description of the Quinnipiac clay (later called New Haven clay) in which deposition in a freshwater glacial lake was postulated, rather than in the sea as previously thought. Ward described the distribution and thickness of the clay and accounted for its layered nature as a result of the seasonal melting of the glacier ice and release of a regular supply of water which carried a uniform amount of silt and clay each season. With an average of 15 silt-clay layers per foot of deposit and thicknesses ranging from 50-150 ft, at least 500 and no more than 2500 years for the life of the lake were postulated. Ward, like Dana, described the stratified drift deposits of the New Haven area as upper gravel unconformably overlying lower sand. His photograph of these units in the excavation for Yale Bowl was printed in Rice and Gregory (1906) and Ward (1920), and reproduced by Lougee (1938). We reprint it (fig. 6) because it is the only visible evidence available at this time for the deltaic nature of the New Haven plain.

In 1928, Ernst Antevs included the Quinnipiac valley clay in his varve chronology of northeastern North America and counted a total of 364 varve couplets in surface exposures in 6 clay pits in Hamden and North Haven. He correlated the 364-year interval with a similar varve record in clay deposits at Haverstraw, NY.

In **The Glacial Geology of Connecticut** (Flint, 1930), interpretation of the deposits of the Quinnipiac-Farmington lowland was based on a hypothesis that the glacier became stagnant over southern New England and melted down from north to south with residual ice remaining longest in Long Island Sound. Accordingly, residual ice masses were left in larger valleys and the meltwater deposits of the Quinnipiac-Farmington lowland were formed in narrow lakes at successive levels along the sides of a shrinking ice mass. Flint (1932) retracted the total-stagnation and north-to-south melting idea; he also changed his interpretation of the stratified glacial deposits of Connecticut from their deposition in principally lacustrine environments to principally fluvial ones. In **Late-glacial Features of the Quinnipiac-Farmington Lowland** (Flint, 1934) he continued to advocate persistence of stagnant ice masses through the length of the lowland during deposition of principally fluvial ice-contact deposits, which formed "an extensive mass, apparently a unit, traceable from near the Massachusetts State line almost continuously south to New Haven." In this paper, Flint recognized the occurrence of a younger unit "laid down by aggrading streams during and after the wasting away of the last remnants of residual ice." The "buff to drab" color of this unit, in sharp contrast to the red or pink ice-contact materials, indicated that they were derived from the crystalline rocks of the Western Highland.

In 1938, **Physiography of the Quinnipiac-Farmington Lowland in Connecticut** by Richard J. Lougee was published. This is a classic paper on the glacial geology of this region. The interpretations of deposits and the deglaciation history described therein are (with some modification) those accepted by the compilers of the recent State maps. Strangely enough, this fine work was largely ignored by later quadrangle mappers of the area, and never referenced by R. F. Flint in later papers or quadrangle reports. Of most significance to the present discussion are the following points discussed by Lougee. 1) From descriptions by Dana (1875-1876, 1883-1884) and Ward (1920), Lougee recognized that the New Haven plain was probably deltaic, built into a lake occupying the Long Island Sound basin. He also concluded that the clay of the Quinnipiac valley was laid down in that lake and that the Muddy River delta was graded to it. He mentioned, but did not favor, the possibility that a lake in the Quinnipiac valley had been impounded by the Fair Haven plain as it built across the mouth of the valley. 2) He also recognized independently, he claimed, from Flint the two generations of red and "yellow" sand and gravel deposits in the Quinnipiac valley. 3) He described glacial-lake features in the lowland north of the Quinnipiac Gorge, especially topset-foreset contacts in ice-marginal deltas, which he projected to a tilted water plane. The plane indicated that post-glacial crustal tilting had occurred in the amount of 5.5 to 6 ft per mile to the north. 4) He described the Quinnipiac Gorge as a late-glacial drainage diversion, cut by the waters draining from the Southington glacial lake.

During the early 1960's, the southern part of the Quinnipiac-Farmington lowland was mapped in detail at 1:24,000 scale under a cooperative mapping program between the U.S. Geological Survey and the Connecticut Geological and Natural History Survey. This mapping produced the following quadrangle maps and reports which have been invaluable in compiling the regional geology: New Haven and Woodmont quadrangles (Flint, 1965); Branford quadrangle (Flint, 1964); Mount Carmel quadrangle (Flint, 1962); Wallingford quadrangle (Porter, 1960); Meriden quadrangle (Hanshaw, 1962); Southington quadrangle (La Sala, 1961).

Also important to the understanding of the glacial deposits of this area are the subsurface data from numerous logs of wells and testholes made available in U.S. Geological Survey, Water Resources Division basin reports (La Sala, 1968; Mazzaferro, 1973; Mazzaferro and others, 1979; Haeni and Anderson, 1980).

### **TILL, DRUMLINS, STRIATIONS AND ICE MOVEMENT DIRECTIONS**

The tills of the Quinnipiac-Farmington lowland are reddish brown to brown with variations to reddish gray or light brown in some places. They are derived from the red sandstones and siltstones of the New Haven Arkose which underlie the lowland. Near the basalt and diabase ridges the till color is grayish brown.

On the western side of the lowland, crystalline-derived light gray to olive gray till in places (Stop 1) underlies the red till. This relationship occurs as much as 3 km (1.9 mi) west of the western border of the Mesozoic rocks (Flint, 1962). On the eastern side of the Central Lowland, red till extends eastward as much as 2 km (1.2 mi) beyond the eastern border fault of the Mesozoic rocks (Flint, 1964; Langer, 1977).

The texture of the lowland till varies from clayey to sandy. In general, it is somewhat less stony than crystalline-derived till and the relatively few large boulders within it are commonly of basalt or diabase.

Drumlins and striations trend south-southeast in most of Connecticut (fig. 3). They trend west-southwest, southwest, and south-southwest in the Quinnipiac-Farmington lowland (fig. 2, fig. 3). West-southwest striations occur as much as 4 km (2.5 mi) west of the steep scarp of the Western Highland on the central west border of the Lowland in the Southington, Bristol and Collinsville quadrangles. Striations trending west of south cross older striations trending east of south in a few places on the west side of

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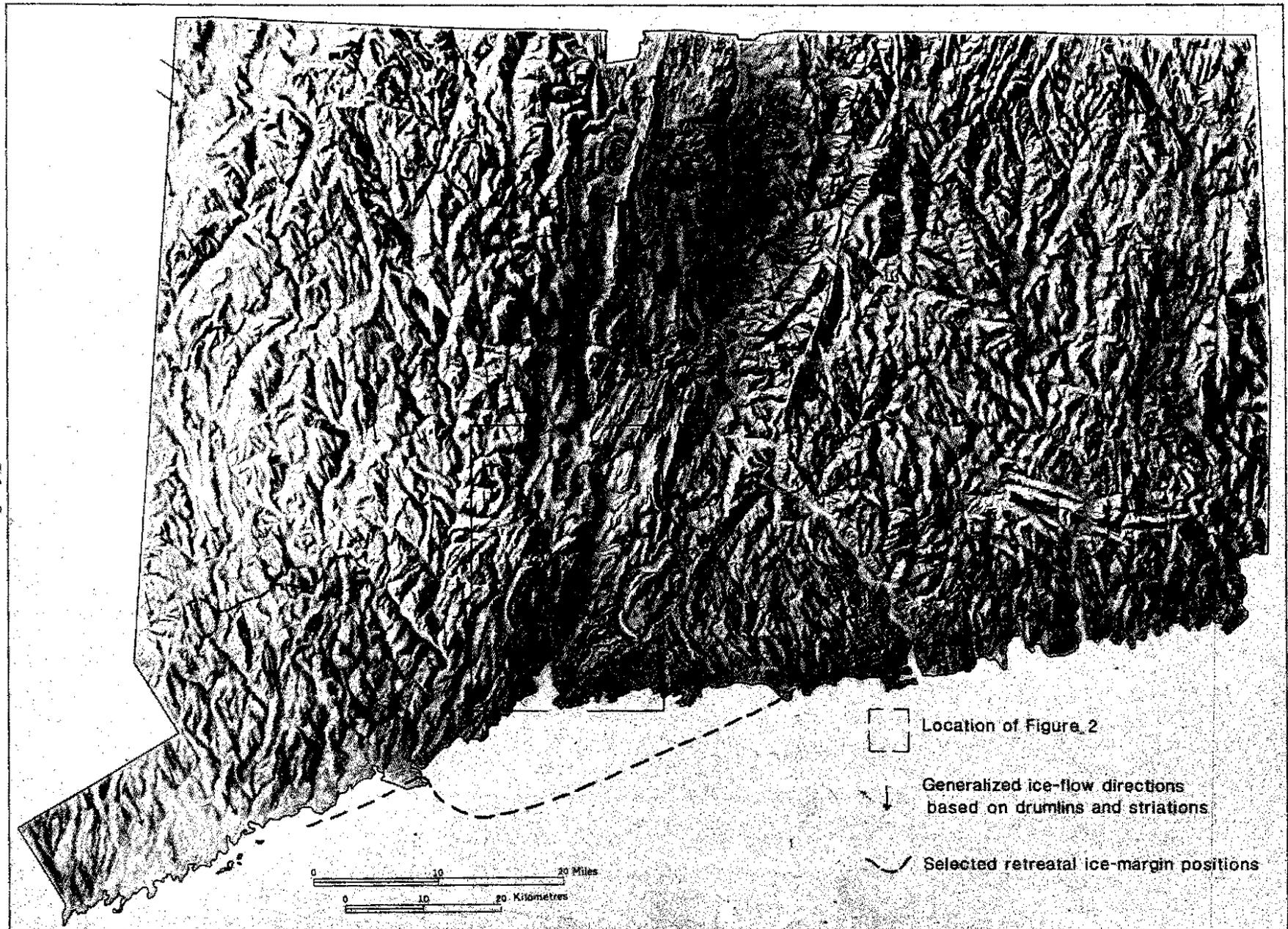


Figure 3.—Generalized ice-flow directions, and selected late Wisconsinan retreatal ice-margin positions in Connecticut. Data from Schafer, Stone, London, and Thompson (unpublished map). Shaded relief base shows Eastern and Western Highlands and the Central Lowland. Retreatal ice-margin lines show the shape and length of the Connecticut Valley ice lobe in the Central Lowland.

the Lowland (Stop 11 of Stone and others, 1982) and in the south-central part (Hanshaw, 1962). These few southeast-trending striations on the west side and within the Lowland provide evidence that when the ice was thickest during maximum glaciation, the difference in altitude between the Highlands and the Lowland was not a significant factor to ice-movement direction. During retreat of the ice sheet, however, the ice became thinner and the difference in altitude resulted in the formation of a lobe of ice in the Central Lowland (fig. 3). The southwest striations on the west side of the Lowland and the occurrence of red till west of the lowland are due to movement during retreat of the Connecticut Valley ice lobe. Although there is no evidence that the margin readvanced, the ice lobe remained vigorously active as is shown by its ability to move west-southwesterly up the steep scarp of the Western Highland.

Drumlins throughout Connecticut are composed of the lower, pre-late Wisconsin till (Stone and others, 1982, in press; Pessl and Schafer, 1968). The southwest trend of drumlins in the Quinnipiac-Farmington lowland, and especially southwest of New Haven, requires further study to determine the relationship between till fabric, provenance, age and topographic trend. It appears that final shaping of these drumlins, presumably composed of the older till, took place during retreat of the last ice sheet.

Throughout the Quinnipiac-Farmington lowland the northeastward retreat of the western side of the Connecticut Valley ice lobe (fig. 3) is recorded by numerous melt-water deposits, especially in upland valleys. Northwest-southeast-trending ice-margin positions were requisite to deposition at these specific high-level positions (fig. 2).

## GLACIAL LAKE CONNECTICUT: OFFSHORE AND ONSHORE EVIDENCE

### Glacial deposits in Long Island Sound

The glacial and postglacial deposits in Long Island Sound recently have been mapped from high-resolution seismic reflection profiles supplemented by information from vibracores (Lewis and Needell, in press). Maps showing depths to the submerged extensions of the Long Island moraines and the distribution and thickness of rhythmically laminated lake clays beneath the Sound, add significantly to our understanding of the damming mechanisms, water levels, extent and duration of the glacial lake that occupied the Sound basin during early retreat of the last ice sheet. Deltas along the coast in Connecticut (fig. 4), the youngest and largest of which is at New Haven, were deposited in this glacial lake, here called glacial Lake Connecticut.

Scattered deposits of till, commonly more than 10 m thick, overlie the bedrock and Coastal Plain remnant (fig. 5) throughout the Long Island Sound basin. Submerged moraine (fig. 4) with crest altitude of about -20 m MSL between eastern Long Island and Block Island (Needell and Lewis, 1984) is the extension of the Ronkonkoma-Shinnecock-Amagansett (Sirkin, 1982) terminal moraine line. Submerged moraine with crest altitude of -10 to -20 m MSL between Plum Island and Fishers Island is the extension of the Harbor Hill-Charlestown recessional moraine line. The now submerged parts of the Harbor Hill-Charlestown and Ronkonkoma moraine heightened a relatively low area across the Coastal Plain cuesta and impounded meltwater as the ice retreated into the isostatically depressed basin. The morainal dams persisted long enough for the entire area to be filled with lake-clay deposits. These lake clays are thickest (up to 150 m) in the pre-glacial valleys that incise the Coastal Plain strata, but they are found throughout the Long Island Sound and Block Island Sound basins. To the south, the lacustrine deposits pinch out against Coastal Plain strata and/or moraine. To the north they pinch out against bedrock or older glacial material. Locally the clays extend into estuaries along the Connecticut coast as in New Haven Harbor (fig. 4) and beneath the deltas, as at New Haven (fig. 7).



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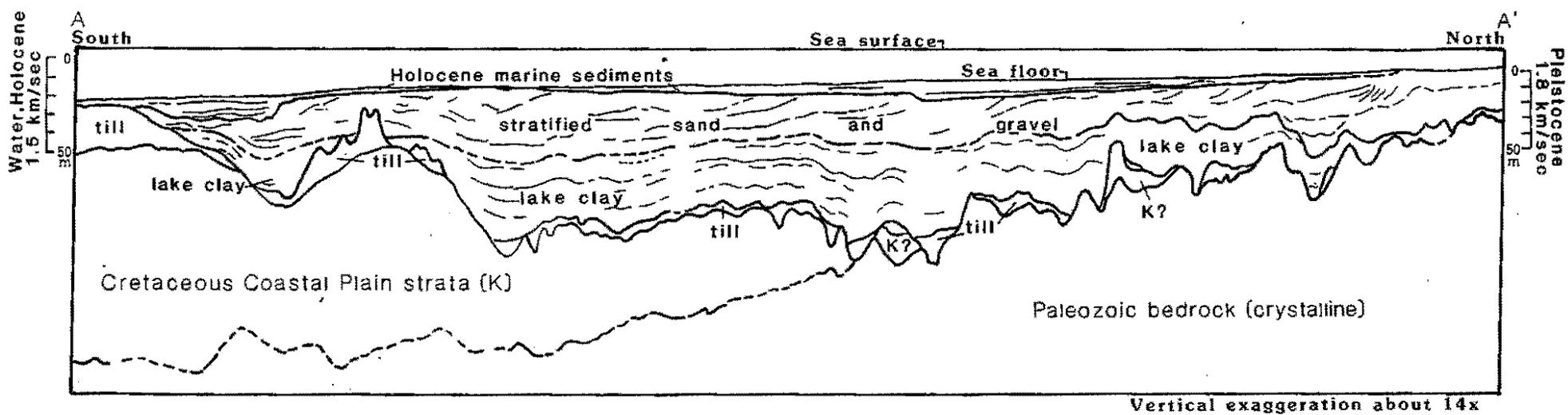


Figure 5.—Interpreted seismic reflection profile from Long Island Sound, south of New Haven, CT (see fig. 4, section A-A'). Length of section is 13 km.

Several features that are relevant to the later history of the lake lie in and around the Long Island Sound basin. A ridge composed of bedrock and Coastal Plain strata extends across the basin south of Branford. East of this high, the lake clays are incised by stream valleys which are in turn truncated by a widespread, flat unconformity. This evidence suggests that the lake clays were subaerially exposed and dissected before wave action associated with the Holocene marine transgression beveled the lake-clay surface. West of the Branford high and as far west as Bridgeport, CT, evidence of subaerial exposure of the lake clays has not been found. In this area, the lake clays are conformably overlain by extensive stratified sand and gravel deposits (fig. 4). Offshore from New Haven these sediments are 20-45 m thick, (fig. 5) and extend southward almost to the north shore of Long Island.

A probable spillway for lake waters and later drainage outlet from the subaerially exposed lake basin exists at the head of Block Channel where the terminal moraine is notched to approximately 75 m below sea level. A channel eroded in the lake-clay surface leads to this outlet and probably provided a path for eastward drainage of the Long Island Sound lake. As the lake lowered, the Branford high may have emerged and prevented or delayed complete draining of the central part of the lake.

South of the mouth of the Connecticut River, remnants of an extensive, internally complex deposit lie above the marine unconformity that beveled the lake clays. As this deposit postdates the last transgression, it is therefore marine.

A cover of Holocene marine mud blankets most of the basin. East of Clinton, strong currents have cut deeply into the glacial and marine deposits, and locally the Coastal Plain strata. Much of the bottom topography of the eastern Sound is a result of this continuing modern erosion.

#### The New Haven delta plain

The oldest meltwater deposits in the Quinnipiac-Farmington lowland are the New Haven delta, along with the one at East Haven and included fluvial feeder deposits in the West, Mill, and Farm River valleys (fig. 2, units Cn and Ce). They were graded to and deposited in glacial Lake Connecticut. That these deposits are deltaic was suggested by Lougee (1938), but in the New Haven quadrangle report, Flint (1965) described the New Haven plain as consisting of more than 250 ft of "stream-deposited sediment down to so great a depth" as the result of deposition in a glacial age when sea level was much lower than today. Although there are no known deep exposures of the subsurface deposits in the New Haven urban area at this time, there is little doubt that the plain is deltaic as evidenced by Dana's descriptions (1875-1876, 1883-1884) and the Yale Bowl excavation photograph (Fig. 6) and other descriptions by Freeman Ward (1920) of upper gravels (topset beds) and lower sands (foreset beds). Neither Dana nor Ward recognized the plain as a delta. However, the following paragraphs from Ward's (1920, p. 45-46) description of the stratified drift of the New Haven region clearly imply the deltaic character of the plain:

"Differences in color, texture, and structure make it advisable to classify stratified drift as upper gravels, lower sands, and clay. Upper gravels, so named because of their position and texture, occupy the upper portion of stratified drift and have a thickness of 8 to 25 feet. Sands are found in it, but gravel, coarse gravel, cobbles, and boulders predominate. The structure shows considerable variety. Cross-bedding is common, and flow and plunge structure ... is found in several places. Layers are short and lenslike and change rapidly in character vertically and horizontally ... But where made of coarse gravel and boulders, the layers may be thick and massive for perhaps 100 yards in length ... The color of the upper gravels is yellow-brown. They lie upon the eroded edges of the lower sands.

The lower sands formation lies directly beneath the upper gravels. It consists of sands of various grades ... Locally gravel layers may be found in the lower portions. The lower sands are characteristically even-bedded, cross-bedding is lacking or on a small scale; the layers are of greater extent; not prominently lenslike, and vary slightly. The color is red-brown ranging from the distinctively bright color to the more common quieter tones.

