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*CHARGING INTO THE FUTURE*  
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**FIELD TRIP GUIDEBOOK**  
**FOR THE**  
**NORTHEASTERN UNITED STATES:**  
**1993 BOSTON GSA**  
**Volume 1**

For  
1993 Geological Society of America Annual Meeting &  
85th Annual New England Intercollegiate Geological Conference  
October 25-28  
Boston, Massachusetts



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## EDITORS' REMARKS

Our goal in assembling this field trip program and accompanying guidebook was to showcase the geology of the northeastern United States and the significant contributions of those who have worked here. The trips and authors were selected in an attempt to offer an interesting excursion for nearly every geologist while insuring access to the rich and diverse geological heritage of New England and environs. However, the papers appearing in these volumes have not been formally edited. In fact, time did not permit even a quick reading of most manuscripts. Our editorial role was limited to that of compiling these two volumes by setting the format, establishing reasonable page limits, attempting to enforce both the page limits and necessary deadlines, and insuring that the original was delivered to the printer in time to produce the books for the running of the field trips. Thus the articles appearing here are the original, uncut and unabridged offerings of their authors.

## ACKNOWLEDGEMENTS

Organizing and bringing to fruition a project of this scope is only possible with the complete and amicable cooperation of all of those trip leaders and co-leaders who planned the trips, set up the logistics, and wrote the articles. We are most grateful to all of these geologists for their friendly understanding, prompt responses to our many requests and professional conduct of this enterprise. The efforts of the GSA staff on our collective behalf are most gratefully acknowledged. We have benefitted immensely from Sue Beggs' experienced best efforts to keep us out of trouble and Kathy Lynch's knowledge. Most of all however, we are deeply appreciative of the organization, dedication and hard work of Becky Martin, the GSA Field Trip Coordinator. Jackie Newberry provided professional, prompt and cheerful assistance at Amherst. Thanks to the efforts of all of these people, the task was immeasurably simplified and indeed the much dreaded job of Field Trip Organizers was in fact most entertaining.

Howard Dimmick was instrumental in securing the flexible transportation contract with Crystal Bus lines of Boston. The University of Massachusetts Duplicating Service, directed by Leo St. Denis, printed the text. The cover, binding and the fold out map in chapter B were done by the Hamilton I. Newell Company of Amherst, Massachusetts.

Finally we would like to thank Professor Peter Robinson of the University of Massachusetts, Amherst for showing us so many times how to do this job right and for giving us the opportunity. Peter was clearly the choice to Chair the field trip committee. However, when asked by Fr. Skehan in 1988, Peter modestly declined due to age and gracefully suggested us!

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Chapter A

# Layered Gabbro-Diorite Intrusions of Coastal Maine: Basaltic Infusions Into Floored Silicic Magma Chambers

By Robert A. Wiebe and Marshall Chapman

Field Trip Guidebook for the Northeastern United States:  
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University of Massachusetts, Amherst, Massachusetts



## LAYERED GABBRO-DIORITE INTRUSIONS OF COASTAL MAINE: BASALTIC INFUSIONS INTO FLOORED SILICIC MAGMA CHAMBERS

by

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### INTRODUCTION

The layered gabbro-diorite intrusions that are the focus of this field trip belong to the Coastal Maine Magmatic Province (CMMP) which consists of more than 100 mafic and felsic plutons emplaced over a time span from the Late Silurian to the Early Carboniferous (Hogan and Sinha, 1989) (Fig. 1). The bimodal character of this province is well established (Chapman, 1962a), and there is widespread evidence for commingling between mafic and felsic magmas (Taylor et al., 1980; Stewart et al, 1988; Chapman and Rhodes, 1992). Chapman (1962a) visualized the mafic plutons as large, sheet-like masses with overlying granite beneath a country rock roof. Gravity studies (Hodge et al., 1982) indicate that many of the granitic plutons are thin with gently-dipping floors, and probably rest on mafic rocks similar to the layered diorite and gabbro that partly surround and dip beneath several of them. The plutons of the CMMP intrude a variety of metasedimentary and metavolcanic rocks in several fault-bounded, northeast-trending terranes featuring different stratigraphies and different structural and metamorphic histories (Williams and Hatcher, 1982). The ages and field relations of these plutons suggest that they postdate the main assembly of these lithotectonic terranes (Ludman, 1986). Hogan and Sinha (1989) suggested that at least some of the magmatism was related to rifting in a region of transtension along a transcurrent fault system.