The division between the upper gravels and lower sands is a plane of unconformity wherever observed. In the New Haven plain it was seen along the whole 'cut' where the Shore Line enters the city, also in more than 20 other scattered localities ... The exposures are sufficient in number and widely enough separated to indicate that the unconformity is a regional feature. The upper gravels rest directly upon the eroded surface of the lower sands."



Figure 6.—Excavation for Yale Bowl showing glaciofluvial delta-topset gravel overlying inclined delta-foreset sand. Photo by Freeman Ward, prior to 1906. Altitude of land surface approximately 45 ft; estimated thickness of topset gravel is 3.6 m (12 ft).

From thicknesses of 12 to 15 ft for the topset beds (Dana, 1875-1876; and fig. 6) beneath downtown New Haven and at Yale Bowl where surface altitudes are about 45 ft, we estimate that the topset-foreset contact at this latitude is at about 30-33 ft in altitude. It is probable that the New Haven delta was built in segments as the ice margin retreated northeastward, although it is not evident from surface profiles of the New Haven plain. If this were the case, the section along the West River to West Haven would have built first, and the section along the Mill River to Fair Haven slightly later. If the level of the lake dropped a little during the successive stages of delta building, topset-foreset contacts might be at lower altitudes in Fair Haven than in West Haven.

This would also require greater aggradation of fluvial topset beds in Fair Haven in order for surface altitudes to be about the same (45 ft) in both areas. The 25-ft-deep section exposed along the "Air Line" Railroad cut through the northern part of Fair Haven, described by Dana (1875-1876, p. 173) and (Ward, 1920, p. 45), revealed a 20-ft thickness of the cross-bedded "upper stratum" unconformably overlying the lower stratum. At the Yale Bowl excavation on the west side of the West River, only about 3.6 m (12 ft) of topsets are present (fig. 6).

Numerous subsurface data from wells and testholes (Mazzaferro, 1973) show the downward stratigraphic sequence of gravel over sand over clay beneath the plain. Figure 7 indicates the extent and thickness of subsurface lake-bottom deposits of very fine sand, silt and clay which lie beneath the deltaic sediments of the New Haven plain. These lake-bottom sediments are probably continuous with the lake clay beneath Long Island Sound (fig. 4).

Coastline deltas of Lake Connecticut indicate a  
minimum value of glacioisostatic tilt

Deltas east of Clinton (fig. 4) are slightly older and were built at slightly higher lake levels than those to the west. An ice-marginal delta at Lordship Point, southwest of New Haven, has a topset-foreset contact at about 15 ft altitude (Hokans, 1952) and a surface altitude of about 25 ft. Deltas just to the north in Bridgeport and Stratford have surface altitudes of about 15 ft. The topset-foreset contact at Bridgeport, estimated from testhole log descriptions, is at 10 ft in altitude. Topset-foreset contacts are at about sea level at Fairfield. The New Haven and East Haven deltas are the youngest of the coastal deltas because they lie farther north than the others and because of persistence of the Connecticut Valley ice lobe from which they were deposited. The Mill River meltwater terrace (fig. 2, unit mt) is cut into the New Haven delta and has a surface altitude of 20-30 ft. Fluvial sand and gravel deposits of this terrace cut into the foreset beds of the New Haven delta down to at least 19 ft in altitude as observed in a recent foundation excavation on State Street. The meltwater erosion and deposition of reworked sediments along the Mill River terrace occurred between the time of completion of New Haven-delta building and the time when the ice margin stood just north of the Mill River-Quinnipiac River drainage divide at Cheshire. After this time, all meltwater drainage was diverted out of Mill River valley and into the valley of the Quinnipiac. Fluvial erosion and redeposition down to an altitude of 19 ft requires that the lake level had dropped by this time to at least 11 ft below the 30-ft level estimated for the New Haven delta.

New Haven lies 7.5 mi (12 km) farther up the isobases of the tilted water plane than Fairfield; the altitudes of 0 ft at Fairfield, 10 ft at Bridgeport, and 30 ft at New Haven give a tilt amount of 4 ft/mi (.76 m/km) if the lake level remained stable during time of ice retreat and delta building between Fairfield and New Haven. Because the lake level was slowly lowering during this time, the total amount of tilt is probably greater.

Glacial Lake Connecticut spilled across either or both the Harbor Hill and the Ronkonkoma-Block Island moraines in the now submerged parts of these moraines east of Long Island and west of Fishers Island at The Race, and west of Block Island at the head of Block Channel (fig. 4). The tilted lake water plane projects to altitudes below the restored crests of both moraines which were deeply eroded by drainage of the lake water at The Race and Block Channel, and possibly by Holocene tidal scour as well at The Race. The projected water plane is at -40 ft (-12.2 m) MSL at The Race, -72 ft (-22 m) MSL at the north shore of Long Island south-southeast of New Haven, and at -92 ft (-28 m) MSL at Block Channel.

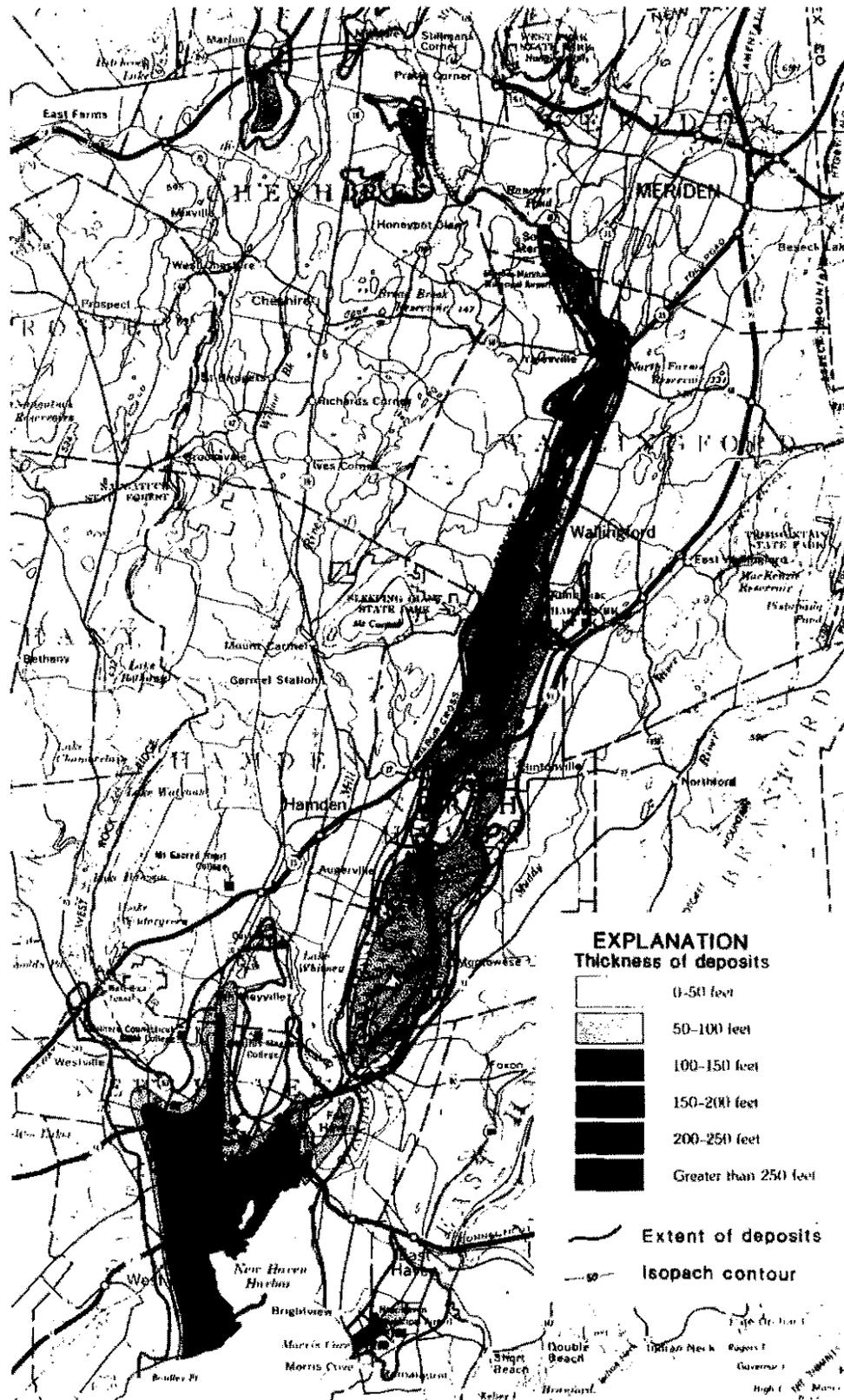


Figure 7.—Map showing distribution and thickness of fine-grained lake-bottom deposits of glacial Lake Connecticut (beneath New Haven), glacial Lake Quinnipiac (Quinnipiac clay in southern part of Quinnipiac River valley), and glacial Lake Southington (north of Quinnipiac Gorge). Modified from Langer (1979). Map shows same area as fig. 2; scale 1 inch equals about 2.6 miles.

A name for the glacial lake in Long Island Sound basin,  
here called glacial Lake Connecticut

The existence of a glacial lake in Long Island Sound was first postulated by Antevs (1922), who envisioned that the water plane recorded by lake deposits had been tilted by postglacial rebound, and was at about 40 ft in altitude at New Haven and projected (with a slope of 8 ft/mi) to about 150 ft below sea level at a presumed spillway across the moraine. Both Antevs and Reeds (1929), who studied lake clays in the New York City area, believed that the clay in the lower Quinnipiac valley had been deposited in the Long Island Sound lake. Reeds (1927) referred to this lake as glacial Lake Flushing although he showed the waterbody continuous with lakes Hudson and Hackensack. The clay at Flushing Meadow, N.Y., was later described by Newman (1977) who used the term "Flushing Formation."

In compilation of the glacial geology of Connecticut for the State Maps, the name "glacial Lake Connecticut" has come into use as it became apparent that the lake occupied the basin at a high level throughout Long Island Sound. It is becoming more apparent however, with recent and continued mapping of offshore areas, that this lake occupied not just Long Island Sound, but Block Island and Rhode Island Sounds as well. It seems therefore, that glacial Lake Connecticut may be a name that is too geographically restrictive, and a better one might be in order; two possibilities are "glacial Lake Antevs" or "glacial Lake of the Sounds."

### GLACIAL LAKE QUINNIPIAC

The bedrock floor of the lower Quinnipiac River valley lies at altitudes of more than 250 ft (76 m) below sea level in its deepest parts. A thick stratigraphic section of glacial and postglacial deposits fills the bedrock valley to about 100 ft (30.5 m) in altitude at the north end near South Meriden and to about sea level at the south end (fig. 8a). The Lake Quinnipiac deposits (fig. 2, unit 1q) consist of: 1) ice-marginal sand and gravel deposits which include the "Muddy River delta" of Flint, (1964), and "ice-contact stratified drift" of Porter (1960); and 2) lake-bottom deposits that underlie, intertongue with, and overlie the ice-marginal sediments. These lake-bottom deposits are here and in Ward (1920) and Loughlin (1905) referred to as the Quinnipiac clay, although in the past the deposit has been called New Haven clay (Flint, 1933, and later references). The Lake Quinnipiac deposits are largely overlain by younger deposits which include: distal meltwater-laid fluvial outwash deposits of the Quinnipiac valley terrace; postglacial stream-terrace deposits; alluvial deposits on the modern floodplain; and peat deposits of the transgressing estuarine marshes. These younger deposits will be discussed in later sections.

The ice-marginal deltaic and lake-bottom sediments of the lower Quinnipiac valley lie at higher altitudes than those at New Haven-West Haven-East Haven that are associated with glacial Lake Connecticut in the Sound basin. Much of the higher altitude is due to postglacial tilting, however water-level indicators provide evidence that Lake Quinnipiac occupied the Quinnipiac valley at a slightly higher level than Lake Connecticut at least during early stages of ice retreat in that valley.

The Muddy River delta, with a surface altitude of 50-60 ft, is the oldest, most extensive, and the best preserved of the Lake Quinnipiac ice-marginal deposits. Because it lies at the southern end of the valley (in the vicinity of Montowese), where the fluvial gradient of the younger valley terrace deposits slopes to lower altitudes, it has not been buried by the younger deposits as has happened to other deposits in most of the valley to the north. The course of the present Muddy River at the north edge of the delta probably follows a "fosse" or collapsed slope marking the position of the ice margin during deposition of the delta. Flowtill in the northernmost exposures of the delta (Lougee, 1938, p. 26; Flint, 1964, p. 20) also indicates that the ice stood along what is now the

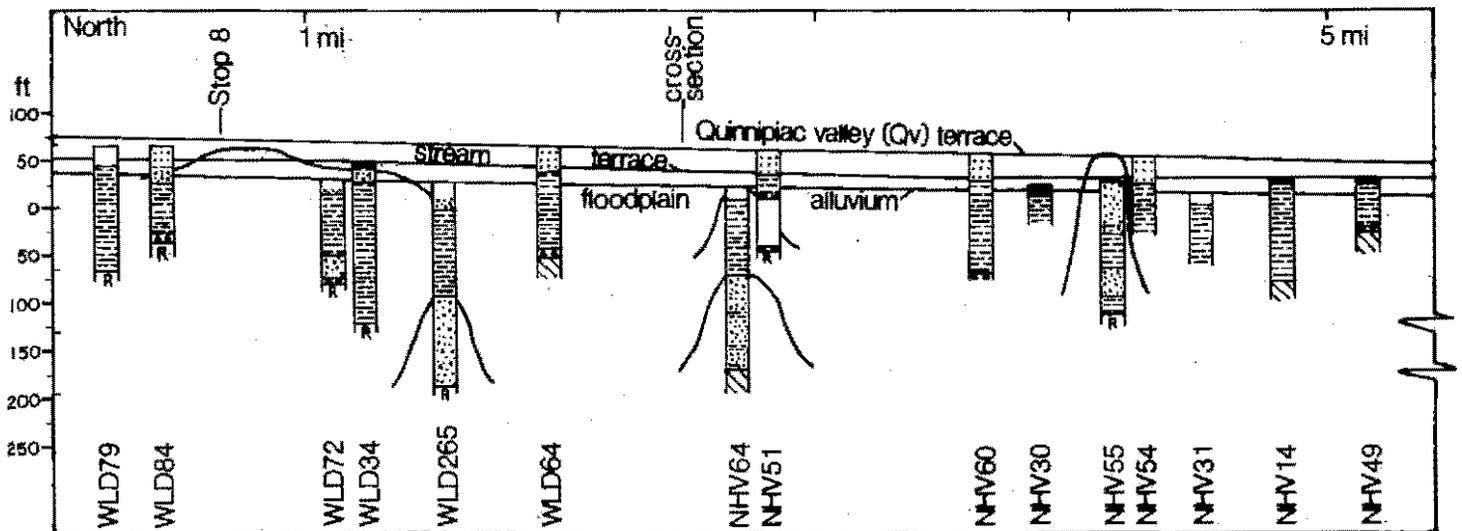


Figure 8a.—Longitudinal profiles of glacial and postglacial deposits in the Quinnipiac River valley from west of Wallingford center to Fair Haven; constructed by projection of highest surface altitudes of each deposit to valley center line. Surfaces of some buried ice-marginal deposits are shown schematically by curved lines. Test hole and well logs (numbers and letter designations for towns as in Mazzaferro, 1973; Haeni and Anderson, 1980) shown graphically, indicate the thickness of the deposits. Test hole log NHN58 at the south end of the profile shows the altitude of the bedrock surface at -13 m (-43 ft) MSL beneath the present channel of the Quinnipiac River east of Fair Haven. This altitude was the lowest possible base level for postglacial terracing.

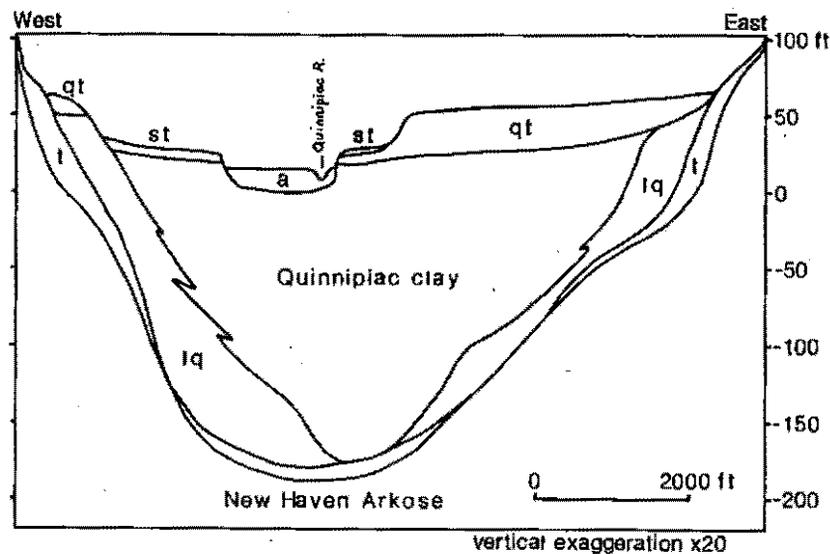


Figure 8b.—Generalized cross-section of glacial and postglacial deposits across the Quinnipiac River valley just south of Wallingford-North Haven, Wallingford-Hamden town lines. Units: lq - ice-marginal deposits of glacial Lake Quinnipiac; qt - fluvial outwash of the Quinnipiac valley terrace; st - postglacial stream terrace deposits; a, Holocene alluvium; t - till.