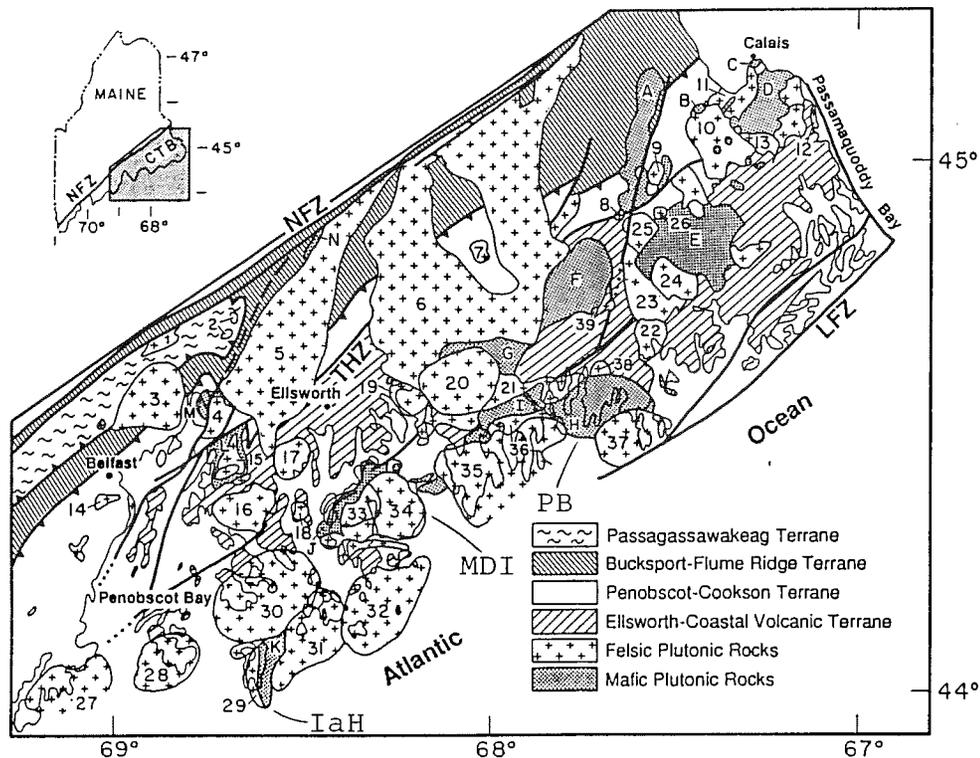


Figure 1. Geologic map of the Coastal Maine Magmatic Province within the lithotectonic terranes of southeastern Maine, taken from Hogan and Sinha (1989). Numbers and letters correspond to their designations for felsic and mafic plutons, respectively. Pleasant Bay Composite Layered Intrusion (PB) (H), Mount Desert Island (MDI) (33, 34, and J), and Isle au Haut (IaH) (29 and K) are described in this field guide.

The three complexes that we will visit (Pleasant Bay, Cadillac Mountain, and Isle au Haut) present superb examples of mafic-silicic layered intrusions (MASLI) which display complex interlayering of gabbroic to granitic rocks (Wiebe, 1993a). MASLI are characterized by gabbroic layers with chilled bases resting on silicic cumulate layers and by the development of silicic pipes and load cast structures along the bases of these gabbroic layers. The basally chilled gabbroic layers record infusions of basaltic magma into shallow-level, pre-existing floored chambers of relatively silicic magma; the pipes and load cast structures record gravitational instabilities caused by the ponding of rapidly cooling basaltic magma on incompletely solidified silicic cumulates. Layers that grade upward from basally-chilled gabbro to highly silicic cumulates provide a cumulate stratigraphic record of the double diffusive interface between stratified basaltic and silicic magmas.

Based on geologic mapping, internal structures, and gravity data these complexes appear to have dimensions, compositions, and a geologic setting that are appropriate for plutonic systems inferred to lie beneath long-lasting silicic volcanic systems in areas of bimodal magmatism - e.g. Yellowstone (Hildreth et al., 1991). Because of their stratigraphic records, and their apparent link to silicic volcanic systems, MASLI have great potential to provide new insights into magma chamber processes and into the connections between the plutonic record and volcanic activity.

## THE PLEASANT BAY LAYERED GABBRO-DIORITE INTRUSION

### Introduction

The Pleasant Bay layered gabbro-diorite intrusion, located on the coast of Maine between Bar Harbor and Machias, is roughly oval in plan, measuring 12 by 20 km (Fig. 2) (Wiebe, 1993b). It was emplaced into the Ellsworth-Coastal Volcanic Terrane (Hogan and Sinha, 1989). Gravity data (Biggi and Hodge, 1982) suggest that it is basiniform in structure with a maximum thickness of about 3 km. The basal contact of the intrusion appears everywhere to dip gently inward, and the disposition of internal layering suggests that the chamber floor consisted of

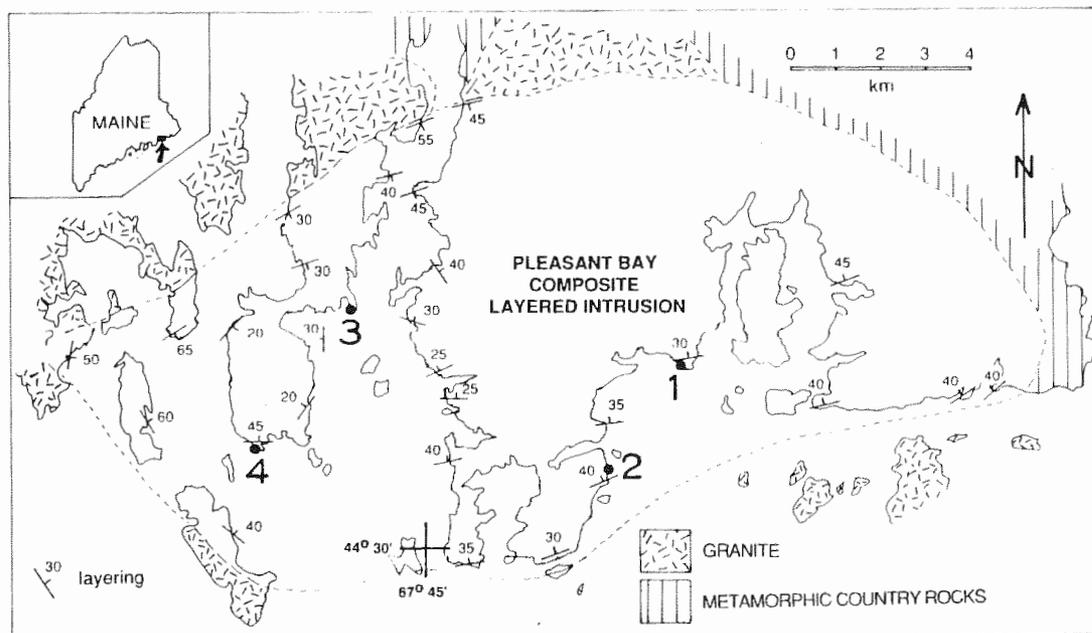


Figure 2. Geologic sketch map of the Pleasant Bay intrusion. The intrusion consists of about 90% medium-grained gabbro and mafic diorite intercalated with (1) subordinate silicic cumulates ranging from leucocratic diorite to granodiorite and (2) strongly chilled gabbroic layers. It has not been subdivided at this scale because of the rapid alternation of these rock types in layers ranging from about 1 to 100 m thick. Bold numbers (1 - 4) refer to field trip stops on Day 1.

two or more irregularly shaped basins (Fig. 2). Where the gabbroic base of the Pleasant Bay intrusion rests on granite, the gabbro is typically chilled against the underlying granite and commonly displays convex-downward lobate forms. At some places, the base of the intrusion consists of pillow-like bodies of chilled gabbro in a hybrid granitic matrix. Basaltic dikes in granite near the Pleasant Bay intrusion commonly trail off into elongate zones of chilled pillows in granite or are strongly re-intruded by the granite. These relations suggest that the granite was incompletely solidified when the gabbro was emplaced. The roof and upper parts of the intrusion have been lost

through erosion. Except for minor faulting, it appears to be undeformed. The occurrence of open vugs (up to a few cm across) in some contemporaneous granitic rocks suggests shallow emplacement. Basaltic, granitic, and composite dikes occur widely; they are mutually intrusive and typically have attitudes close to N10°W, vertical.

The intrusion consists of about 90% massive to weakly layered gabbro and mafic diorite, both of which vary widely in grain-size and texture. It also contains medium-grained leucocratic layers and lenses from a few cm to several meters thick that range in composition from diorite to granodiorite and commonly contain chilled gabbroic pillows and partly digested mafic inclusions that resemble mafic enclaves in granitic plutons (Didier and Barbarin, 1991). While these leucocratic layers commonly grade downward to gabbro, overlying gabbroic layers are typically chilled against them.

### Field Relations

**Layered Units.** Two stratigraphic sections in Fig. 3 illustrate sequences of layers and the occurrences of distinctive structures along layer boundaries that are typical of the Pleasant Bay intrusion. These sections have been subdivided into macrorhythmic units with chilled gabbroic bases that grade upward to medium-grained gabbro, diorite or highly evolved silicic cumulates. The top of each unit is truncated by the chilled base of the overlying

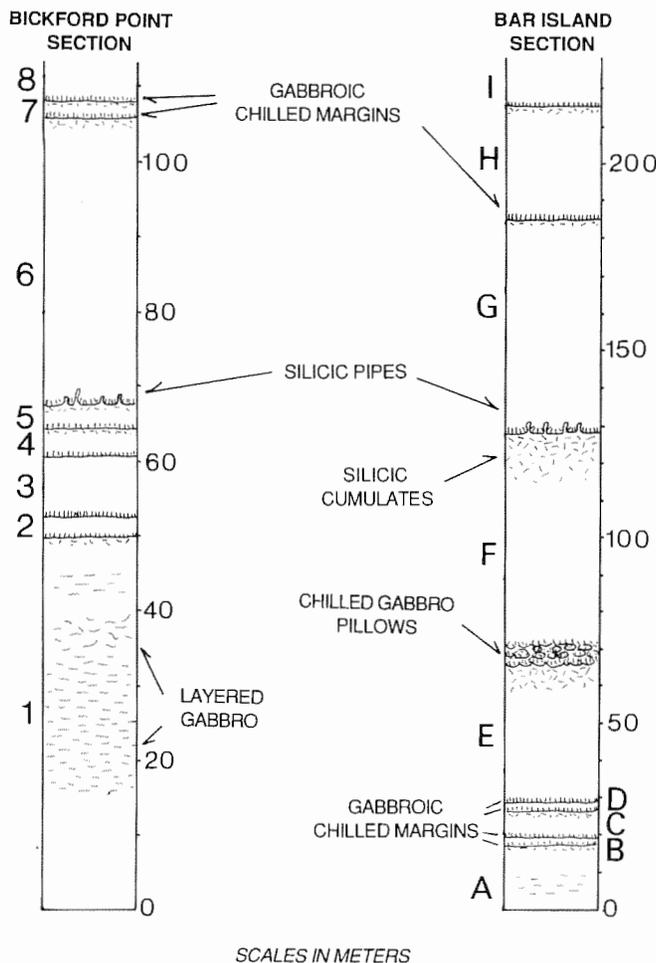


Figure 3. Stratigraphic sections from two locations in the Pleasant Bay intrusion. The Bickford Point section is at Stop #1 in Fig. 2; the Bar Island traverse is at Stop #2. Unpatterned parts of the columns are massive medium-grained gabbro. Individual macrorhythmic units are labelled by letters and numbers alongside the sections. All transitions between different rock types in a macrorhythmic unit are gradational.

unit. Each chilled base records the infusion of basaltic magma into the magma chamber. Cumulates beneath a macrorhythmic unit vary widely in composition, indicating that the composition of resident magma in contact with the chamber floor also varied widely when basalt was injected. The degree of chilling at the base of a unit generally increases as the compositional contrast with the underlying cumulates increases. Although the base of each layer consists of similar chilled gabbroic material with an irregular crenulate, convex-downward boundary, the extent and rate of compositional variation toward the top of each layer is highly variable. Extrapolation from measured sections suggests that the total number of macrorhythmic units in the intrusion probably exceeds several hundred.