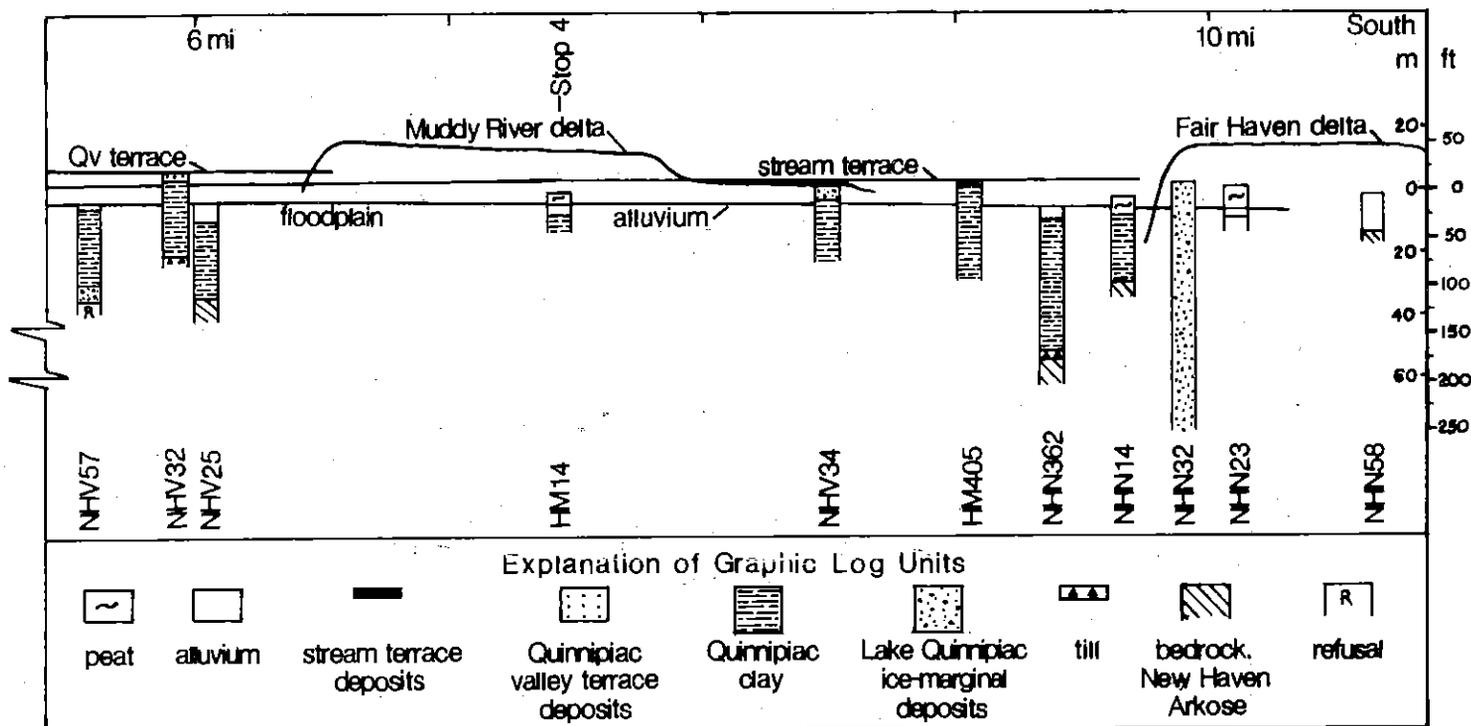


Figure 8a.—continued

north edge of the delta. Delta-foreset dip directions are predominantly west to southwesterly (Flint, 1964) indicating that sediment was contributed primarily by meltwater streams that entered the lake from the Muddy River valley upstream to the northeast. There are no known exposures at present of the topset-foreset contact of this delta, although we estimate that it is 40-50 ft in altitude. Foresets are exposed to an altitude of 30 ft in the remnants of the "once spectacular" Elm City Construction Co. pit (Stop 3). Lougee (1938, p. 25) described a topset-foreset contact in the northeast part of the delta at 44 ft in altitude; however, his photograph of "both topsets and foresets...of reddish sand" (Lougee, 1938, p. 25 and Plate III B) appears to us to show only cross-cutting foreset beds. Flint (1964) was of the opinion that the surface of the delta might have been planed off by post-delta streams, because "at every place where the delta sediments were seen exposed extensively, including four of the highest places on the surface of the delta, the topset beds which every delta possesses were missing." The four highest places on the delta ranged from 50-60 ft in altitude. Flint's photograph (1964, plate 4, p. 43) from the Elm City Construction Co. pit shows cross-cutting lobes of foreset beds to the top of pit face, with no topsets apparent; but whether this pit face was cut into the 60 ft surface shown on the topographic base of the Branford quadrangle map is not known. A water level of 60 ft or higher at this locality is difficult to postulate because this altitude, adjusting for postglacial tilt, projects to the vicinity of Fair Haven at about 50 ft. Because progradation of the Fair Haven delta lobe into Lake Connecticut across the southern end of the Quinnipiac valley provided the dam for the lake, the water level could be no higher than the 40-ft surface of the Fair Haven delta-plain dam. At that time the dam must have extended 150-240 m (500-800 ft) farther east to the till slope on the east side of the present course of the Quinnipiac River. The 40-ft altitude, adjusted for tilt, projects to the Elm City Construction Co. pit at about 50 ft. Because the water that spilled from Lake Quinnipiac flowed over easily eroded sediments, the spillway floor probably lowered rapidly nearly to the level of Lake Connecticut to the south.

Ice-marginal deposits of Lake Quinnipiac north of the Muddy River delta are largely buried by the younger valley terrace deposits. The Lake Quinnipiac deposits overlie, intertongue with, and are overlain by lacustrine silt and clay. A past exposure in a body of ice-marginal sediment near North Haven village, the 58-ft surface of which protrudes through the terrace sediments, showed deltaic foreset bedding as high as 50 ft in altitude (Lougee, 1938, p. 26). Between North Haven and Yalesville the ice-marginal deposits lie mainly on the sides of the valley; their collapsed slopes, toward the center of the valley, are overlain by lake-bottom sediments (fig. 8b). On the extreme west edge of the valley, postglacial terracing has exposed narrow segments of the ice-marginal deposits from beneath the valley terrace sediments at altitudes of 50-75 ft (Porter, 1960). Several exposures in these segments (e.g. Stop 8) show collapse-deformed beds of sand, silt and pebble gravel of probable lacustrine origin. At the north end of the valley, in the vicinity of South Meriden, exposures to a depth of about 6.1 m (20 ft) have shown only fluvial gravel which is not collapsed (Stop 9). Subsurface data indicate that the gravel overlies as much as 24.4 m (80 ft) of lake-bottom sediment. Lougee (1938) described steeply dipping foreset beds of red gravel beneath the yellow terrace sediment half a mile south of Hanover Pond. The pebble-cobble gravel deposits at South Meriden have an ice-marginal head north of Meriden (fig. 2). Subsurface data made available after quadrangle mapping was done show that the bulk of material that fills the lower Quinnipiac valley is lake-bottom silt, clay and fine sand (figs. 8, 7) rather than coarse-grained ice-contact deposits as indicated by Porter (1960, p. 9).

The extent and thickness of the Lake Quinnipiac bottom sediment is shown in figure 7. The Quinnipiac clay extends from just north of the Fair Haven delta plain northward to South Meriden. The deposit is as much as 60.1 m (200 ft) thick in the deepest part of the valley west of the village of Quinnipiac, but the clays average 15.2-30.5 m (50-100 ft) in thickness over much of the area (fig. 7). In the past as much as 5 m has been exposed in various clay pits in the towns of Hamden and North Haven. Since this body of varved clay is not exposed at the surface in the city of New Haven and since well and testhole logs indicate that it pinches out in the subsurface less than half a mile south of the New Haven city line (fig. 7), it is not clear why it was called "New Haven clay" by Flint (1933). The thick body of lake-bottom sediment that now is known to lie beneath New Haven (fig. 7) more properly might be called the New Haven clay.

The Quinnipiac clay deposit consists in order of decreasing abundance of silt, clay and very fine sand, dark reddish brown 5YR3/4-reddish brown 5YR4/3 in color. Where exposed in the clay pits, it is "varved" in silt-dominated couplets that average about an inch (2.3 cm) in thickness. Although the clays have not been exposed in recent years, they were well-exposed during brick-making operations in the past (Ward, 1920, p. 54):

"The clay ... is of so fine a texture that no grit can be felt by the teeth ... These layers are a quarter to half an inch thick. Silt is also of a fine texture. It has some true clay in it, but it consists essentially of silt and a certain amount of very fine sand. It feels gritty to the teeth ... At the bottom of each layer there is likely to be a thin lamina of small scattered grains of sand. The thickness of silt varies more than clay; ordinarily it is half an inch to 1 1/2 inches thick, in a few instances attaining a thickness of 4 1/2 inches and rarely even several feet. These layers may show a subdivision into very thin laminae, about fifteen to half an inch. The contact of the lower part of the silt and the upper part of the clay layers is uniformly regular, distinct and abrupt; but the lower part of the clay layers grades into the upper part of the silt."

### GLACIAL LAKE SOUTHINGTON

Deposits of Lake Southington fill the Quinnipiac-Farmington lowland northwest of the Quinnipiac Gorge and through the town of Southington. These deposits are predominantly ice-marginal deltas and lake-bottom sediments laid down in successive morphosequences as the ice retreated northward (figs. 1 and 2 of Stone and others, 1982). Delta surfaces range from 195-210 ft altitude in the southern part of lake basin to 235-245 ft in the north. The oldest deposits abut slightly older ice-marginal deltaic deposits that filled small basins in three north-draining tributary valley to the Quinnipiac River (fig. 2, small dot pattern). The ice-marginal ponds in the small basins spilled through cols across the Quinnipiac River-Mill River drainage divide at 245 ft from the Broad Brook valley, 195 ft from the Honeypot Brook valley and 186 ft in the valley of an unnamed brook west of Honeypot Brook. Deposits of the ice-marginal pond in Broad Brook valley, having surface altitudes in noncollapsed areas of 200-250 ft, lie across the Quinnipiac valley northwest of the gorge (fig. 9 and Stop 10). These deposits possibly formed part of the "dam" that impounded glacial Lake Southington. At the present Quinnipiac Gorge, the river flows through a bedrock-floored and -walled canyon (fig. 9) at 90-100 ft in altitude. In order for Lake Southington to have existed with a water plane at 180-185 ft at the latitude of the Gorge, the Quinnipiac Gorge either did not yet exist, or was filled with till, ice, stratified deposits or a combination thereof. Lougee (1938, p. 39-42) proposed that the rock gorge was carved by headward erosion of a late-glacial



Figure 9.—Quinnipiac Gorge, a possible late-glacial drainage diversion in the Meriden 7.5' quadrangle, scale 1:24,000; from Hanshaw (1962). Ruled pattern within the Gorge indicates area of abundant bedrock outcrop.

waterfall. The surface level of glacial deposits in the valley to the south is about 24 m (80 ft) below the level of Lake Southington, and it is conceivable that headward erosion by a waterfall of this magnitude could have resulted in the removal of 15-18 m (50-60 ft) of weathered arkose similar to that seen in the Gorge today. If this were the case, the Quinnipiac River may not have flowed southward prior to the last glaciation, but instead may have been a north-flowing tributary to the Farmington River.

The deposits of Lake Southington include the conspicuous ice-marginal deltas in Cheshire and near Plantsville, and lake-bottom sediments in front of and locally beneath the deltas. The deltas are spaced at 2.5-km (1.5 mi) intervals in the southern part of the lake basin and record stagnation-zone retreat of the ice margin northward through the lowland. Lake clay is as thick as 33 m (108 ft) in the narrow basin south of the distal part of deltaic deposits at Marion (figs. 2 and 7). Clay south of the proglacial delta in Cheshire, 1 km (0.6 mi) southeast of Milldale (discussed below) is as thick as 27 m (88 ft) (figs. 2 and 7).

At some time after the deposition of the last ice-marginal delta into Lake Southington and the development of glacial Lake Farmington to the north, the dam at the Quinnipiac Gorge was lowered, and the Southington deposits were terraced and eroded, first by water that spilled from Lake Farmington and then by meltwater that deposited the Quinnipiac valley terrace sediments (see discussion below).

#### THE "LOUGEE DELTA" OF GLACIAL LAKE SOUTHINGTON, CHESHIRE

One of the best examples of the morphology of ice-marginal glaciolacustrine deltas of New England is the proglacial, esker-fed delta of Lake Southington in the northern part of Cheshire (fig. 10). R. J. Lougee's (1938) plane-table map of this delta, using a 5-ft contour interval, shows the fine detail of its morphology, and his map matches well with the modern topographic maps of the delta and adjacent deposits. The "Lougee delta" ideally exhibits the characteristic continuity of ice-proximal to -distal topographic elements that compose such ice-marginal deposits in the region. The delta thus includes the ice-channel filling (esker), an irregular ice-contact scarp, a zone of kettles and collapsed deformation, a noncollapsed fluvial plane, and a distal margin containing distributary lobes, the lower slopes of which tangentially merge with a lake-bottom plane to the south. The continuity of such forms in a single mappable sedimentary unit was recognized nearly a century ago (Davis, 1890), used in early studies of glacial lake successions (Goldthwait, 1905; Alden, 1924), and exemplified in the first detailed quadrangle maps of glacial deposits (Jahns, 1953; Richmond, 1953). Some of these deltas have been referred to variously as **proglacial deltas** (Lougee, 1938), **esker-fed deltas** (Stone and others, 1982), **kame deltas** (Richmond, 1953; La Sala, 1961), **ice-contact deltas**, **ice-marginal deltas** (Stone and Force, 1982), or **ice-contact lacustrine morphosequences** (in the morphogenetic classification of Koteff, 1974; see also Koteff and Pessi, 1981). Most of these terms seem appropriate, the more descriptive, i.e., **ice-marginal deltas**, are preferred by us. The term **kame delta** uses the word **kame** as a modifier, synonymous with **ice contact**. To some geologists, the term **kame** connotes ice hole deposition (Bates and Jackson, 1980), as in **moulin kame**. Transferral of this connotation to studies of deltas like the "Lougee delta" obscures the depositional setting and stratigraphic relationships of such deltas. These ice-marginal deltas are bounded by ice contacts at their northerly margins, and their free-front, lobate distal margins contain foreset strata that grade tangentially into bottomset beds in front of them. Further, the extent of lake bottom sediments around and among related deltas indicates areas of open water in the basin, and topset-foreset contact altitudes of related deltas define a single plane that closely approximates the water plane of the glacial lake.

The "Lougee delta" (fig. 10) includes map unit Qkd<sub>g</sub> and part of unit Qk<sub>g</sub> of La Sala (1961). Adjacent deposits of Lake Southington, are other parts of La Sala's unit Qk<sub>g</sub>. At the west end of the ice contact of the delta is a small rounded hill that is just over 210 ft in altitude, and which is bounded on its northern and eastern sides by steep ice contacts. This hill appears to project above the depositional gradient of the delta, and is possibly a slightly older, chiefly ice-bordered deposit. The adjacent flat-topped 200-ft hill, welded on to its western side, also appears not to be part of the main delta.

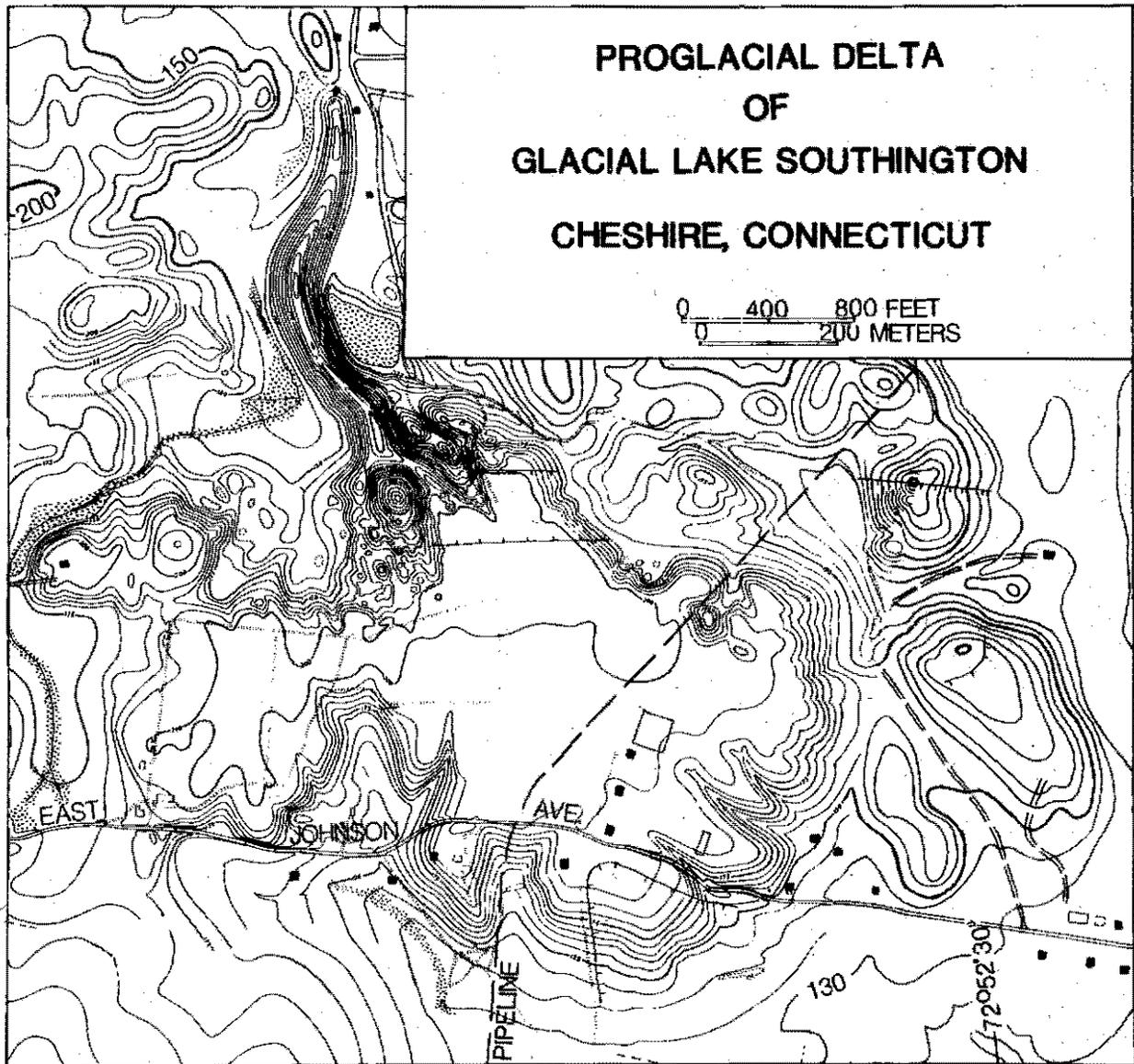


Figure 10.—Topography of the "Lougee delta" of glacial Lake Southington. Topography in area of 5-ft contours, some boundaries of which are shown by dashed lines at edge of map, is modified from Lougee (1938, Plate XV). Topography in areas of 10-ft contours is modified from the Southington and Meriden 7 1/2' topographic quadrangles. Hachured lines show partial extent of large excavations.

Besides the ice-channel filling, the "Lougee delta" at Cheshire is composed of two parts, as previously recognized (Lougee, 1938, p. 38): 1) a flat-topped western part, the surface of which is everywhere above 195 ft and mostly above 200 ft; and 2) a larger flat-topped eastern part, the surface of which extends from the ice-channel filling at altitudes above 200 ft to the lobate southern margin at altitudes of 185-195 ft.

From Lougee's finely detailed topographic map some observations and measurements of features related to processes that formed this ideal example can be made. The delta-top fluvial plain slopes from above 204 ft at the proximal end to 185-195 ft at the distal margin. Depositional slopes on the plain are 0.0087 (8.7 m/km; 46 ft/mi) - 0.0106 (56 ft/mi) at the proximal part above 200 ft altitude. The slopes are about 0.0045 (24 ft/mi) in the middle reaches between altitudes of 195-200 ft. Fluvial slopes at the distal lobate margin cannot be calculated because the precise altitude of the top and bottom of fluvial beds is not known, and the sediments in the distributary lobes are not well exposed. Lougee (1938, p. 38) previously estimated the distal edge of the fluvial plain to be at 193-ft on the basis of morphology. Distributary lobes of the eastern delta plain vary in length/width dimensions from 152.4/76.2 m (500 ft/250 ft) to 91.5/39.6 m (300 ft/130 ft), as measured from altitudes of about 193 ft to the 185-ft contour (length), and orthogonally from the 185-ft contour at the midpoint of the lobes. Interlobate troughs are V-shaped and widen from less than 15.2 m (50 ft) at their heads to about 53.3 m (175 ft) at their mouths.

The size of gravel clasts in the delta-topset beds fines remarkably from the ice contact to the lobate margin. Small rounded boulders at the ice-contact head of the delta have long axes of 30-50 cm. Cobbles in the middle reaches of the delta have long axes of 20-32 cm. Clasts with long axes of about 10 cm are common in the surface gravel of some of the distributary lobes.