In addition to the macrorhythmic units, some areas of the intrusion have many smaller gabbroic layers and lenses (0.5 to 2 meters thick) that have chilled upper and lower margins against more leucocratic dioritic material. In some areas, parts of layered sequence also contain closely spaced, chilled gabbroic pillows in a dioritic matrix.

**Veins, pipes, and diapirs.** A variety of veins, diapirs and pipes project from the tops of dioritic layers into overlying basally chilled gabbro. Leucocratic, often pegmatitic granite is commonly concentrated where dioritic material extends between convex-downward lobes of fine-grained gabbro and commonly connects to sharply defined, cross-cutting granitic veins in the overlying gabbro. These structures appear to represent filter-pressing of granitic liquid from the dioritic cumulates. Well-laminated diorite also appears to have moved upward as bulbous diapirs and pipes into chilled gabbro. These structures were apparently caused by the density inversion of light dioritic crystal mush beneath denser, rapidly-cooling, but still mobile gabbroic material. Pipes range in diameter from a few cm to more than 3 m. Their long axes are roughly perpendicular to layering, suggesting that the chamber floor was initially nearly flat and bowed downward during consolidation (Wiebe, 1993b). Gabbro in contact with a pipe is typically chilled near the base of the pipe, and becomes coarser-grained and more contaminated at higher levels. Diorite in the pipes typically has a weak lamination parallel to the pipe margins. Pipe compositions commonly grade upward from diorite at the base to pegmatitic leucogranite and open vugs.

### Petrography

Rocks of the Pleasant Bay intrusion can be placed into four broad petrographic groups: (1) fine-grained gabbroic rocks and basaltic dikes, (2) medium-grained gabbroic rocks, (3) leucocratic dioritic layers (diorite to granodiorite), and (4) granitic dikes and veins.

(1) Chilled margins of gabbroic layers and pillows and basaltic dikes have basaltic textures marked by thin, radial clusters of plagioclase laths and strong normal zoning (An<sub>65-35</sub>). Average grain-sizes typically increase from 0.1 to about 1 mm within several cm of the chilled margins. Euhedral olivine and plagioclase phenocrysts occur sparsely in some rocks. Olivine and augite are the dominant mafic phases; brown hornblende rims are widespread; and many rocks are dominated by brown hornblende. Fe-Ti oxide minerals, biotite, and apatite are ubiquitous accessory minerals. Minor interstitial quartz and alkali-feldspar occur locally. Partly resorbed xenocrysts of quartz and feldspar occur in many chilled rocks. Chilled basaltic pillows (either enclosed by granite in composite dikes or stratabound within silicic cumulates) commonly have margins that are greatly enriched in hornblende and biotite.

(2) Medium-grained gabbroic rocks range from olivine-rich gabbroic cumulates to massive quartz-bearing gabbros to scarce highly evolved Fe-rich cumulates. The most common gabbroic rocks range from hornblende-poor olivine gabbro and gabbro-norite to hornblende gabbro that lacks olivine and pyroxene. Orthopyroxene rims on olivine are ubiquitous. Titanomagnetite, ilmenite, biotite and apatite are common accessory minerals; brown hornblende commonly rims the olivine and pyroxene and is particularly abundant near leucocratic dioritic layers. The most primitive gabbros have cumulus olivine and plagioclase ± augite. Plagioclase typically has strong normal zoning (An<sub>65-35</sub>) and commonly displays complex oscillatory and patchy zoning. Plagioclase zoning and the abundance of intercumulus material indicates these rocks are orthocumulates. The most evolved rocks contain cumulus oligoclase, fayalite, Fe-Ti oxides and apatite.

(3) Leucocratic dioritic layers include rocks that range in composition from diorite to granodiorite (in the classification scheme of Streckeisen, 1976). Some rocks with higher color index (CI) are commonly dominated by subhedral calcic to intermediate, zoned plagioclase similar to that in the gabbro except that sodic patches and rims are more abundant. Other dioritic rocks contain antiperthitic sodic plagioclase with little zoning (An<sub>30-20</sub>) or coarsely exsolved ternary feldspar. Perthitic alkali feldspar occurs as separate subhedral grains in some silicic layers. Quartz is generally absent or occurs as a minor interstitial phase. Both the textures and the high total feldspar content of many diorites indicate that they are feldspar cumulates. Brown hornblende is generally more abundant than pyroxene except in some more silicic dioritic rocks where two pyroxenes or fayalite may dominate. Hornblende commonly contains irregular cores of relict augite or cores of fibrous cummingtonite that apparently replaced primary orthopyroxene. Biotite is most abundant in the more mafic diorites. Titanomagnetite, apatite and zircon are common accessory minerals. Petrographic evidence for hybridization is widespread: alkali feldspar occurs as cores in plagioclase; many samples contain both strongly zoned calcic plagioclase laths and larger tabular crystals of weakly zoned antiperthitic oligoclase; the more mafic diorites commonly contain biotite and quartz xenocrysts rimmed by clinopyroxene or hornblende.

(4) The granitic dikes are relatively fine-grained and have roughly equal proportions of plagioclase, perthitic alkali-feldspar, and quartz. CI is typically less than 10, with biotite more abundant than hornblende. Plagioclase crystals show both normal and oscillatory zoning (An<sub>38-10</sub>). Accessory opaque minerals and apatite are ubiquitous; zircon and allanite are common in many samples.

## Geochemistry

Approximately 100 samples from the Pleasant Bay intrusion were analyzed by XRF and ICP for major and trace elements (Wiebe, 1993b). These analyzed rocks fall into four groups: (1) chilled and fine-grained mafic rocks including dikes and layers, (2) fine-grained granitic dikes and veins, (3) medium-grained gabbroic layers, and (4) leucocratic dioritic layers. The first two groups approximate the compositions of basaltic and granitic liquids; the last two have compositions that grade to each other and are strongly affected by crystal accumulation.

Chilled and fine-grained mafic rocks from dikes and layers define smooth curved trendlines on most variation diagrams (Fig. 4). Most of the dike samples plot near the high-MgO end of these trendlines and also plot on the primitive ends of trendlines for incompatible trace elements (Fig. 4). These relations indicate that the dikes are almost certainly feeders for the basally chilled gabbroic layers in the intrusion. The Mg# [=  $100\text{Mg}/(\text{Mg}+\text{FeT})$ ] of chilled margins of layers varies widely, indicating that the mafic magmas underwent variable amounts of fractionation prior to being emplaced in the Pleasant Bay intrusion. In terms of major and trace elements, the least contaminated and most primitive basaltic dikes and chills more closely resemble tholeiites from back-arc basins and ocean islands (cf. Wilson, 1989) rather than basalts related to subduction or continental rifting (Wiebe, 1993b).

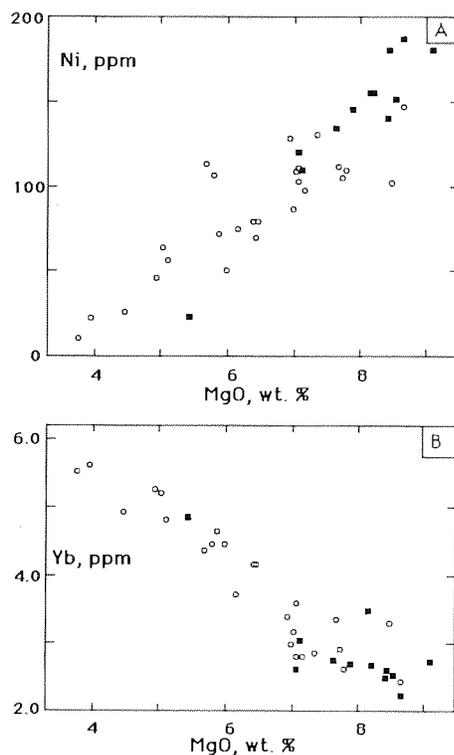


Figure 4. Whole-rock relationships of Ni and Yb to MgO in chilled basaltic material associated with the Pleasant Bay intrusion: chilled gabbroic layers and pillows (open circles) that are intercalated with dioritic layers, and basaltic dikes (solid squares).

Fine-grained granitic dikes and veins typically have 70-75%  $\text{SiO}_2$  with a moderate range in  $\text{K}_2\text{O}/\text{Na}_2\text{O}$ . Compared to dioritic layers in the intrusion, the dikes are higher in normative orthoclase and quartz and lower in anorthite (Wiebe, 1993b). Dike compositions plot in a narrow band along the low-temperature trough in the system quartz-albite-orthoclase (Fig. 5). Their compositions are consistent with crystallization at approximately 1 kb under  $\text{H}_2\text{O}$ -saturated conditions (Tuttle & Bowen, 1958).

Medium-grained gabbroic rocks have major- and trace-element compositions consistent with their being cumulates of plagioclase and olivine (+/- augite and Fe-Ti oxides) from basaltic liquids resembling in composition the basaltic dikes and chilled gabbroic margins (Wiebe, 1993b). Many of the medium-grained gabbroic layers have compositions essentially equivalent to those of the dikes and chilled layers.

Stratabound leucocratic rocks (diorite to granodiorite) have a wide compositional range that does not overlap the granitic dikes and veins (Fig. 5). These rocks are lower in  $\text{K}_2\text{O}$  and  $\text{SiO}_2$  and higher in  $\text{Al}_2\text{O}_3$  and  $\text{Na}_2\text{O}$  (Fig. 6), and their compositions are consistent with the petrographic evidence that they are cumulates dominated by sodic plagioclase.

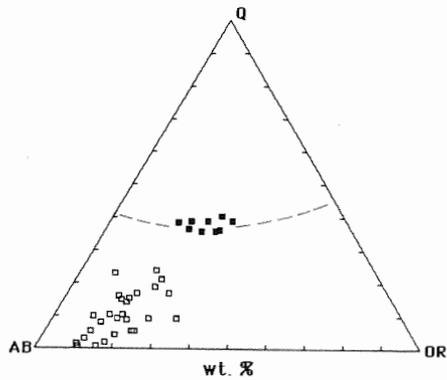


Figure 5. Plot of normative Q-Ab-Or for granitic dikes cutting the Pleasant Bay intrusion (solid squares) and dioritic to granitic layers intercalated with gabbro in the Pleasant Bay intrusion (open squares). The compositions of the granite dikes plot in a tight group near the 1 kbar  $\text{SiO}_2$ -Ab-Or- $\text{H}_2\text{O}$  minimum (Tuttle and Bowen, 1958).