Past exposures in parts of the esker-fed delta and adjacent deposits showed coarse gravel-dominated topset beds overlying steeply inclined delta-foreset strata of coarse sand, and granule-pebble gravel (stop 11). The sequence of fluvial beds beneath the sloping delta-topset plain is bounded at the base by a sharp, horizontal unconformity that erosionally truncates the underlying sandy foreset strata. This unconformity is approximately at the altitude of the paleolake level (Gilbert, 1885; Lougee, 1938; Stone and Force, 1982). Pits on either side of the pipeline at the southern foreset slope of the delta previously exposed coarse sand and pebble gravel foreset beds, the products of depositional avalanching of debris down the foreset slope, and minor interbedded sand beds that contain sets and cosets of climbing-ripple cross-laminations. An oblique cut through foreset strata in the elongate hill just east of the delta and 300 m (984 ft) north of East Johnson Ave. (fig. 10) showed cross-cutting sets of proximal foreset strata composed of sand and pebble and cobble gravel in beds that dipped southwesterly 25-30° (fig. 11). A conspicuous bed of indistinctly bedded sand and cobble-boulder gravel probably resulted from a subaqueous gravity flow or slide on the foreset slope. The proximal delta foresets extend to the ice contact at the north side of the pit; backset beds (Davis, 1890) have not been seen at the ice contact. Also, neither flowtills (Hartshorn, 1958) nor other matrix-supported diamict sediments were observed by participants in the 1982 NEIGC fieldtrip (Stone and others, 1982) to this exposure, so the presence of these nonsorted, probable ice-slope-derived sediments is not always obvious in ice-marginal deltaic sequences. Other delta-foreset strata containing characteristic climbing-ripple cross-laminations and associated draped laminations, as well as sets of planar laminations of coarse sand, have been exposed until the present in the more distal southwesterly part of the pit.

Highway excavations in 1983 in the ice-contact head of the "Lougee delta" (fig. 10) revealed about 4 m of horizontally bedded gravel topset strata. Beds were generally less than 1 m thick. Coarse cobble gravel was in erosionally bounded, channel-filling beds commonly 6-8 m wide. Other beds of pebbly sand and small cobble gravel, in sets of

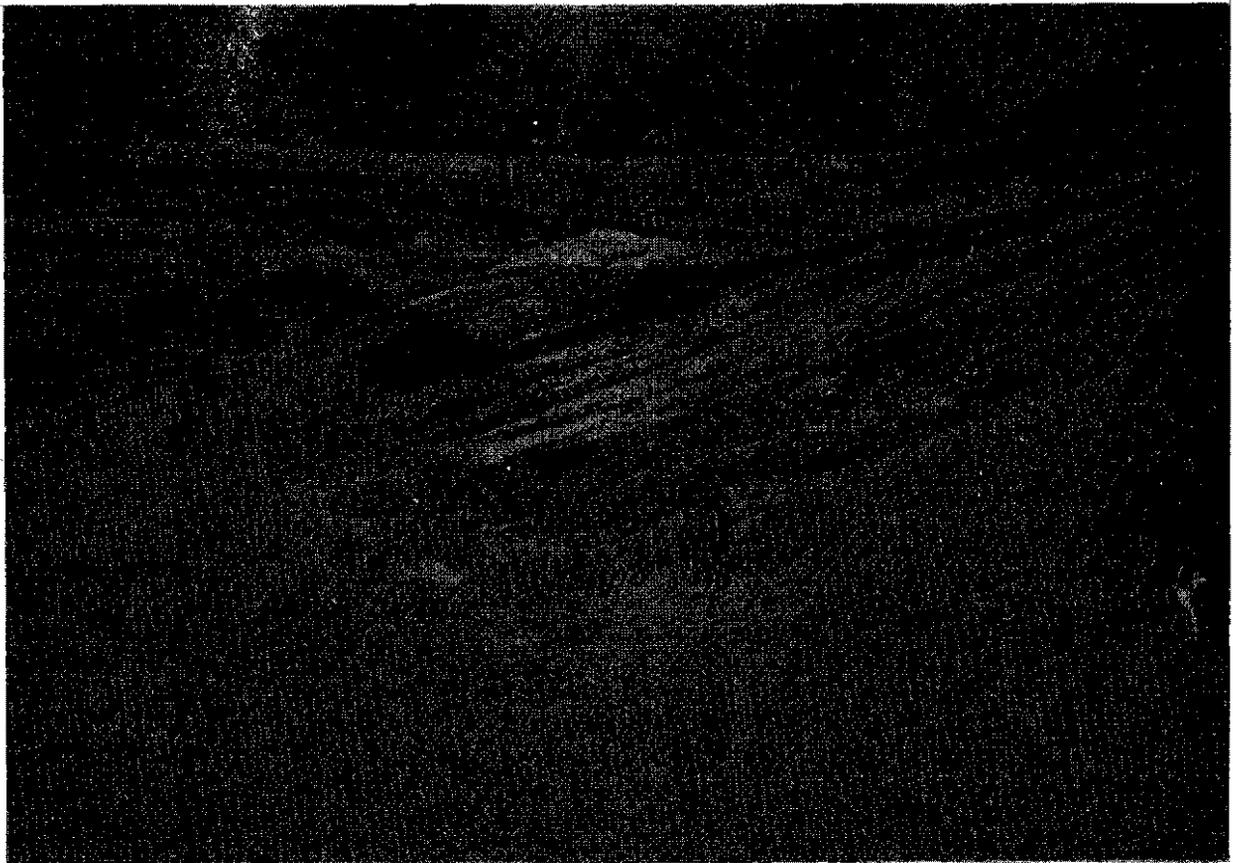


Figure 11.—Delta foreset strata exposed in elongate hill east of the "Lougee delta" (fig. 10). Oblique view of delta foresets that dip southwesterly. Topset beds have been excavated. Note trough-filling set of foreset strata near top of section. Delta ice contact is less than 25 m to right of view.

tabular cross-beds 0.3 to 0.6 m thick, were exposed along the length of the excavation. At places in these cuts, and in a cut in the part of the ice-channel filling that was on the east side of the prominent circular kettle, coarse sand in inclined delta-foreset strata was exposed. A single set of steeply inclined delta foresets was exposed to a total thickness of 3.5 m in one place. The occurrence of foreset strata and unconformably overlying fluvial topset beds within the ice-channel-filling feeder, indicates that the deltaic depositional environment extended into the channel between walls of stagnant ice. The altitudes of the delta topset-foreset contact in the channel and the top of the surface of the deposit indicate that the lake level in the channel filling was at the same level to which the rest of the "Lougee delta" was graded, thus showing that deltaic sediments filled an ice channel that extended to till or bedrock at its base. Furthermore, the continuity and noncollapsed topography of the proximal part of the delta indicate continuous progradational sedimentation of the upper part of the delta southward from the ice-channel mouth to the lobate margin of the delta.

#### THE QUINNIPIAC VALLEY TERRACE

Like other meltwater terraces in the State, this terrace deposit was laid down by distal meltwater that eroded and reworked slightly older, more proximally deposited sediments. The Quinnipiac valley terrace deposit is unique because of its extent,

continuity, and composition distinct from the deposits that it overlies. The terrace sands and gravels are a fluvial unit that extends through the lowland from Farmington, where the surface is at 215 ft in altitude, about 47 km (29 mi) southward to North Haven where the surface is at 35 ft. The deposit probably once extended farther south but has been removed by postglacial erosion. Presumably some of the material that overlies lake-bottom deposits of Lake Connecticut in Long Island Sound basin off New Haven is a continuation of the valley terrace deposit.

The color of this deposit ranges from yellowish gray to light gray in its upper part; it approaches reddish gray where contaminated by reworked ice-contact sediments, most commonly in lower parts. The yellowish color of the valley terrace sediments is due to derivation in large part from the crystalline rocks of the Western Highland. The difference in color between the terrace deposit and the ice-contact sediments was recognized in early work by Dana (1875, p. 179-180), Flint (1934), and Lougee (1938). Krynine (1937) showed by heavy mineral analysis that the red or pink ice-contact deposits are characterized by a hornblende-kyanite ratio of  $>1.6$  and a hornblende-garnet ratio  $>1.0$ , and that their principal source was from local Mesozoic bedrock. The later generation of "buff" or yellow "ice-free" sediments where uncontaminated by reworked pink sediments is characterized by a hornblende-kyanite ratio of  $>9.0$  and a hornblende-garnet ratio of  $>4.5$ ; the provenance for these is the crystalline rocks of the Farmington and Pequabuck drainage areas in the Western Highland.

The Quinnipiac valley terrace slopes southward at 0.95-1.14 m/km (5-6 ft/mi). The fluvial gradient during deposition, however, was only 0.38-0.47 m/km (2-2.5 ft/mi), the greater present slope being a result of postglacial tilting. Evidence from elsewhere in southern New England (Koteff and Larsen, 1985) indicates that crustal rebound did not begin until well after deposition of the terrace deposits.

Textures within the deposit vary both laterally and vertically but exhibit an overall decrease in grain-size downstream. In Farmington at the upstream end of the deposit, the average largest clast sizes are 10-14 cm long; at South Meriden the average largest clast size is about 5 cm long. Through the length of the Wallingford quadrangle, where the valley terrace deposits are most extensive, Porter (1960) showed a decrease in grain-size from fine pebbles and coarse sand at the north end of the quadrangle to medium sand at the south end, which is close to the southernmost present surface exposure of the deposit.

The internal structure reveals characteristic cut-and-fill stratification within horizontally layered fluvial beds as is typical in braided-stream deposits. Cross-bedding indicates southward current directions (see Stops 5, 6, 7, 8, and 9). Thicknesses ranges from less than 1 m to as much as 10 m. The deposit is generally thickest in the middle of the valley and thins toward the margins. These sediments lie on an erosional surface which in the north half of the valley is cut into mostly deltaic deposits of Lake Farmington and Lake Southington. Here the lake deposits stand at higher altitudes than the terrace. South of the Quinnipiac Gorge, the valley terrace deposits unconformably overlie and bury most of the Lake Quinnipiac ice-marginal and lake-bottom sediments. Several past exposures of the unconformable contact between terrace sands and Lake Quinnipiac sediments provide evidence that the pre-terrace surface was subaerially exposed and underwent a period (however short) of colluviation, wind abrasion and fluvial erosion prior to the arrival of terrace sediments (Dana, 1875, p. 179-180, 419; Hartshorn and Schafer, 1965, p. 9-10; Flint, 1964, p. 25). Porter (1960, p. 15 and fig. 7) described from many exposures a lag gravel at the unconformable contact which he attributed to "the winnowing of fine sediment during a period of fluvial erosion prior to outwash sedimentation." A pebble gravel unit in similar stratigraphic position can be seen at Stop 8 where it is clearly part of the terrace deposit.

Meltwater erosion and reworking of the surface of Lake Quinnipiac deposits could have begun south of the Quinnipiac Gorge during the life of glacial Lake Southington. Meltwater that spilled from Lake Southington across the dam at the Gorge (possibly after

dropping in an 80-ft waterfall) could have begun the terracing process as soon as the level of Lake Quinnipiac had lowered enough to expose the lake bed. North of the Gorge, through the town of Southington, terrace development began after Lake Southington drained; here early erosion and reworking were accomplished by meltwater that spilled from Lake Farmington. During the life of Lake Farmington, the Pequabuck and upper Farmington River drainages were deglaciated. These were the source areas for the crystalline-derived material that in large part comprises the Quinnipiac valley terrace deposit. But until Lake Farmington drained, all sediment carried by the Pequabuck and upper Farmington meltwater streams was deposited in the lake as deltas and lake-bottom sediments. Only after Lake Farmington drained could the yellow crystalline-derived material be carried and deposited southward. Crystalline-derived material, mixed with varying amounts of reworked older, red or pink sediments was deposited along the entrenched terrace surface through the Lake Southington deposits. It was carried through the Quinnipiac Gorge and out onto the surface of the Lake Quinnipiac deposits. In the southern Quinnipiac valley, deeper cuts in the terrace deposit commonly show a high proportion of reworked red material at depth (Stop 7).

The surface of the Quinnipiac valley terrace appears to be a depositional surface rather than an erosional one. Regional mapping and correlation do not support the idea (Flint, 1964, 1965) that the valley terrace deposit is but an erosional remnant of a once more extensive and greatly thicker deposit.

#### POSTGLACIAL DEPOSITS

Postglacial degradation of the glacial deposits in the Quinnipiac-Farmington lowland probably began soon after abandonment of this valley as a meltwater drainageway, perhaps about 17,000 radiocarbon years ago. Thin stream terrace deposits (fig. 2, unit st) along the Quinnipiac River are remnants of early, shallow entrenchment. These terraces are most extensive south of the Quinnipiac Gorge through the Wallingford quadrangle. Here the stream terrace consists of non-paired terraces produced by a meandering stream (Porter, 1960), cut mostly into the Quinnipiac valley meltwater terrace surface. The stream terrace surfaces are thinly veneered with commonly 0.3-0.6 m (1-2 ft), rarely 1.2-1.5 m (4-5 ft) of principally sand composed of reworked red and yellow sediments (Porter, 1960). In some places the stream terrace sediments overlie the meltwater terrace sediments. In other places the base of stream terrace deposits is slightly lower in altitude than the base of the valley terrace outwash (fig. 8), indicating that the stream terrace entrenchment cut slightly deeper into the surface of the glacial lake clay than did the meltwater terrace entrenchment. The deep erosional trench cut into the Quinnipiac clay probably also dates from early in postglacial history; it is now filled with alluvial and estuarine sediments (fig. 2, units a, sm). From the profile (fig. 8a) it appears that base level for the entrenched channel beneath the present flood-plain alluvium was the bedrock surface, at about -40 ft (-13 m) MSL in altitude beneath the present Quinnipiac River east of Fair Haven. The stream terrace segments and Quinnipiac valley terrace, by contrast, project to levels above present sea level at Fair Haven. The base level for these terraces was probably a somewhat lowered level of Lake Connecticut in Long Island Sound basin.

The floodplain alluvium consists mainly of sand and silt, moderate brown to dark reddish brown in color. In the clay pit (Stop 4), the sandy alluvium is light gray in color. In the upper Quinnipiac valley the alluvium is generally less than 3 m (10 ft) in thickness. In the southern part of the valley, thickness of alluvial and estuarine deposits is as much as 7.6 m (25 ft) in some places. Much organic plant material including logs and tree stumps, some of which were in growth position (R. W. Brown, in Flint, 1930, p. 263-266) occur in the alluvium in the southern part of the valley. In the clay pit (Stop 4) a log within the alluvium was dated at  $3560 \pm 80$  years B.P. and one from slightly lower in the section at  $6810 \pm 170$  years B.P. (Bloom and Stuiver, 1963). Older alluvial sediments may

be preserved locally.

Estuarine silt and peat and salt-marsh peat overlie alluvium in the southern part of the Quinnipiac valley (Stop 4; fig. 2, unit sm; and fig. 8). These were deposited near sea level, and are probably younger than about 3,000 years B.P., which is the time that the transgressing sea reached about -2.8 m in altitude (Bloom and Ellis, 1965, sea level curve p. 5).

The sedge peat seen in the upper part of the Stop 4 exposure is a deposit of an estuarine freshwater marsh where cattail, sedge and reed grow to high-tide level in an estuarine environment of low salinity (Bloom and Ellis, 1965). Salt-marsh grasses occur only farther south in the valley.

### SUMMARY OF LATE QUATERNARY HISTORY

The history of ice retreat and resultant meltwater drainage patterns through Connecticut has direct bearing on the sequence of deposits and position of the marine unconformity in the eastern and central parts of Long Island Sound. The thick sand and gravel deposit that conformably overlies the lake clay offshore from New Haven (fig. 5) in the central part of the Sound (fig. 4) is expectable in that meltwater-transported sediment reached Lake Connecticut during most of the time of ice-margin retreat through the Quinnipiac-Farmington lowland until deglaciation of the gap at Tariffville. This time interval may have spanned at least a thousand radiocarbon years. These sediments could have derived initially from the deltas built into the lake (fig. 2, units Cn, Ce), slightly later from extensions of the Mill River terrace deposits (fig. 2, unit mt), and still later from distal parts of the Quinnipiac valley terrace deposits (fig. 2, unit qt). The time interval for deposition of the Quinnipiac valley terrace deposits is dependent upon the change from southward flow to northward flow at Farmington of the Farmington valley meltwater drainage. This seems most likely to have happened when the Tariffville gap was deglaciated, since water levels would have dropped about 30.5 m (100 ft) at that time. It is possible, however, that northward diversion occurred later, which would allow a longer time for meltwater to bring sediment down the Quinnipiac-Farmington drainage route to Lake Connecticut. Meltwater continued to enter the upper Farmington River valley until the ice margin retreated into the Westfield River valley in Massachusetts.

The sequence of deposits in the eastern part of the Sound is different than that in the central part south of New Haven probably because a major sediment trap, glacial Lake Hitchcock, existed in the Connecticut lowland during deglaciation and for a long time following deglaciation. Lake Hitchcock persisted possibly until about 10,000 years ago (Flint, 1956; see discussion in Stone and others, 1982). By the time that Lake Hitchcock drained, Lake Connecticut had lowered considerably, and at least in the eastern part of the basin, the lake-bottom deposits had undergone subaerial exposure, fluvial erosion, and marine transgression. The sea probably came into the eastern part of the basin about 12,000 to 13,000 years B.P., based on the adjusted sea-level curve (Needell and Lewis, 1985). The complex sand and gravel deposit that lies on top of the marine unconformity offshore from the mouth of the Connecticut River in eastern Long Island Sound possibly is sediment that was derived from the terracing and reworking of Lake Hitchcock deposits after that lake had drained and before vegetation had become established. A more certain interpretation of this complex deposit, however, awaits more detailed investigation.

The provisional radiocarbon chronology of glacial and postglacial deposits in the southern New England-Long Island Sound area (Stone and Borns, in press) is based in part on dates from the organic fraction of both pre- and postglacial fine-grained elastic sediments. It is assumed that these whole-sediment dates accurately date the time of sediment deposition and that these dates are compatible with other dates obtained from macroscopic organic materials.

Following retreat from the Roanoke Point (Sirkin, 1982) segment of the Harbor

Hill moraine, central north shore of Long Island (fig. 4), perhaps about 19,000 years B.P., lacustrine sediments began to accumulate in the eastern Long Island Sound basin of Lake Connecticut. Ice-margin retreat in western Long Island Sound caused probable rapid expansion of the glacial lake into the western basin. Ice-margin positions associated with offshore moraines on the Norwalk and nearby Captain Islands are correlated generally with some of the onshore moraines of southeastern Connecticut (fig. 3, southernmost ice line). Further tentative correlation of these offshore moraines with ice recessional deposits in northern New Jersey that are younger than  $18,570 \pm 250$  B.P. (Cotter and others, 1984) indicates a minimum age of these moraines of about 18,000 radiocarbon years B.P. A whole-sediment date of  $17,950 \pm 620$  B.P. (Weiss, 1971) from glaciolacustrine sediments in the Hudson River channel north of the correlated moraines corroborates their minimum age. Deltas of Lake Connecticut along the Connecticut coast (fig. 4) are younger than the moraines; the New Haven delta is the youngest of these. Sediments of glacial Lake Quinnipiac including the varved Quinnipiac clay, and younger deltaic and lake bottom deposits of glacial lakes Southington and Farmington are older than about 17,000 radiocarbon years B.P. This age is suggested for deglaciation of west central Connecticut (fig. 3, northernmost ice line) and resultant abandonment of the southern Quinnipiac-Farmington lowland as a meltwater drainage route, by opening of the Tariffville gap. The Quinnipiac valley terrace and some of the stream terrace deposits are of about this age. Deep fluvial erosion into the Quinnipiac clay, succeeding alluvial sediment and some of the "stratified sand and gravel" overlying lake clay (fig. 4, fig. 5) in the central Long Island Sound basin are probably younger than about 16,000 B.P. The age of the late Pleistocene-early Holocene marine transgression in eastern and western Long Island Sound basins is estimated from a regionally adjusted sea-level-rise curve for the last 16,000 years (Needell and Lewis, 1985; Oldale and O'Hara, 1980; and Dillon and Oldale, 1978). Transgression of the sea into eastern Long Island Sound basin probably took place 12,000-13,000 years B.P. and transgression across the Branford high (see earlier discussion) somewhat later. A date on peat of  $10,200 \pm 400$  B.P. (Schaffel, 1971) at -39.6 m MSL from a borehole in western Long Island Sound is the lowest of three dated submerged peats from that area (Newman, 1977). This date indicates marine transgression in the western part of the Sound at -39.6 m MSL. Lower parts of some of the drowned floodplain alluvial deposits in the Quinnipiac River estuary are only a few thousand years old. The brackish water peat that caps the lower floodplain is related to the limit of the latest transgressive on-lap of the last 2000 years.

## FIELD TRIP STOPS

Although extensive exposures of surficial sediments are rare at present in the New Haven-Meriden urban area, small and reexcavated exposures show important sedimentologic and stratigraphic details. The field trip stops (fig. 2) are arranged in south-to-north order and reflect a general oldest-to-youngest depositional chronology of the ice-marginal glacial deposits. Almost all stops are on private property and permission should be obtained before visiting them. Stops are in the following quadrangles, which are covered by U.S. Geological Survey topographic maps (scale 1:24,000, contour interval 10 ft): New Haven, Branford, Wallingford, Meriden, and Southington. The Mount Carmel quadrangle is useful for travel between stops. References for published surficial geologic quadrangle maps for these quadrangles are: Flint (1965), Flint (1964), Porter (1960), Hanshaw (1962), and La Sala (1961), respectively.

**ASSEMBLY POINT.** East Rock Park, monument at East Rock, altitude 360 ft, city of New Haven, New Haven quadrangle. From State Street, 0.6 mi (1 km) north of the New Haven-Hamden town line, entrance to north end of park is reached by turning west onto Ridge Road, thence proceed 0.35 mi (0.6 km) to light at intersection with Davis Street. Turn left (west) onto Davis and proceed 0.1 mi (0.15 km) to park entrance on left. Winding park road leads to monument at crest of East Rock.

From this vantage point the Long Island Sound glacial lake basin and the north shore of Long Island can be seen to the south. To the southwest, smooth crests of south to southwest-trending drumlins at altitudes as high as 350 ft and which lie on the irregular crystalline bedrock surface west of the Mesozoic lowland can be seen. The broad New Haven delta plain, at about 45 ft altitude in downtown New Haven and which was graded to the glacial lake in Long Island Sound (glacial Lake Connecticut of this report), extends from west of Yale Bowl (seen to the west) to Fair Haven (seen to the south-southeast). The western scarp of the Mill River meltwater terrace that is incised into the New Haven delta plain is just behind the row of three churches on the green in downtown New Haven. To the east and northeast, the extensive salt and freshwater marshes of the Quinnipiac valley are dotted by ponds that fill abandoned brickyard pits in the Quinnipiac clay. The Fair Haven part of the New Haven delta overlies a deep bedrock valley. The Quinnipiac River flows between the eastern valley wall and this part of the New Haven delta; the original extent of the delta at Fair Haven dammed glacial meltwater in the Quinnipiac basin to the north, thus impounding glacial Lake Quinnipiac.

Striations and grooves preserved in the glacially polished diabase on top of East Rock record southwesterly ice flow. Thirty-six feet northwest of the stone block wall that is west of the monument, grooves 1-2 cm wide and 20 cm long trend S.19°W. Thirty feet further northwest, grooves 1-2 cm wide and 30 cm long trend S.32°W.

**STOP 1.** Till cut behind Elm City Subaru, town of Woodbridge, New Haven quadrangle. Cut is behind auto dealership, on west side of Amity Road, State Route 63; 0.4 mi (.64 km) north of Wilbur Cross Parkway, just west of Interchange 59.

This cut exposes superposed distinctive facies of the stratigraphically separate two tills of southern New England (Schafer and Hartshorn, 1965). It is not a typical two-till cut, however, in that the upper till of late Wisconsinan age is reddish brown, which contrasts with the gray color of the lower, older till. The upper till contains a red matrix and clasts derived from early Mesozoic rocks. This locality is just west of the western border of the Mesozoic rocks; bedrock beneath the cut is schist and greenstone of the upper part of Maltby Lakes Metavolcanics, and Wepawaug Schist lies a short distance to the northwest (Rodgers, 1985). Red New Haven Arkose of Triassic age lies to the east and northeast; intrusive diabase of prominent West Rock Ridge is conspicuous across the West River valley east of the cut. Thus the red matrix of upper till derives from southwesterly ice flow on the western side of the Connecticut valley ice lobe.

The cut shows: 1-2.5 m of sandy, stoney, reddish brown upper till, fairly compact with abundant diabase boulders, many of which are striated. Upper till overlies as much as 6 m of olive gray, silty, lower till with relatively few stones in massive basal parts; distinctive flat-iron-shaped stones in the lower till are dark gray Wepawaug Schist. The contact between the tills is highest in the middle of the cut and slopes down at both ends. At the south end of the cut, iron- and manganese-stained subhorizontal and subvertical surfaces bound angular, lenticular plates of gray lower till in the top half-meter of that till. At the north end, the iron/manganese-stained horizon of the upper part of the lower till is not present. Between the two tills is a zone of laminated and thinly bedded sand, silt, and clay. This zone is less than 1 m thick in the middle of the face where irregularly laminated silt and clay are intermixed with subhorizontal, lens-shaped blocks of gray lower till. Laminae of white fine-medium sand penetrate the included blocks of lower till. At the north end of the pit, the laminated zone is 0.5 m thick and contains fine sand beds 1 cm or less thick, silt beds less than 0.5 cm thick, and clay laminae less than about 3 mm thick. These beds are contorted, and one thrust fault showed 5 cm of horizontal offset in a southerly direction. The origin of the deformed bedded zone between the two tills may be related to the zone of shearing, brecciation, and glacial dislocation at this stratigraphic position reported elsewhere in western Connecticut (Pessl and Schafer, 1968).

**STOP 2.** Glacial grooves and crag and tail behind Amity Shopping Center, city of New Haven, New Haven quadrangle. Exposure is behind southeast end of shopping center building, on west side of Amity Road at end of Sunset Drive, just north of Wilbur Cross Parkway.

The glacially polished bedrock ledges are composed of fine-grained greenschist of the lower unit of Maltby Lakes Metavolcanics (Rodgers, 1985), which contains prominent nodules of quartz and epidote. Striations, grooves, and unique tails behind nodular crags trend S5-10W, oblique to the foliation. Some tails are as long as 2 m.

**STOP 3.** Muddy River delta pit at Elm City Construction Co., near village of Montowese, town of North Haven, Branford quadrangle. Entrance to pit is off Montowese Ave., at end of small road that parallels north-bound entrance ramp at Interchange 9 of Interstate Route 91. Pit is north of railroad overpass, 2000 ft (610 m) north of Montowese Ave.

This pit previously exposed crosscutting sets of foreset strata of the Muddy River delta (Flint, 1964, plate 4a, p. 43) of glacial Lake Quinnipiac. The original land surface at the pit probably was above 50 ft altitude. R. F. Flint's description and photograph of foreset strata were from an exposure now occupied by a pond at the north end of the property. A small exposure on the eastern side of the modern pit shows foreset strata as high as 30 ft altitude, as high as the surface of the adjacent rail yard. These foresets consist of coarse sand, and granule-pebble gravel beds, 0.5-15 cm thick, containing well-rounded outsized pebbles and small-to-medium cobbles of diabase and red sandstone. Some beds are inversely graded. Beds dip as much as 34° to the west. East of the rail yard, gravel underlies the 50-ft plain, but in the early 1960's Flint (1964, p. 21) saw no delta topset beds in the large exposure in the area of the present pond.

**STOP 4.** Pit in peat, alluvium and Quinnipiac clay at Hamden Building Salvage, Inc. (formerly Stiles Corporation clay pit of Bloom and Ellis, 1965, and Flint, 1965), town of Hamden, New Haven quadrangle. Entrance to pit is at 2891 State Street, 0.7 mi (1.1 km) south of Sackett Point Road intersection. Pit is south of old brickyard buildings east of the railroad.

The top of the Quinnipiac clay lies just below the standing water level in the clay pit at the time of this writing. The water must be pumped from the pit before the clay can be seen in place, however current excavation of clay below the water level has brought up partial sections of varved clay. About 6.5 m of Holocene alluvial and

estuarine sediments disconformably overlies the clay and these are well exposed in the present pit. A composite stratigraphic section, culled from the present exposure and from sections seen previously in this pit (Bloom and Ellis, 1965; Hartshorn and Schafer, 1965) and a section from the pit immediately to the north (R. W. Brown, in Flint, 1930, p. 263-266) is given.

Thickness (m) Description of units

1-5.1 Sedge peat; brown, compact, containing abundant fragments of cattail (*Typha*) and reed (*Phragmites*); black peat is locally at the surface, and gray muddy peat or peaty mud grades down to brown muddy peat. Plant remains include: grasses, leaves of deciduous trees, hickory nuts, beech nuts, and butternuts. The peat has been compacted by the overlying earthen dike.

—Sharp contact—

0.3-1 Sandy silt, gray, containing twigs, nuts, leaves, and logs; probable fluvial overbank deposit; a log from top of silt yielded a  $^{14}\text{C}$  date of  $3560 \pm 80$  B.P. (Y-1077).

—Gradational contact—

1-4 Medium to coarse sand, well-sorted, light gray, cross-bedded; contains thin beds of dark gray fine sand and silt with organic matter, and twigs and logs; a log yielded a  $^{14}\text{C}$  date of  $6810 \pm 170$  B.P. (Y-843); base of unit contains 0.5 m of coarse to very coarse sand, granule gravel and small pebbles.

—Disconformity—

4.5 Silt and clay, red, rhythmically laminated in alternating silt and clay layers, probable varve couplets; leaves of tundra plants reported in upper part of clay. This is the Quinnipiac clay of this report.

In the abandoned pit immediately to the north, Antevs (1928) measured 179 varve couplets in the top 13 ft (4 m) of the clay (varves average 2.21 cm thick). Below a thin disturbed zone he measured an additional 33 varves in 2.66 ft (0.81 m) (varves average 2.46 cm thick). Nearby subsurface data indicate that the total thickness of fine-grained lacustrine sediment in the vicinity of the pit is 26-41 m (85-135 ft). The surface of the Quinnipiac clay in this pit is about 5-8 m lower than its surface beneath stream terrace deposits on the east and west sides of the valley. As shown on the profile (fig. 8a), erosion of the clay surface to this depth occurs only in the broad channel beneath the floodplain alluvium and saltmarsh deposits (fig. 2, units a and sm). This deep erosion is related to a lowering of base level in the lower valley after deposition of the stream terrace deposits. A previously described section (Hartshorn and Schafer, 1965) included a discontinuous unit of yellow and red sand and pebble gravel, suggested to be part of the Quinnipiac valley terrace outwash deposit. Correlation of this unit with the valley terrace outwash is questionable. As seen on the profile (fig. 8a), the base of the Quinnipiac valley outwash projects some 20 ft (6 m) above the -15 ft altitude of the coarse sand and pebble gravel at the base of the alluvium in the pit. The discontinuous unit of sand and pebble gravel is probably part of the alluvial deposit.

**STOP 5.** Bruce's Ice Pond cut at North Haven Industrial Park, town of North Haven, Branford quadrangle. Entrance to cut is north off Sackett Point Road on small dirt road just east of Interstate 91. Cut is 500 ft (152.4 m) southeast of Bruce's Ice Pond.

The cut exposes 3-5 m of Quinnipiac valley terrace deposits disconformably overlying 1.4 m of Quinnipiac clay. Laminated silt and clay at the base of the cut contains couplets 4-16 cm thick. Clay layers are 0.6-1.5 cm thick, and some display graded texture from darker, coarser-grained zone at the base to lighter, finer zone at top. Silt layers are planar microlaminated, and locally contain graded beds which fill shallow erosional scours. Massive silt, reddish brown (5YR4/3) and 0.9 m thick, overlies the clay. Laminated fine sand at the base of the silt is deformed by fluidization/load structures. The overlying coarse sand of the Quinnipiac valley terrace is light gray (2.5Y7/2) in dried areas and brown (7.5YR5/4) where moist. It is composed chiefly of subrounded clear quartz grains, some of which are stained with red or orange iron oxide. Other coarse sand grain constituents include red sandstone, black diabase, and mica-schist rock fragments, and flakes of muscovite. Cross-beds are in 0.3-0.5-cm-thick sets which indicate southerly current flow directions. The upper two meters of the terrace deposit are oxidized to yellowish red (5YR4/6). The surface of the Quinnipiac valley terrace is above 30 ft altitude in the vicinity of the pit; the terrace extends only 0.5 mi south of the pit to where it has been truncated by subsequent alluvial entrenchment.

**STOP 6.** Small cut at south end of abandoned Stiles Corporation claypit at North Haven; town of North Haven, Wallingford quadrangle. Turn west from Elm Street on Stoddard Ave., North Haven; access to cut is south off Stoddard Ave., past Plasticrete Corporation to south end of water-filled old claypit.

The cut exposes less than a meter of postglacial stream-terrace sediment disconformably overlying about 2 meters of Quinnipiac valley meltwater terrace sediments. This section is:

<u>Thickness (m)</u>	<u>Unit Description</u>
.5	Soil, red (2.5YR4/8-4/6), high chroma
.5	Medium-coarse sand and pebble gravel, light brown (7.5YR6/4) to light yellowish brown (10YR6/4), in trough cross-beds 10-20 cm thick; also thin beds and lenses of pebble gravel containing indistinct horizontal bedding. Stream terrace deposit.
	—Disconformity—
2	Medium-coarse sand and few pebble gravel beds, light gray (2.5YR7/2), in trough and festoon cross-beds 20-40 cm thick; laterally interbedded with coarse sand, granule and small pebble gravel in a 0.9 m cross-bed set, reddish in color with high Mesozoic clast content. Valley terrace outwash, probably low in stratigraphic position where the Mesozoic clast content is higher than near the top.

Lake-bottom sediment, 12-25 m thick in this vicinity, lies below the valley terrace outwash. At the former clay pit near this locale, Antevs (1928, loc. 128, p. 185-186) counted 56 varves in 14.5 ft (4.4 m) (average varve thickness 7.89 cm). The lake-bottom sediment here was laid down in close proximity to deltaic deposits and this is reflected in the thickness of silt layers in the varves.

**STOP 7.** Pit in Quinnipiac valley terrace deposit on the property of American Cyanamid Company near the village of Quinnipiac, town of Wallingford, Wallingford quadrangle. Pit is located 250 ft (76 m) north of Toelles Road and 1500 ft (460 m) southwest of the corner of American Cyanamid building. Access is through American Cyanamid, west off Rt. 5.

The fresh cut shows a good section of the Quinnipiac valley meltwater terrace deposit. The pit is cut into the 55-ft upper surface of the terrace. Within the exposed 4-5 m (13-16 ft) of the deposit, a change in provenance of the valley terrace sediment is notable. Sand derived from crystalline rocks is dominant at the top of the section, where the color is the typical yellowish gray and where pebbles are mostly white quartz. In the middle of the section the color is light gray but more abundant red sandstone rock fragments impart a pinkish-gray cast. The color at the bottom of the section is much redder, and red sandstone pebbles and pink sand predominate.

The deposit is overall a pebbly sand; cross-bed sets are as thick as .8 m; some sets of large cross-beds have flat basal contacts, indicating their origin as channel bars of positive relief. Pebble gravel occurs in 0.2-0.4 m-thick cross-bed sets. The thickest of these are in tabular sets in channel-bar deposits, 6 m wide obliquely and more than 3 m wide in cross-section view. Paleocurrent directions are southerly. The vertical succession of sedimentary features appears to be conformable through the exposed section; there is no extensive disconformity between the red and yellow-gray material.

**STOP 8.** Pit behind Nutmeg Farms convenience store, 500 ft (152 m) northeast of Oakdale Summer Theater, town of Wallingford, Wallingford quadrangle. Cut is on west side of Turnpike Road South and immediately south of Cook Hill Road intersection.

The pit exposes yellowish brown sand and pebble gravel of the Quinnipiac valley terrace outwash deposit which unconformably overlies collapsed reddish brown sand and gravel, which are ice-marginal deposits of glacial Lake Quinnipiac.

The upper unit consists of about 2 m of pebbly sand, medium to coarse sand, and pebble gravel beds; cut-and-fill structure is present as well as tabular cross-beds as thick as 0.8 m that fill channels several meters wide. At the base of the yellowish crystalline-derived material there is an erosional unconformity. In places on the west side of the cut a 20-25 cm-thick pebble gravel bed lies along the unconformity and is clearly part of the upper unit. Porter (1960) described the pebble gravel layer as a common unit between the base of distal outwash and the ice-contact sediments and called it a "lag gravel".

The underlying reddish brown sand and gravel is collapsed, but probable foreset bedding including avalanche foreset beds can be seen. Rhythmically bedded fine-medium sand, coarse sand with granules, and clay beds are common.

A section similar to this one about a mile south of here was described for the 1965 INQUA fieldtrip (Hartshorn and Schafer, 1965). That section included a zone of joint blocks of basalt, 5-45 cm in diameter, resting on the unconformity. These blocks of basalt moved downslope from an outcrop 20 m to the west, and showed effects of wind abrasion. Their presence indicates that the surface of the ice-marginal deposits was subjected to a period of colluviation and wind abrasion before deposition of the distal outwash deposit.

**STOP 9.** Pit on west side of Meriden-Markham Municipal Airport, town of Wallingford, Meriden quadrangle. Access is by dirt road that turns west off Hanover Street near power line, 1.3 mi (2 km) south of junction (as Evansville Avenue) with Main Street (Connecticut Route 70) in South Meriden. Pit is just west of the landing strip and north of a small stream.

The pit exposes two glaciofluvial units. The lower unit is non-collapsed, reddish-brown pebble-cobble gravel, mostly in horizontal planar beds 10-20 cm thick. Stones are imbricated; crossbeds occur in 20-cm sets. The upper unit is outwash of the Quinnipiac valley meltwater terrace deposit and consists of 1.5 m of yellowish-gray, pebbly coarse sand containing tangential crossbeds in 10-20 cm sets. Stone counts in these units shows yellow sand - 64% crystalline clasts, 34% Triassic-Jurassic sedimentary clasts, and 2% Triassic-Jurassic basalt clasts; red-brown gravel - 14% crystalline clasts, 48% Triassic-Jurassic sedimentary clasts, and 38% Triassic-Jurassic basalt clasts.

The ice margin probably stood in the vicinity of the Hanging Hills in Meriden when the lower gravel was deposited (fig. 2). Gravel deposits in Sodom Brook valley coarsen northward and are cobble and boulder gravel near Beaver Pond. Base level for this unit was probably glacial Lake Quinnipiac to the south, and deltaic parts may occur in this area. Lake Quinnipiac varved silt and clay underlie these fluvial units from here southward.

**STOP 10.** Pit in ice-channel filling at the west end of the Quinnipiac Gorge, Cheshire, Meriden quadrangle. Entrance to pit is on south side of Cheshire Road (State Route 70) north of Broad Brook Reservoir.