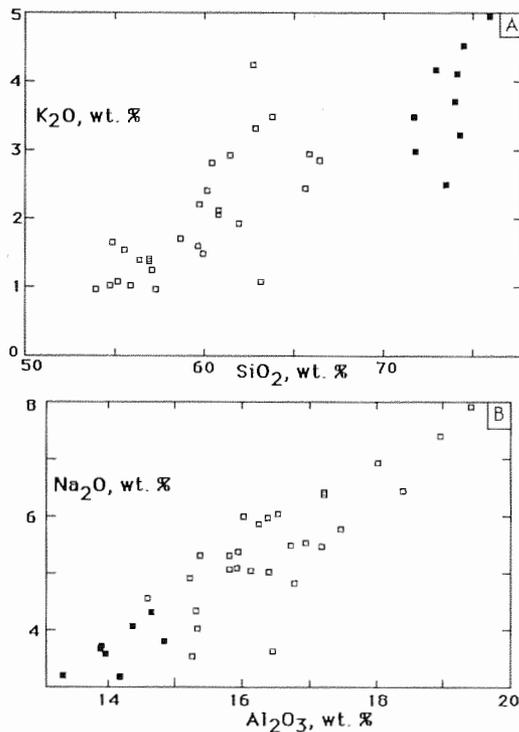


Figure 6. Plots of (A)  $\text{K}_2\text{O}$  vs  $\text{SiO}_2$  and (B)  $\text{Na}_2\text{O}$  vs  $\text{Al}_2\text{O}_3$  comparing dioritic cumulates (open squares) with granitic dikes (solid squares).

**Compositional variation in a macrorhythmic unit.** In order to characterize the stratigraphic compositional changes in a macrorhythmic unit, six samples from the upper part of macrorhythmic unit F in the Bar Island section (Fig. 3) were analyzed for major and trace elements. Most oxides and trace elements show smooth trends when samples are plotted against their stratigraphic positions (Fig. 7):  $\text{CaO}$ ,  $\text{MgO}$ , and  $\text{Sr}$  decrease continuously upward;  $\text{FeO}_T$ ,  $\text{TiO}_2$ ,  $\text{P}_2\text{O}_5$ , and  $\text{V}$  have maxima, reflecting the incoming of cumulus titanomagnetite, ilmenite, and apatite; and  $\text{K}_2\text{O}$ ,  $\text{Na}_2\text{O}$ ,  $\text{Al}_2\text{O}_3$ ,  $\text{Zr}$ , and  $\text{Ba}$  all increase strongly upward. The gabbroic cumulate near the base of the sampled interval has incompatible trace-element abundances that are comparable to those of a chilled gabbroic layer, suggesting (along with textural evidence) that it embodies a large percentage of trapped liquid. The uppermost dioritic cumulate contains cumulus ternary feldspar, Fe-Ti oxides, and pyroxenes, so it presumably crystallized from highly evolved, high-temperature silicic liquid. Its low Rb content suggests that it contains relatively little trapped liquid. Intercumulus liquid in the uppermost dioritic cumulates was probably largely removed by compaction and filter pressing caused by the loading of overlying macrorhythmic units. The expelled liquid probably escaped through silicic pipes that project upward from the top of this unit.

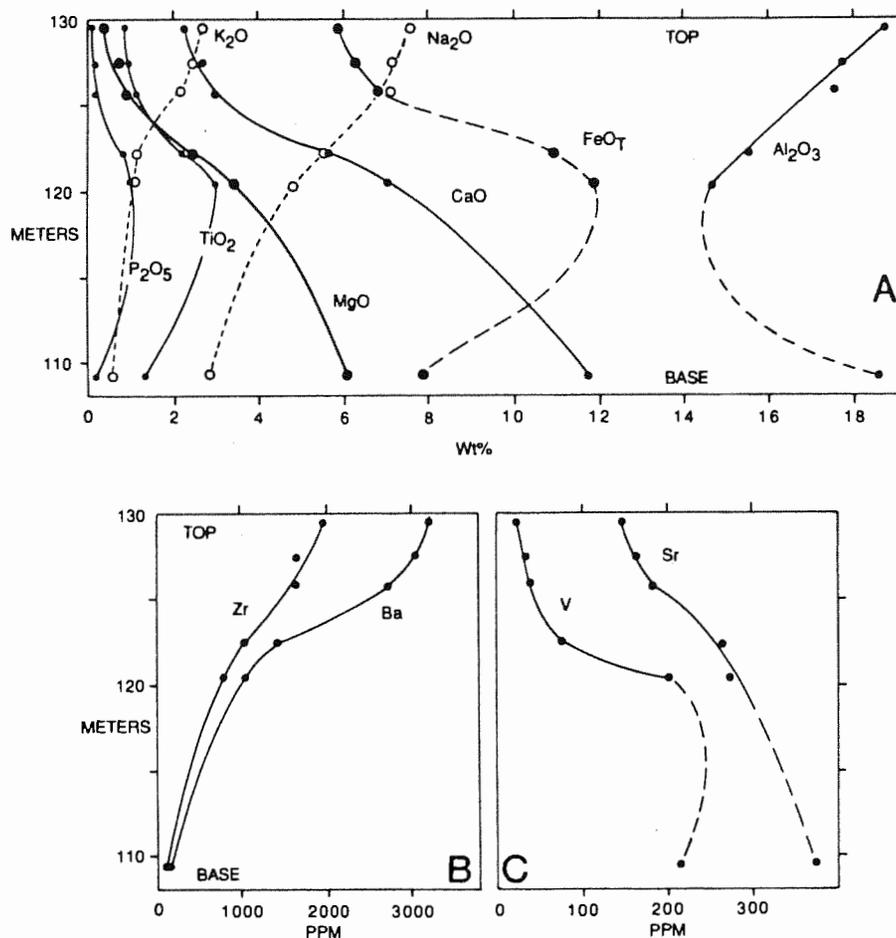


Figure 7. Whole-rock chemical variations in the upper 20 meters of macrorhythmic unit F of the Bar Island section illustrated in Fig. 3. This interval displays a continuous gradation from hornblende-rich gabbro with cumulus plagioclase and augite at the base to leucocratic diorite with cumulus ternary feldspar, two pyroxenes and Fe-Ti oxides at the top. (A) Major elements; (B) Zr and Ba; (C) V and Sr.

#### Interactions between layered basaltic and silicic magmas

Infusions of basaltic magma apparently occurred at different times when magma at the base of the chamber ranged from relatively primitive basalt to highly silicic magma. The compositions of chilled gabbroic margins suggest that most basaltic infusions were in equilibrium with plagioclase and olivine. The density of this newly resident magma should have increased while plagioclase, olivine and augite crystallized and begun to decrease when Fe-Ti oxides began to crystallize (Sparks and Huppert, 1984). Prior to the incoming of cumulus oxides, new infusions of basaltic magma should have tended to rise into and mix with resident basaltic magma. After resident magmas began to fractionate Fe-Ti oxides or hybridize with overlying silicic magma, basaltic infusions should have ponded on the chamber floor and established the base of a new macrorhythmic unit.

When silicic cumulates were forming on the floor, additions of basaltic magma should have spread across the chamber floor, ponding in the low spots over the silicic cumulates that just formed and displacing upward the silicic magma that had just been on the floor. The stratification of light, relatively cool silicic liquid over hot, dense basaltic liquid would be gravitationally stable, and, if the volume of basalt was sufficiently large, a double diffusive interface would form between the two magmas (Huppert and Sparks, 1984; Clark et al., 1987).

There are several lines of evidence which suggest that upward transitions in macrorhythmic units from gabbro to the most evolved silicic cumulates reflect, at least in part, strong compositional gradients established by mixing between two or more liquids rather than simple fractional crystallization of an originally homogeneous liquid:

(1) The bulk composition of most macrorhythmic units is significantly more silicic than the composition of the basaltic infusion (i.e., the chilled basal gabbro at the base of the unit). Resident silicic magma must contribute to the upper parts of these units.

(2) The lower gabbroic rocks in any macrorhythmic unit have textures and compositions characteristic of extreme orthocumulates, and some even approach liquid compositions. Rocks with such a large proportion of trapped liquid should be very inefficient at promoting fractional crystallization to the highly evolved silicic cumulates.

(3) The very high abundance of Zr and Ba of the highest diorite in Fig. 7 cannot have been produced by fractional crystallization of a basaltic parent liquid, especially considering that the transition occurs within 20 meters of section.

(4) There is abundant petrographic evidence for hybridization between two distinct magmas: alkali-feldspar cores in plagioclase, highly variable development of hornblende and biotite in gabbro, calcic plagioclase xenocrysts in diorite, and highly-digested enclaves of mafic rock in diorite.

(5) The lower gabbroic rocks in a macrorhythmic unit typically are richer in biotite and hornblende than the upper dioritic rocks, and the most evolved silicic cumulates are commonly anhydrous. Fractional crystallization should not produce increasingly dry evolved silicic cumulates.

(6) A study of Sr-isotopes indicates significant and systematic increases in Sr initial ratio across the transition from gabbro to diorite (Wiebe and Sinha, in prep.).

## THE CADILLAC MOUNTAIN INTRUSIVE COMPLEX

### Introduction

The Cadillac Mountain intrusive complex (CMIC) occurs on Mount Desert Island, Maine, and lies largely within the Acadia National Park (Fig. 8). It is defined here as consisting of three major units: the Cadillac Mountain granite, a hybrid unit of gabbroic to granitic rocks, and the Somesville granite. There are two areas of homogeneous Cadillac Mountain granite (CMG) that are separated by the hybrid unit of complexly interlayered gabbroic, dioritic, and granitic rocks (G-D) (Fig. 8). The Somesville granite (SG) forms a small intrusive body between the layered gabbro-diorites and the larger eastern area of the CMG. The western area of CMG also contains scarce mappable lenses of gabbro-diorite. These units occur in an oval area roughly 14 by 20 km. Gravity data (Hodge et al., 1982) suggest that the gabbro-diorite unit forms a saucer-shaped body, 2 to 3 km thick, (similar to the Pleasant Bay intrusion) beneath a saucer-shaped mass of CMG that is less than 3 km thick at its center. The 3-D shape of the Somesville granite is not well constrained, but gravity data suggest that it is thin. This intrusive complex appears to have been emplaced very close to an unconformity between the Ellsworth schist and the Bar Harbor Formation. The unconformity also appears to have controlled the emplacement of sill-like masses of homogeneous gabbroic rocks with diabasic textures located to the north and south of the CMIC. Steeply bounded masses of gabbro within the Ellsworth schist appear to have been feeders for the gabbroic rocks. Brief descriptions of the units that make up the country rock are provided in Gilman et al. (1988). Hornblende from the hybrid dioritic unit gives an Ar-Ar age of about 418 +/- 5 Ma (Lux, personal comm., 1993).