This gravel pit is cut into the "feeder esker" of an ice-marginal delta, built into a small lake impounded in the north-draining Broad Brook valley (fig. 2). Hanshaw (1962) showed the esker as map unit Q1c and the delta as unit Q1d. Figure 9 shows part of this map. This large unit of related ice-channel fillings extends north of the pit and adjacent Quinnipiac River to hillslopes north of the Quinnipiac Gorge. The highest parts of the unit attain 230 ft altitude. Because of its position, altitude, and coarse grain size, this deposit of ice-channel fillings and related blocks of stagnant ice may have contributed to the blockage at the Quinnipiac Gorge that impounded glacial Lake Southington to the north (La Sala, 1961).

The pit is cut into the north end of an ice-channel filling or esker that is at 190 ft at the top of the pit. The crest of the ridge rises to 240 ft altitude at the proximal part of an ice-marginal delta to the south (fig. 9). The pit exposes a large face that is parallel to the crest of the ridge, and lesser oblique faces. The gravel is composed dominantly of red Mesozoic sandstone; clasts are well rounded, not striated and some are nearly spherical. Overall, bedding has not been deformed by collapse, possibly indicative of valley-bottom, ice-tunnel deposition. Beds in the central core of the ridge are cobble gravel with abundant sandy matrix, indistinctly bedded, and containing outsized small boulders throughout; sand beds are a minor constituent. Sediments in lateral positions in the ridge include indistinctly bedded pebble and cobble beds, 0.1-0.3 m thick, and cross-beds in 0.2-0.5 m sets. Depositional processes have sorted this sediment into coarse sand to medium cobble gravel. No boulders are included; large boulders on the sides of the ridge are probably derived from ice or upper-core ridge slopes.

**STOP 11.** "Lougee delta" of Lake Southington, town of Cheshire, Southington quadrangle. Small exposures in the top of the delta and in upper foreset strata are open from time to time in areas off East Johnson Ave. (fig. 10). Deep cuts in the ice-contact head of the delta are now graded along State Route 66 (hachured lines); an old pit in the ice channel can be reached via the small residential street east of the channel filling.

The original morphology of the "Lougee delta" is shown by Lougee's (1938) topographic map and modern maps of adjacent areas (fig. 10). Small pits in the distributary lobes south and north of East Johnson Ave. show steeply inclined coarse sandy delta-foreset strata, with subordinate beds of climbing-ripple cross-laminations. Shallow surface excavations in the fluvial plain of the delta yield enough stones to measure for clast-size study. Some morphologic details have been preserved in the ice contact next to the Route 66 corridor, and some large cobbles and small boulders indicate the size of the coarsest stones in the proglacial streams next to the ice contact. Coarse cobble-boulder gravel can still be seen in the old pit in the ice-channel filling. A sand pit in the 200-205-ft flat-topped hill 200 m west of the bend in the ice-channel filling shows gravel topset beds overlying sandy delta-foreset beds.

**STOP 12.** Pit in 200-ft altitude hill just east of "Lougee delta", County Wide Construction Co. pit, town of Cheshire, Southington and Meriden quadrangles. Entrance to pit is north from East Johnson Ave. 0.3 mi west of junction with Cheshire Street.

On the 1982 NEIGC fieldtrip (Stone and others, 1982, Trip Q1, Stop 8) this pit showed an excellent oblique view of foreset beds dipping southwesterly. At that time, the face exposed gravel-dominated foresets at the north end, interbedded sand and gravel strata in the middle part of the face, and sandy foresets at the distal southern end of the pit. The exposure in 1985 is about 20-30 m west of the 1982 exposure, providing us with a chance to compare some of the three-dimensional details of the beds. Also, the present exposure extends to the ice contact at the north end, where no backset beds are seen. In recent visits to the pit, no flowtills or other matrix-supported diamict sediments were seen.

This deltaic deposit is probably the same age as the "Lougee delta"; the internal features of the "Lougee delta" may be similar to those exposed in this pit. The small 200-ft hill 100 m north of the ice contact of the County Wide pit deposit has recently been removed. Exposed in 1983, it was seen to be composed of sandy, ripple-dominated delta foresets and bottomsets, notably finer-grained than the section in the northern part of the County Wide pit. This small deposit to the north is therefore younger than the "Lougee delta" and the County Wide pit deposit.

**STOP 13.** Proglacial delta of glacial Lake Southington, Mulberry Street near Plantsville, town of Southington, Southington and Meriden quadrangles. Entrance to pit is north off Mulberry Street, 0.5 mi (0.8 km) west of intersection with Old Turnpike Road.

The pit exposes the upper part of a distributary lobe of another ice-marginal delta of Lake Southington. R. J. Lougee (1938) obtained a topset-foreset contact altitude of 196 ft in another delta 1 km to the west. The delta is one of two that were included by La Sala (1961) in his map unit Qkd<sub>10</sub>. The delta extends from the 200-ft surface at its lobate margin at the pit to its feeding ice-channel filling to the north, which attains an altitude of just over 220'. Exposures in the ice-channel filling at its point of attachment to the delta reveal delta-foreset beds of gravel and sand. Thus, the continuity of the delta-topset plain with the ice-channel filling and the internal structures indicate that delta-foreset and topset beds prograded outward from a till- or bedrock-floored channel between blocks of stagnant ice.

At the north end of the pit 1.5 m of pebbly sand topset beds are chiefly trough-filling cross-beds, exposed in cross- and longitudinal section. The erosional base of the topset sequence is remarkably straight and horizontal in the pit walls. Proximal upper foresets in the north end of the pit consist chiefly of avalanche sets of pebbly sand and fine gravel beds that dip as much as 32°. Dip direction is southerly though laterally variable. Distal foreset strata at the south end of the pit include silty beds over sandy beds. Sedimentary structures are planar beds, solitary and climbing-ripple cross-laminations, local regressive ripple laminae, and draped laminae. Clayey silt laminae in the upper 1 m of the section are deformed in fluidization/load structures, possibly related to near-surface dewatering of the loose sediments on the front of the distributary lobe.

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## MORPHOLOGY OF COASTAL MARSHES, SOUTHERN CONNECTICUT

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(Pages C-7-1 to C-7-6 of this field trip guide are reprinted from Bloom, A.L. and Ellis, C.W., Jr., 1965, Postglacial Stratigraphy and Morphology of Coastal Connecticut: State Geological and Natural History Survey of Connecticut Guidebook No. 1).

## PALUDAL STRATIGRAPHY AND MORPHOLOGY

*Introduction*

There were before human intervention an estimated 43 sq. mi. of tidal marsh along the 98-mi. straight-line length of the Connecticut coast. In the last decades B.P. (before physics) some work had been done on the paludal stratigraphy and morphology, but no regional study had been attempted. Brown (1930) described a section in a now-flooded clay pit near North Haven and discussed its significance. Knight (1934) described a small marsh in Branford that "revealed a section preserving the hitherto unrecorded early stages of a New England salt marsh developed in accordance with Shaler's classic theory, coupled with later stages developed in accordance with the theory first proposed by Mudge and later repropounded and elaborated by Davis."

In 1960, encouraged by preliminary work and the reports of Brown and Knight, a systematic study of the Connecticut coastal marshes was begun by Bloom. The initial goal was to collect sufficient samples for radiocarbon dating so that the age and rate of postglacial submergence could be determined. Field work was supported in part during 1960 by the Connecticut Geological and Natural History Survey, and since 1960 by the Office of Naval Research, Project NR 388-065. Since the initial goal was achieved (Bloom and Stuiver, 1963) the project has been modified to include research on sedimentation rates and shoreline erosion of the coastal marshes.

Consideration of the relationship between sedimentation and submergence pervades the interpretation of the Connecticut coastal marsh environment. Three significant paludal environments can be distinguished, wherein the interaction of the two variables has produced three distinct stratigraphic records.

(a) *The estuarine "fresh-water" marsh.* Where a sufficiently large river enters an estuarine marsh, the fall and rise of the tide causes alternate accelerated stream flow and slack water. The salinity is low, but the nutrient content of the water is apparently high. A dense growth of *Typha* (cattail), *Phragmites* (reed), and *Scirpus* (bulrush), commonly more than 6 ft tall, characterizes this environment. Harshberger (in Nichols, 1920, p. 540) likened these marshes to the British "fens." Production and accumulation

of organic debris has kept pace with submergence, and a thick layer of sedge peat has been built up to the local high-tide level in the marsh.

(b) *The former deep (9-50 ft) bay or lagoon.* During rapid submergence, until about 3,000 years ago, the sea transgressed into coastal valleys and produced bays or lagoons. In an environment of generally low wave energy and low sediment supply, early submergence exceeded the rate of sedimentation, and open water of near-normal salinity persisted in the embayments. However, during the last 3,000 years submergence has been slow enough to be equaled by the sedimentation rate, and salt marshes have filled former bays and lagoons. A typical stratigraphic section of these salt marshes is composed of a veneer of muddy salt-marsh peat, 9 ft or less in thickness, overlying a thick wedge of mud that has an open-bay fauna. Below the mud in many marshes there is a thin layer of sedge peat in sharp contact with the substratum. This peat represents the fringe of "fresh-water" rushes and reeds that grew at the transgressing shoreline.

(c) *The shallow (less than 9 ft) coastal marsh.* A coastal embayment less than 9 ft deep below present high-tide level was not affected by submergence prior to 3,000 years ago. Many of the shallow marshes are on submerged coastal lowlands, especially outwash plains, which continue their gentle seaward slope up to a mile beyond the high-water line. Many of these low-relief areas were marshy even before submergence raised the water table. The vegetation on these marshes is zoned landward from salt marsh through belts of progressively lower salinity tolerance to either normal upland vegetation or fresh-water marsh. The stratigraphy of a shallow marsh is similar to that of the upper 9 ft of a salt marsh in a former deep bay, except that lenses and tongues of sedge peat complexly alternate with salt-marsh peat. The alternations reflect shifts of vegetation belts across the marsh as seasons of abnormal high tides or excessive fresh-water runoff displaced the zone of salt marsh respectively landward or seaward. Deeper parts of the marsh apparently represent former topographic basins that filled with sedge peat as a result of the rising ground-water table prior to marine inundation. A normal upland soil profile on glacial drift commonly underlies a shallow coastal marsh.

Three marshes (fig. 1) have been chosen to represent the three paludal environments outlined.

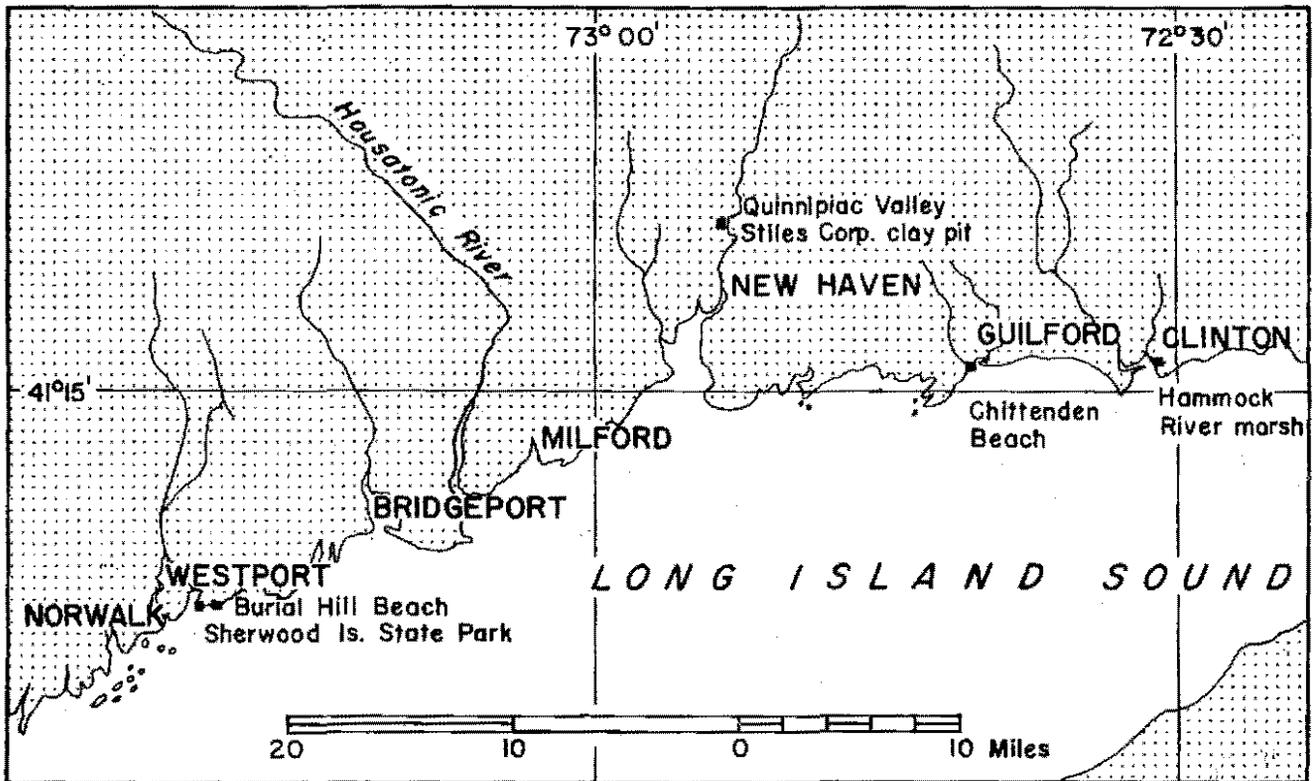


Fig. 1. Index map of the central Connecticut coast

### Quinnipiac Valley, Hamden — an estuarine "fresh-water" marsh

Excellent exposures of late-glacial and postglacial deposits have been available for many years in the brickyard clay pits of the Quinnipiac Valley, near New Haven. The section described by Brown (1930, p. 263-266) was obtained from a now-flooded pit north of the Stiles Corporation brickyard. However, a similar section (fig. 2) is currently exposed in the pit south of the brickyard.

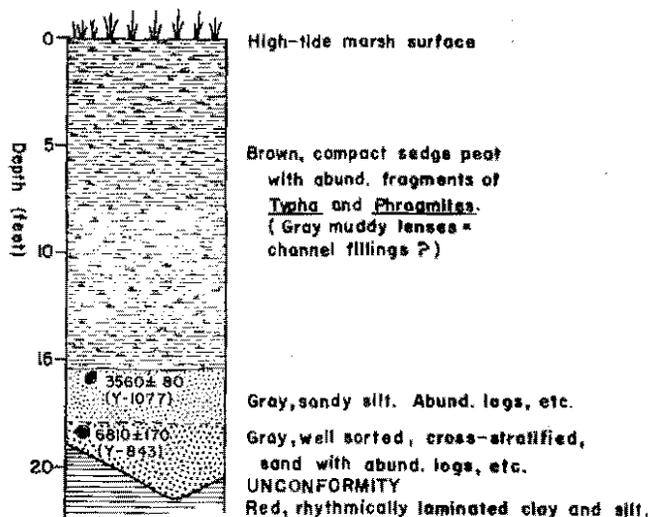


Fig. 2. Generalized stratigraphy of the Stiles Corporation clay pit, Hamden, Connecticut

The postglacial stratigraphy of the Quinnipiac Valley begins at the erosional unconformity between underlying glacial deposits and overlying alluvium. An episode of stream erosion to a lower-than-present base level followed deposition of the late-glacial lacustrine New Haven Clay. Erosion was followed or accompanied by fluvial deposition of cross-stratified sand and gravel. In other parts of the Quinnipiac Valley yellowish-gray outwash unconformably overlies the channeled upper surface of the New Haven Clay (Porter, 1960, p. 18). The alluvium in the Stiles clay pit is not outwash, as is evidenced by (1) the arkosic composition of the alluvium, (2) the dominantly hardwood composition of the enclosed logs and leaf mats, and (3) a radiocarbon age of  $6810 \pm 170$  years B.P. for a log from the alluvium (Y-843). Postglacial erosion in the southern end of the Quinnipiac Valley apparently not only removed the outwash that formed the final glacial deposit of the valley, but cut into the underlying reddish, arkose-derived ice-contact stratified drift and lacustrine clay and silt. This unconformity represents a hiatus of approximately 6,000-7,000 years prior to 6800 B.P.

The basal alluvium of the postglacial section exposed in the Stiles Corporation clay pit grades upward into gray sandy silt of questionable origin. The silt represents the loss of former stream transporting power. It is a slack-water deposit but whether it is a fresh-, brackish-, or salt-water deposit has not been determined. Foraminifera or sponge spicules are not present. The silt contains abundant logs, twigs, nuts, and leaf trash of species similar to those preserved in the underlying alluvium. A log from the top of the silt was radiocarbon dated at  $3560 \pm 80$  years B.P. (Y-1077).

Brown sedge peat, 12 to 17 ft thick and similar to that which is presently accumulating on the marsh surface, immediately overlies the silt. The nature of the contact indicates an abrupt change in the depositional environment from the time of silt accumulation to the time of peat accumulation, although no erosional unconformity is apparent. Old reports (Davis, 1913, p. 700; Brown, 1930, p. 265) described a "forest soil" and tree stumps rooted in place beneath the peat of the Quinnipiac Valley, but no recent observers have verified these reports. At the Stiles Corporation clay pit, the transition from silt deposition to peat accumulation represents only a change in depositional environment without an interval of weathering and soil formation. This change took place shortly after 3560 B.P. Since then, the Quinnipiac Valley has had its present appearance, with a cattail, sedge, and reed marsh growing to high-tide level in an estuarine environment of low salinity. Salt-marsh grasses do not now enter the valley in significant quantity north of the railroad yards, 2 mi. south of the clay pit.

Depth measurements on the pit face are subject to error because of compaction of the clay-pit wall by an overlying earth dike. Figure 3 shows a section through the south face of the Stiles Corporation clay pit on June 16, 1962, shortly after the earth dike had been moved back for a new cut in the pit. A shallow sag pond parallel to the outer edge of the dike and tension cracks on the inner slope indicated that compaction was in progress. The peat at boring 1 had been compressed from an original thickness of 15.7 ft to 13.0 ft, or to about 83 percent of its original thickness. That much compaction was accomplished by earth fill about 12 ft deep on the boring site for an estimated 2 months. At the pit face, where the dike is believed to have lain through the preceding winter, the compaction was to about 63 percent of original peat thickness. To further complicate depth

measurements, some "heave" or relaxation at the site of boring 1 seemed to have resulted from the removal of the dike. Vertical faults in the New Haven Clay, parallel to the pit face and upthrown on the pit side, suggested that both compression under the load of the dike and subsequent relaxation also may take place in the underlying silt-clay rhythmites. Thus, the depths of radiocarbon-dated samples from the clay pit are not considered as reliable as those of samples collected by coring undisturbed marshes.

### Hammock River Marsh, Clinton — a former deep bay or lagoon

The Hammock River marsh in Clinton (fig. 4) has the appearance and stratigraphy typical of many Connecticut salt marshes. The surface is a thick mat of short, wiry salt-marsh grasses, especially *Spartina patens*. Along the banks of channels, the taller salt thatch, *S. alterniflora*, grows. Whereas *Spartina patens* can tolerate only a brief wetting by salt water at normal high tide, *S. alterniflora* can tolerate submergence of its roots for 5 to 16 hours daily. The combined effect of these two plants and similar species has been to build and maintain the marsh surface very close to the local mean high-water level.

(A tidal gate installed under the Hammock River bridge of Route 145 now inhibits the inflow of salt water to the northeastern part of this marsh, and reeds, shrubs, and weeds are rapidly destroying the smooth beauty of the salt meadow. The southwestern arm of the salt marsh is flooded by high tides through a drainage ditch extending through Hammock Point Beach to the southwest, and has not yet degenerated.)

Prior to submergence, the Hammock River probably flowed west on a flood plain about 38 ft below the present marsh surface. A tributary valley sloped northeast toward

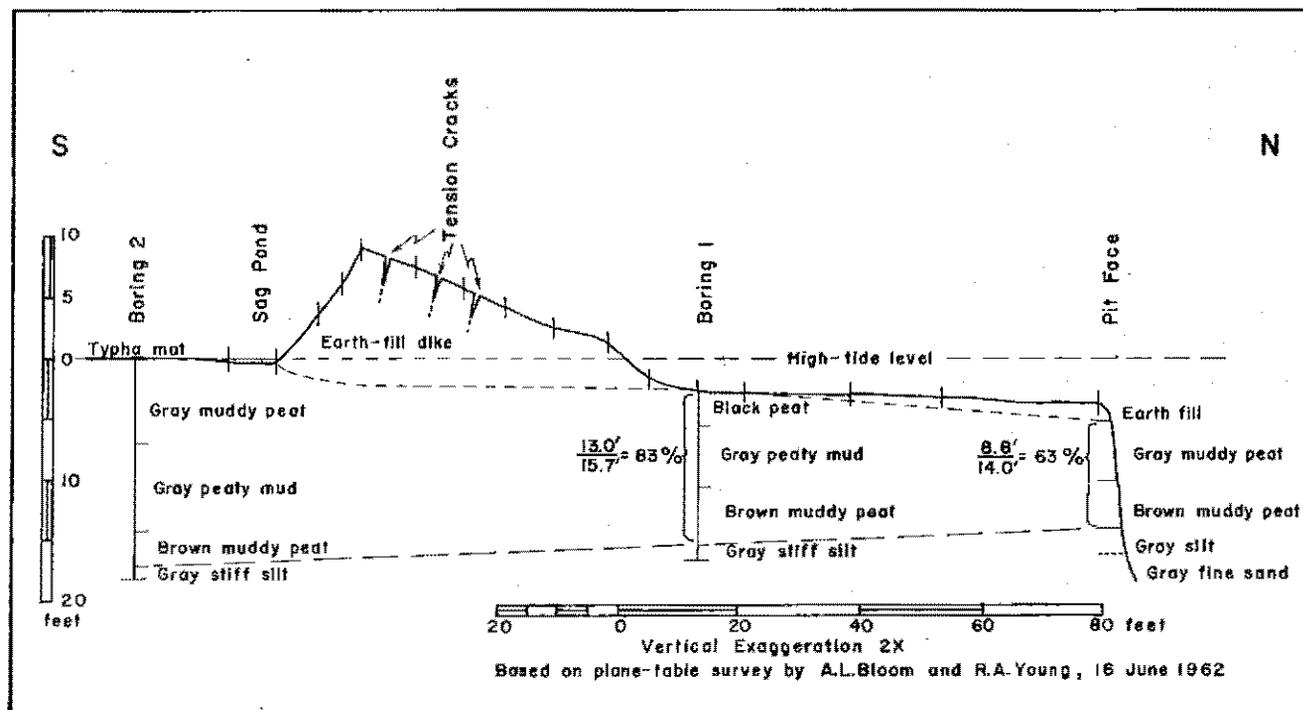


Fig. 3. Section through south face of Stiles Corporation clay pit, Hamden, Connecticut

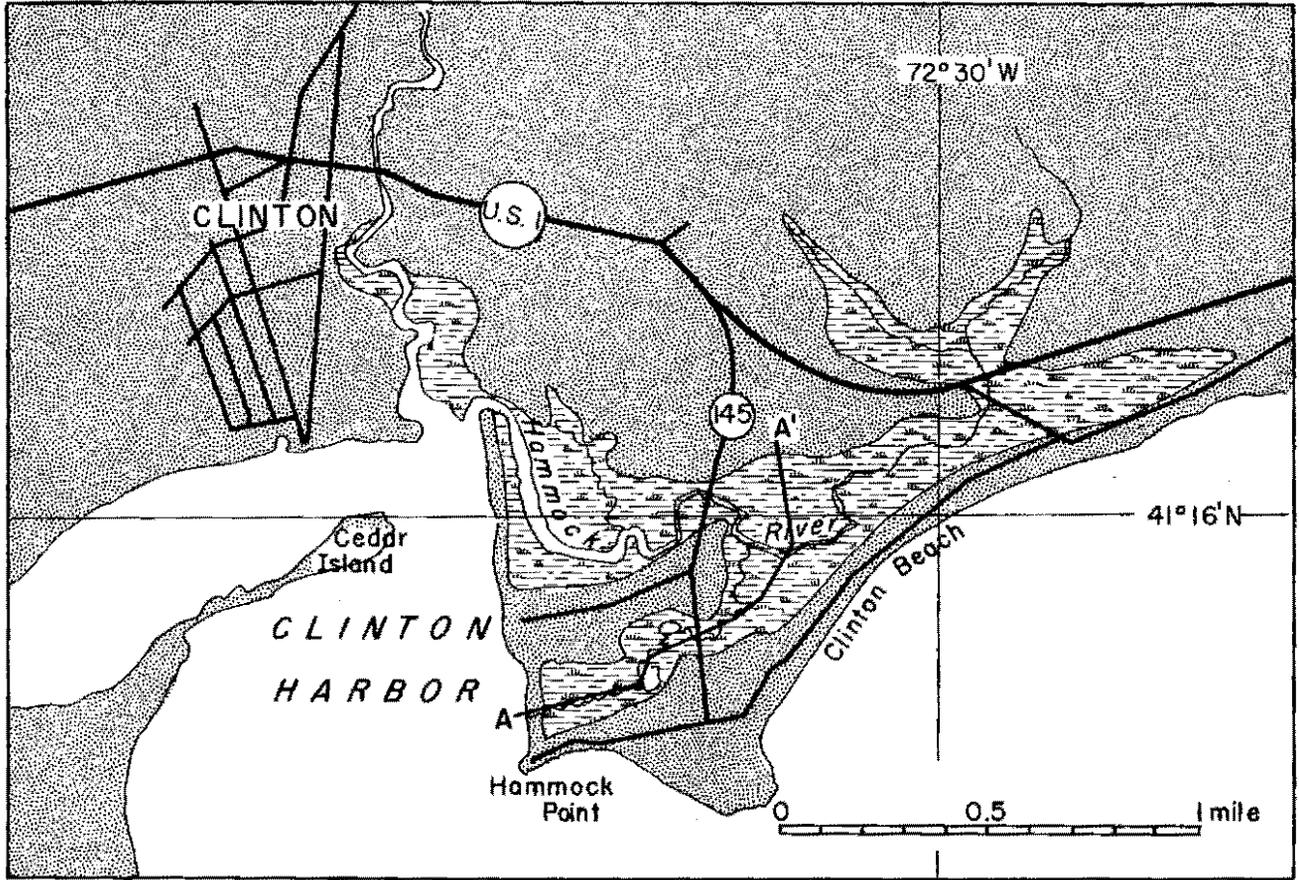


Fig. 4. Hammock River marsh and vicinity, Clinton, Connecticut

the river along line of section A-A' (fig. 4). Figure 5 shows the stratigraphy of section A-A'. The base of the section is the sand and gravel of the former valley floor, which had a northeastward gradient of about one percent.

As the sea transgressed eastward into the Hammock River valley, then southwestward into the tributary valley along the line of section, the shoreline was fringed by rushes and sedges. The basal unit of the stratigraphic section is a layer of sedge peat that accumulated at the transgressing high-tide

shoreline. The sedge peat is overlain by mud of a shallow open-bay environment. The mud contains an abundant shallow-water, muddy-bottom fauna of snails, clams, and Foraminifera. Frances L. Parker (1962, personal communication) reported the following notes on the Foraminifera of boring 15 of the section:

The upper 8 samples (8 ft.) contain a marsh fauna, either tidal marsh or shallow marsh pools. With sample 8, a rather meager bay fauna appears. I would guess that the

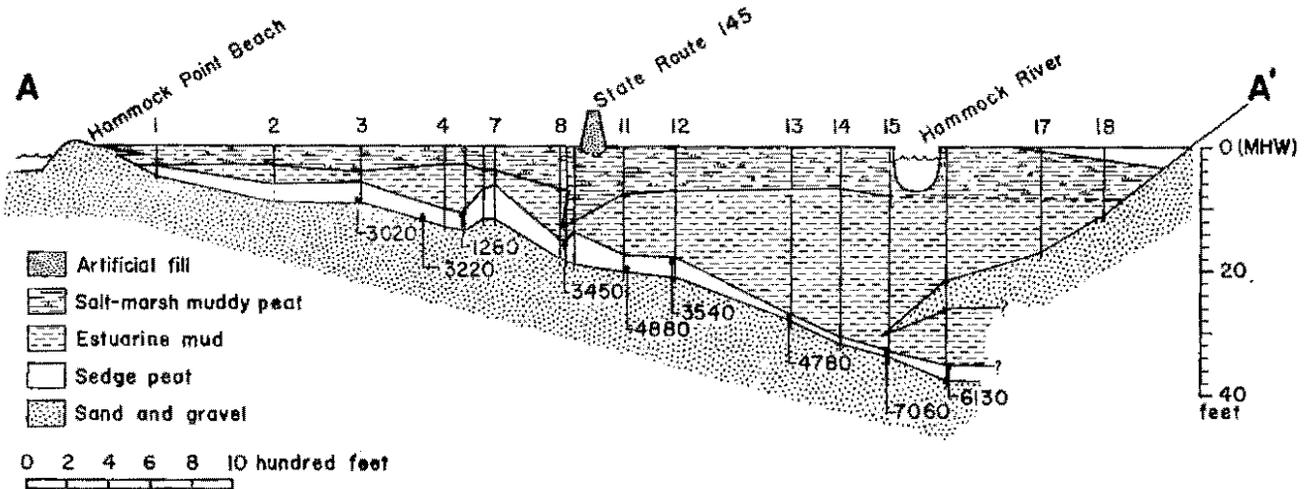


Fig. 5. Cross section of the Hammock River tidal marsh, Clinton, Connecticut

water was probably shallow and conditions not too good for the Foraminifera. In one or two samples, there was evidence of solution of the  $\text{CaCO}_3$  tests. The bay fauna is best developed at about samples 12-17. By bay, I don't mean a nice big open bay but rather some kind of semi-enclosed bay, probably with salinities somewhat lower than truly marine ones.

Sedimentation in the Hammock River estuary or lagoon (the nature of the embayment and the distribution of former barriers has not been determined) did not keep pace with submergence prior to 3,000 years ago, and open-water conditions persisted. However, when the rate of submergence decreased about 3,000 years ago, mudflats built up to the mid-tide level and were populated by *Spartina alterniflora*. The mid-tide marsh that developed was an efficient sediment trap, and in a short time the marsh surface had been built to high-tide level, where *S. patens* and related species became established. The lower third of the "salt-marsh muddy peat" of figure 5 consists of strawlike *S. alterniflora* fragments in mud, whereas the upper two-thirds consists of the fibrous roots of *S. patens* and similar high-tide species. Submergence of about 9 ft during the 3,000 years of marsh formation produced the thick section of peat derived from plants that live in a narrow vertical range near high tide (the "Mudge-Davis" type of salt marsh).

The positions and radiocarbon ages of peat samples from the Hammock River marsh are shown in figure 5. Table 1 is a list of radiocarbon-dated samples from coastal Connecticut (after Bloom and Stuiver, 1963, p. 333). The dates are plotted against sample depth in figure 6 (Bloom and Stuiver, 1963, p. 333) and a curve is drawn through the samples whose depths have not been affected by compaction. The most reliable samples used in preparing the submergence curve came from the base of the sedge peat in the Hammock River marsh, where a nearly ideal combination of permeable substratum and sloping valley floor per-

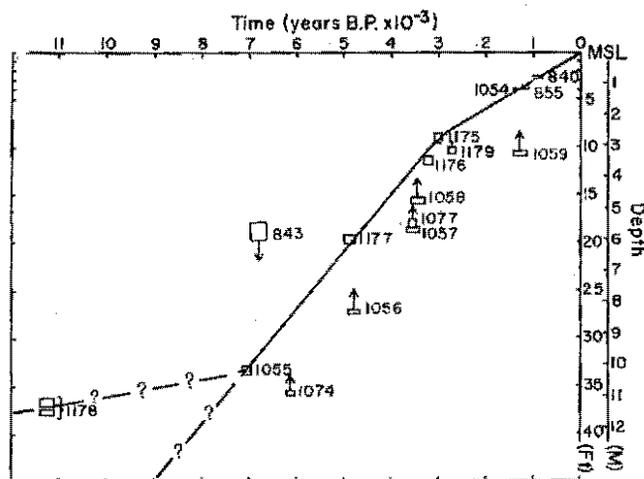


Fig. 6. Submergence of the Connecticut coast (Bloom and Stuiver, 1963). The line is the locus of a point now at mean sea level.

mitted the accumulation of sedge peat only very near the high-water shoreline of the transgressing sea. The depth of these samples below the present high-tide marsh surface in the same embayment is an accurate measure of the relative change of level since peat accumulation began. Samples collected from the top of the buried sedge-peat bed plot below the curve of submergence by an amount equal to the compaction of the peat. The displacement by compaction plus the present thickness of peat equals the original thickness, and the ratio of present to original thickness can be calculated. The sedge-peat bed beneath the Hammock River marsh has been compacted to between 13 and 44 percent of original thickness. Older and deeper samples (6,130 yrs. = 22 percent; 4,780 yrs. = 13 percent) show greater compaction.

Table 1. Radiocarbon-dated samples from coastal Connecticut<sup>1</sup>

Laboratory No.	Locality	Sample	Depth (ft)	Age (years before present)
Y-840 <sup>2</sup>	Branford	Cedar root	2.7 ± 0.2	910 ± 120
Y-843	North Haven	Log	18.5 ± 1.0	6,810 ± 170
Y-855 <sup>2</sup>	Guilford	Oak log	3.8 ± 0.2	1,180 ± 80
Y-1054 <sup>2</sup>	East Norwalk	Tree root	4.0 ± 0.2	1,400 ± 70
Y-1055 <sup>2</sup>	Clinton	Peaty sand	33.3 ± 0.4	7,060 ± 100
Y-1056	Clinton	Sedge peat	27.2 ± 0.3	4,780 ± 130
Y-1057	Clinton	Sedge peat	18.6 ± 0.3	3,540 ± 130
Y-1058	Clinton	Sedge peat	15.6 ± 0.3	3,450 ± 160
Y-1059	Clinton	Sedge peat	10.7 ± 0.3	1,280 ± 150
Y-1074	Clinton	Sedge peat	35.7 ± 0.4	6,130 ± 90
Y-1077	North Haven	Log	18.0 ± 0.5	3,560 ± 80
Y-1175 <sup>2</sup>	Clinton	Sedge peat	9.1 ± 0.6	3,020 ± 90
Y-1176 <sup>2</sup>	Clinton	Sedge peat	11.4 ± 0.5	3,220 ± 90
Y-1177 <sup>2</sup>	Clinton	Wood and bark	19.6 ± 0.5	4,880 ± 120
Y-1178 <sup>2</sup>	Clinton	Sedge peat (combined)	37.7 ± 0.3	11,240 ± 160
Y-1179	Westport	Sedge peat	10.4 ± 0.4	2,710 ± 90

<sup>1</sup> Bloom and Stuiver, 1963, p. 333

<sup>2</sup> Samples whose depth range does not require correction because of compaction

## Chittenden Beach, Guilford — a shallow coastal marsh

The small marsh at the back of Chittenden Beach formed on an outwash plain. The outwash appears to be thin, as numerous bedrock knobs protrude through it. The smooth profile offshore indicates that the outwash plain formerly extended at least a mile seaward of the present shoreline. Whether or not a barrier beach formerly protected the marsh has not been determined. The present beach is under-nourished, and is little more than a fringe of sand and shells being "bulldozed" landward over the marsh by storm waves. The marsh in back of the eastern part of the beach is now only about half as wide as it was in 1960.

It is possible that the marsh formerly extended thousands of feet seaward, and at its outer edge a barrier beach extended from headland to headland. If so, the present marsh and beach remnants represent the final stage of landward retreat of a barrier beach across a filled lagoon. However, development of this marsh may not have been the result of a protecting barrier. Former glacial deposits, now submerged or eroded, may have provided protection for the early marsh. In that case, the present beach is only a re-worked remnant of drift, rather than of a former barrier beach. As a third possibility, it may be that no more protection has ever been provided for this marsh than it now has. Erosional retreat of the marsh edge has been at the rate of 10 ft or more per year since 1960, and at least 2 to 3 ft per year during recent decades, according to local residents; however, these rates may not be typical of erosional retreat during the several thousand years of marsh history. The widespread destruction of eel-grass beds offshore in the early 1930s may have exposed this shoreline to more rapid erosion. Late spring storms of the past two years have been responsible in large part for the recent erosion, but their frequency in the past has not been investigated in this study. Furthermore, the submergence of the New York City tide gauge between 1893 and 1953 averaged 0.011 ft per year (Disney, 1955), about four times the average rate of submergence in Connecticut of 0.3 ft per century through the last 3,000 years. If submergence has accelerated in the past century, the effects would be most noticeable on exposed peat shorelines such as Chittenden Beach.

Some indications of a change in shoreline development at Chittenden Beach appeared in 1963. Formerly, the wave-cut intertidal peat cliff fronted on a barren tidal flat, but in 1963 a heavy growth of *Spartina alterniflora* covered much of the flat. If this vegetation persists, it may trap enough sediment from the river mouth and offshore to rebuild the marsh to high-tide level, leaving the present beach as a "chenier" across the marsh. Future years will determine the validity of this hypothesis.

The stratigraphy of Chittenden Beach marsh has been studied by coring, and is also exposed in the wave-cut cliff at low tide. The pollen profile of a 270 cm (9 ft) core from a site now beneath the beach was prepared by Sears (1963, p. 59). Figure 7 is reproduced from his report. The oxidized peat zone at the base of the section probably represents chemical activity by ground water from the underlying drift, but it could have paleoclimatic significance. The transition from underlying sedge peat to overlying salt-marsh peat was 95 cm (3.1 ft) below the marsh surface. The arboreal pollen content of the core shows a general shift upward from oak to pine and hemlock. Sears (1963,

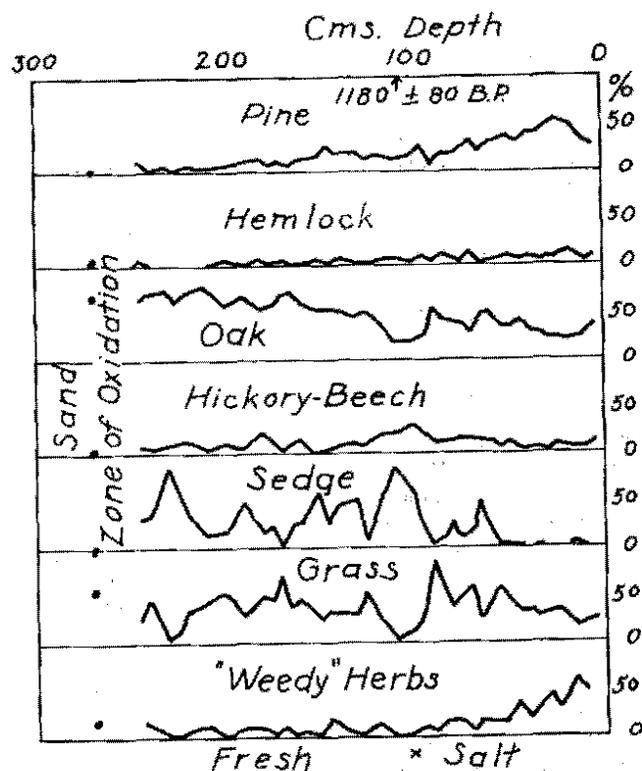


Fig. 7. Pollen diagram of important indicators in the Guilford, Connecticut, coastal marsh (Sears, 1963)

p. 59) interpreted this as a trend toward a cooler and moister climate during the time of marsh accumulation. Superimposed on the climatic change inferred from the arboreal pollen are a series of reciprocal alternations between sedge and grass pollen. Sears (1963, p. 59) interpreted these recurrent fluctuations as showing a pulsating rather than steady rise of the water table during submergence on the hypothesis that "slight rises in the water table normally favor sedges at the expense of grasses." However, in coastal marshes salt-marsh grasses normally displace sedges during growing seasons of abnormally high tides, the reverse of Sears' hypothesis. Most of the sedge-to-grass fluctuations are recorded in the sedge-peat part of the core, but some are shown in the upper salt-marsh peat as well. Because the environment has been in delicate balance with several variables, climatic interpretation is not easy.

In the wave-cut cliff of peat, sedge peat is interbedded with salt-marsh muddy peat. The best indicator of accumulation in a low-salinity marsh is the distinctive curved-triangular culm of *Scirpus maritimus*, the common coastal bulrush. Fibrous, wiry mats of roots represent growth of *Spartina patens* and related high-salinity salt-marsh plants. An oak log from the base of the wave-cut bank at Chittenden Beach was 1,180 ± 80 years old (Y-855). It came from a layer of black "fresh-water" peat 115 cm (3.8 ft) below present high tide, nearly at the contact with underlying sand.

## Shoreline Erosion at Chittenden Beach, Guilford

The shore zone of Chittenden Beach was plane-table surveyed at a scale of 1:240 (1 inch = 20 feet) in 1964, 1965, 1973, and 1983 (Bloom, 1967; Harrison, 1975; Young, 1985). The maps are too large for reproduction in the guidebook, but will be available during the excursion. In 1965, the most extensive recent erosion along the Connecticut coast was measured at Chittenden Beach, Guilford and at the Chaffinch Island shore immediately west of Chittenden Beach across the mouth of West River. From aerial photographs, the measured erosion at the center of the Chittenden Beach marsh had been 55 m between 1949 and 1965. Most of the erosion took place before the summer of 1962.

During the "Great Atlantic Storm" of March, 1962, the sand beach ridge at Chittenden Beach was thrown back onto the marsh surface 15 to 30 m behind a wave-cut peat bank in the intertidal zone. Compaction has lowered the overridden peat so that high spring tides still reach the foot of the beach ridge, but in general the beach became isolated from a sand supply and could be called a "chenier". By 1965, the small tidal marsh behind Chittenden Beach had been narrowed to half its 1960 width by beach retreat. Erosion of the marsh edge may have been accelerated when the succession of hurricanes between 1954 and 1960 pushed the beach excessively far inland and left the marsh exposed. Inadequate sediment supply has subsequently prevented a new protective beach from forming. Chittenden Beach is obviously undernourished (Bloom, 1967).

In the relatively storm-free decade prior to 1973, new growth of S. alterniflora added 4000 m<sup>2</sup> to the marsh area in the central part of Chittenden Beach, while 800 m<sup>2</sup> was eroded from the eastern and western margins (Harrison, 1975). The former wave-cut peat bank of the central beach became almost buried by new sediment trapped on the foreshore by S. alterniflora.

The 1983 survey showed several meters of progradation at the east end of the beach, where erosion had been dominant previously. By contrast, the central beach area had retreated 10 to 20 m, and a new wave-cut scarp had formed. The pre-1965 scarp between the lower and higher levels of S. alterniflora marsh was no longer continuous, due to erosion on the upper marsh or possibly vertical accretion on the lower marsh surface in the previous decade. The western end of the shore zone had retreated more than any other part, by up to 20 m (Young, 1985). The west end of Chittenden Beach had been a river-side wharf in the last century, and stone-filled log cribbing had progressively protruded into the bay along the edge of West River as the beach eroded. Since 1973 the remnants of the old structure have been largely destroyed, but the mounds of cobbles are still obvious.

After disappointing attempts in the early 1960's to convert Chittenden Beach into a recreational swimming area, the citizens of Guilford wisely dedicated the area for classes in nature study and shoreline biology. With this enlightened outlook, they can watch future changes with interest rather than concern.

## Vertical Accretion at Hammock River March, Clinton

Beginning in the summer of 1962, immediately after the major storm in March of that year, marker beds of about 1 m<sup>2</sup> were established on six selected Connecticut coastal marshes. Various materials have been used, but the most successful is the variously colored decorative "glitter" that is readily available in variety stores and stationers. Annual increments of glitter in contrasting colors were added to the marker beds from 1962 through 1966 (Bloom, 1967). The annual vertical accretion of mud on the marsh surface is recorded by measuring the average separation of successive layers of glitter. Most of the sites were relocated and refurbished in 1973 and 1974 by Harrison (1975) and again in 1983 by Young (1985). Young (1985) also resurveyed 12 similar marker beds on the shores of Long Island that had been established in 1974-1976. Her results are summarized in Table 1.

On the Hammock River marsh in Clinton (Figure 4, p. C-7-4) two marker beds were established in 1962. Both are near the bank of the Hammock River but one (Hammock River east) is upstream of a tidal gate under the bridge at State Route 145, and the other (Hammock River west) is downstream. The tidal gate is a "flapper" type, designed to open on a falling tide and drain the eastern part of the marsh, but to close on a rising tide and prevent most of the tidal flooding. Some tide water enters the eastern part of the marsh through drainage ditches that cross the marsh from the southwest, but full high-tide flooding is prevented unless the river gate is intentionally opened. As a result of the restricted tidal flooding, the former high-tide salt marsh east of the road was rapidly invaded by brackish-water plants, including Phragmites reed, upland weeds, and woody shrubs. West of the tidal gate the salt-marsh vegetation is a normal S. patens and Distichlis association.

The Hammock River west marker bed was established in 1962 on an area of marsh that probably had been burned within the previous year. No decaying grass matted the surface of the mud. Vigorous new grass sprouted from the firm, level mud surface. When the 1963 marker layer of coal was spread, the 1962 layer was covered with only 1 to 2 mm of mud, plus some decaying grass debris that was not included in the thickness measurements (Figure 7).

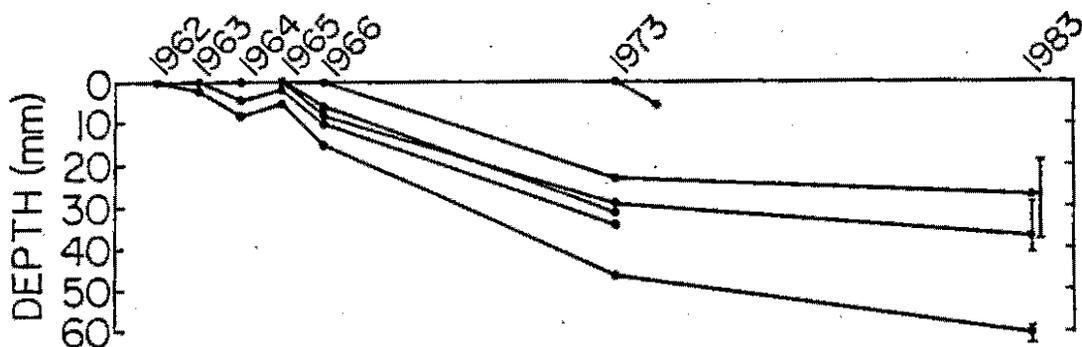


Figure 7. Vertical accretion, Hammock River west. (Young, 1985, Fig. 29).

TABLE 1

Site	Vegetation	Sed. Rate (mm/year)	Length of Record (years)
Connecticut Sites			
Barn Island	Sp	1.7	21
Great Island	Sp	4.8	11
Hammock River	Sp	2.3	21
Stony Creek	Sas	6.7	20
Nells Island	Sp	4.8	20
North Shore Long Island			
Flax Pond	Sp	1.5	2
Wading River	Sp	4.6	7
Mattituck Inlet	Sas	3.8	9
Eastern Bays			
Orient Point	Sas	0.5	2
Cedar Beach Point	Sas	1.5	2
Birch Creek	Sp	2.4	7
Squire Pond	Sp	2.5	9
Clam Island	Sp	1.8	9
Acabonack Harbor	Sp	1.9	9
Napeague Harbor			
South Shore			
Shinnecock Res.	Sp	4.5	2
Westhampton Beach	Sp	3.0	9
Carmans River	Sp	4.6	9

Sas = Spartina alterniflora, short (dwarf)

Sp = Spartina patens

The marker bed lost some sediment between 1964 and 1965. In 1965, gold glitter from 1964 was widely exposed at the mud surface beneath rotting grass, and coal of the 1963 layer, which was buried in 1964, was again exposed over part of the marked bed. Blue glitter from 1962, which had been found to a depth of 7 mm in 1964, was found only 3 mm below the surface of 1965. It must be emphasized that the measurements of Figure 7 are based on only a few turf samples cut from various parts of the marker bed, and may include some anomalies caused by microrelief at the mud surface.

In 1966, all four previous years' layers were found and measured in three turf samples. In addition to the mud accumulation recorded in Figure 7, about 25 mm of loose rotting humus covered the marker bed area. The grass had fully recovered from the pre-1962 burn. Without defining terms, the Hammock River west is subjectively regarded as a "typical" Connecticut high-tide marsh. Accumulation averaged 4 mm per year between 1962 and 1966 (Bloom, 1967). After 1966, the decade intervals of restudy preclude any significant statement about shorter term fluctuations in accretion rates. In 1983, the 1962 marker bed and two younger horizons were relocated, and the average accretion rate for 21 years was determined to be 2.3 mm/year (Young, 1985).

Hammock River east (Figure 8) is building upward at an exceptional rate. The cause is primarily the rapid accumulation of organic debris from reeds and shrubs that are invading the eastern, fresher area of the marsh upstream of the tidal gate.

Local residents reported that the tidal gate was fastened open during the 1964-65 and 1965-66 winters, and the eastern marsh was flooded and covered with a sheet of ice during much of the winter. Winter flooding and ice-rafted sediment may contribute to the exceptional sedimentation rate in the eastern part of the marsh, and yet not inhibit the summer growth of upland annual weeds and shrubs. Some pebbles and coarse sand were collected near the eastern marker bed. Ice rafting or human transport are the only plausible explanation for such coarse material in the center of a large tidal marsh.

By 1966 an extensive portion of a former creek channel in the Hammock River east marsh was completely filled and marked only by a slightly wetter meandering trace of unusually bright green grass. The eastern part of the main Hammock River channel was also filling with sediment. Steep, undercut river banks in 1962 had become gentle, muddy slopes by 1966, on which S. alterniflora was pioneering. Large areas of this marsh changed from salt-marsh grass to dense thickets of reeds and shrubs within a few years. The increasing fire hazard in this drying marsh is of great local concern. The average vertical accretion rate (Figure 8) was about 17 mm per year from 1963 to 1973 (Harrison and Bloom, 1977). In 1983 the reference stakes for this bed could not be relocated, and the vegetation, channel configuration and nearby references for compass bearings had changed so much that this interesting and anomalous marker bed was lost.

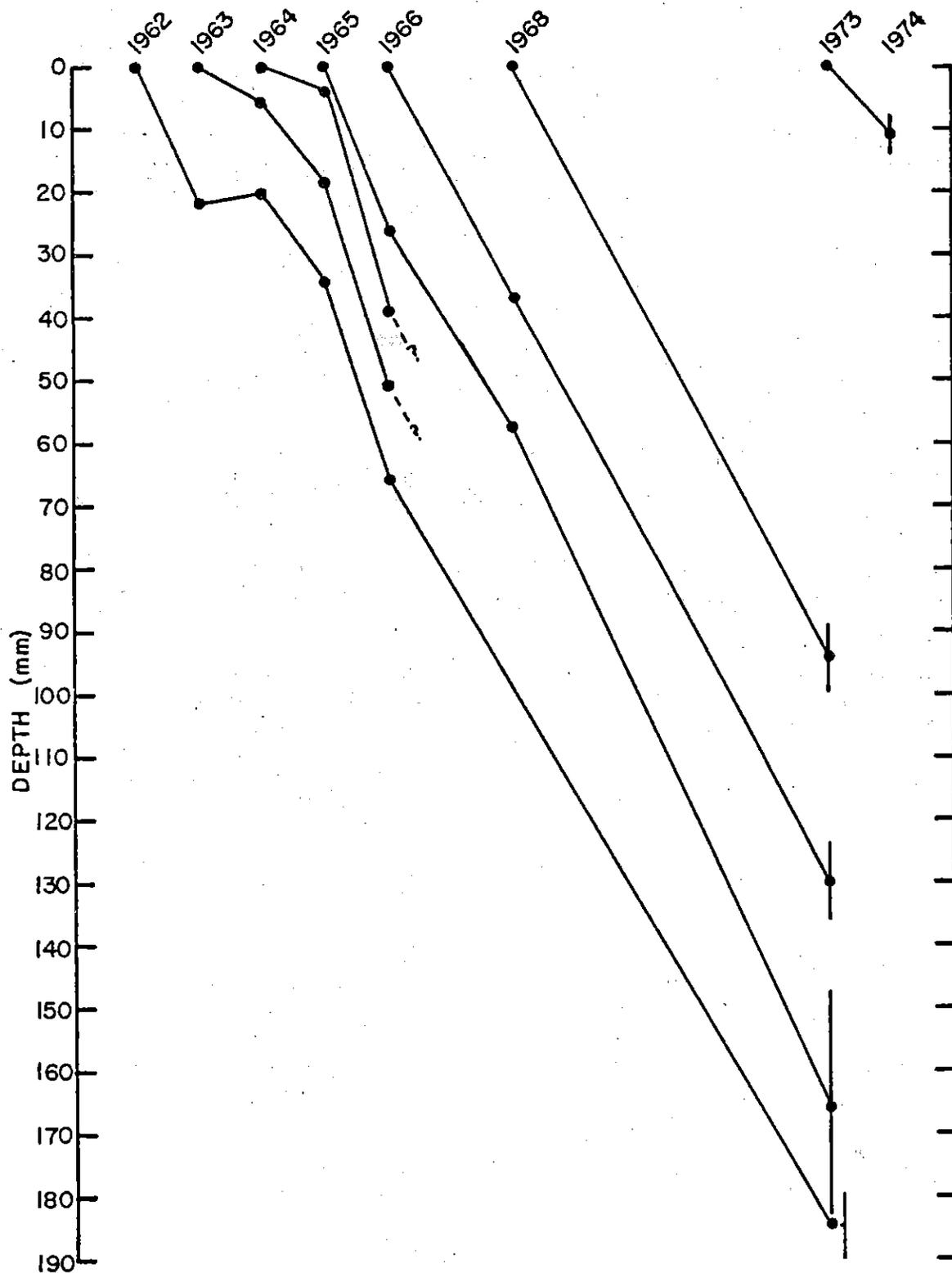
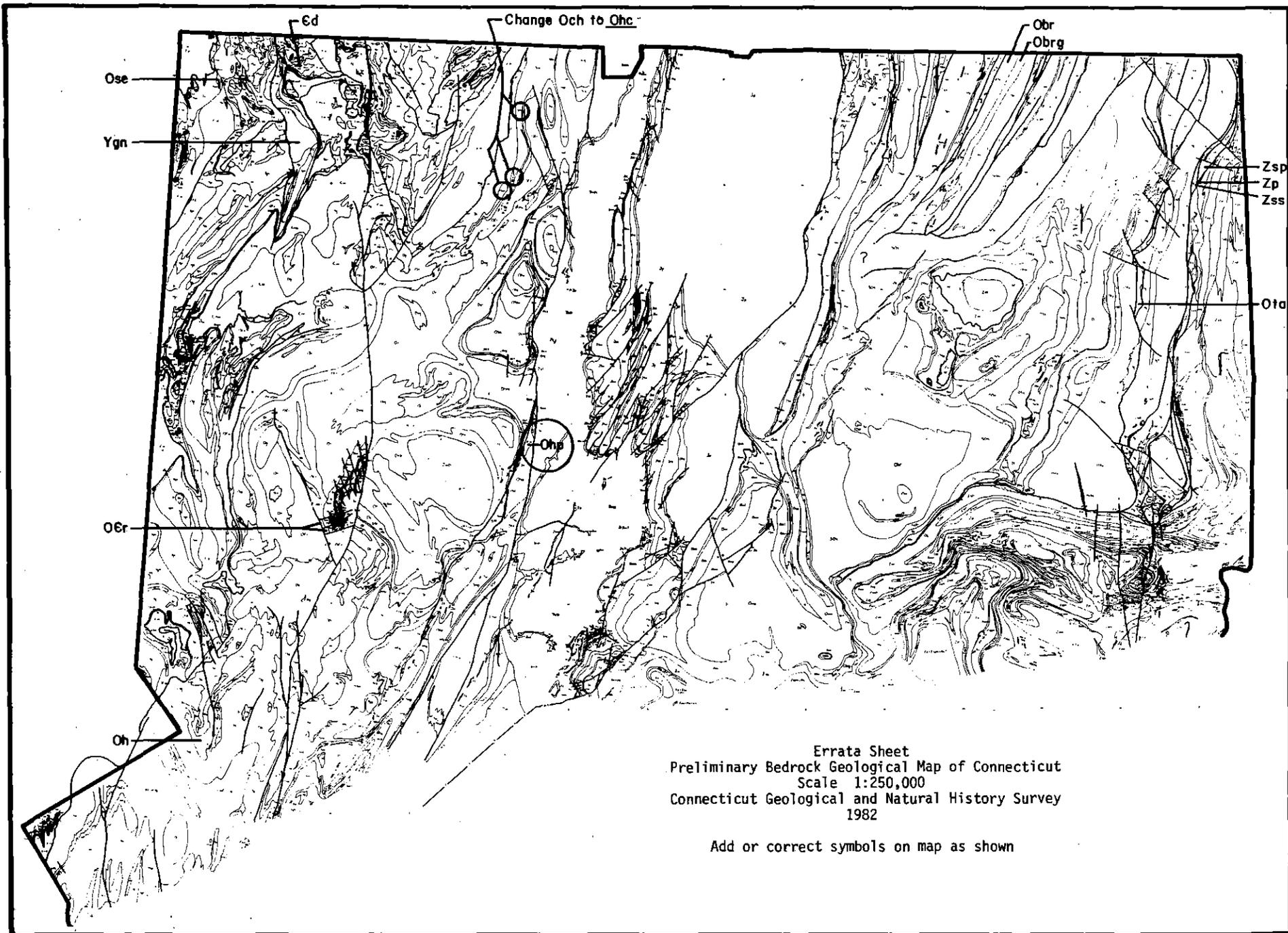


Figure 8. Hammock River east - sediment accumulation in Phragmites communis. To minimize disturbance only the 1973 horizon was recovered in 1974 (Harrison, and Bloom, 1977, Fig. 9).

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Errata Sheet  
Preliminary Bedrock Geological Map of Connecticut  
Scale 1:250,000  
Connecticut Geological and Natural History Survey  
1982

Add or correct symbols on map as shown