### Field relations in the CMIC

The larger eastern area of CMG is rimmed everywhere except on its western margin by a basal intrusive breccia (termed a "shatter zone" by Gilman et al., 1988). The western margin of the granite is either in contact with the gabbro-diorite unit or cut by the younger Somesville granite. The intrusive breccia contains blocks of the surrounding country rock (e.g. Ellsworth Schist, Bar Harbor Formation, gabbro); along its southern margin it also contains blocks of the Cranberry Island Series (mainly bimodal volcanics). The breccia appears to truncate the granite of Southwest Harbor. There is little or no evidence in this zone for commingling between mafic and silicic magmas.

Lensoid mafic enclaves occur sparsely in the larger eastern area of the CMG; their attitudes define a basin-form shape that is consistent with the shape indicated by gravity. The granite is cut by at least two arcuate zones of variably porphyritic granophyre that Chapman (1970) originally interpreted as recrystallized zones. The boundaries between granophyre and granite range from sharp to gradational. Near the summit of Cadillac Mountain the CMG contains rounded fine-grained "blobs" of granitic material up to a few meters in diameter. In some areas, the arrangements of these blobs define irregular curvilinear zones, suggesting they formed from the breakdown of silicic dikes passing upward through partially solidified granite. The blobs appear to be randomly distributed in other areas.

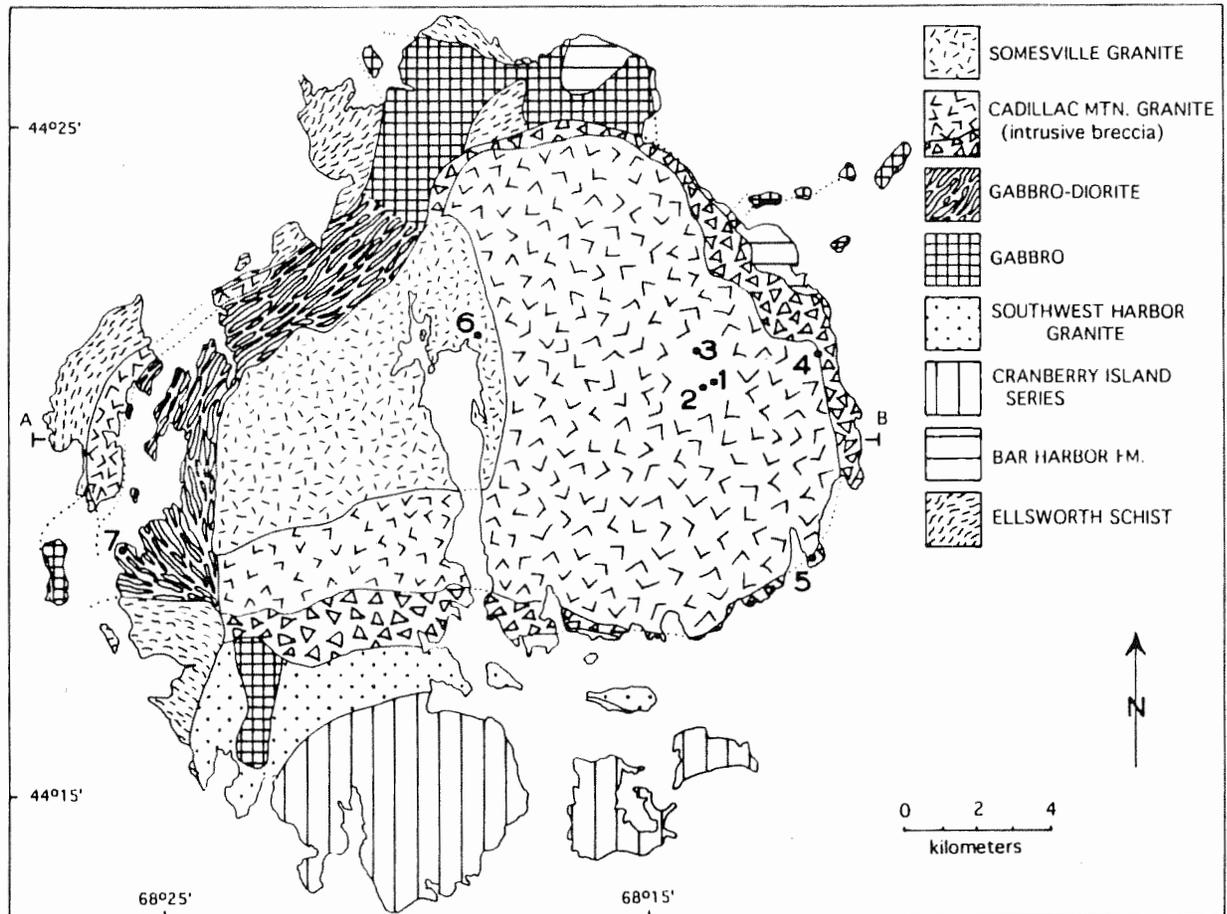


Figure 8. Geologic map of Mount Desert Island modified from maps in Chapman (1970) and Gilman et al. (1988). Numbers refer to field trip stops on day 2.

The CMG is sparsely cut by steep, roughly N-S trending basaltic dikes; rarely some of these dikes trail off into a linear zone of basaltic pillows in contaminated granite.

The smaller western area of the CMG contains several sheet-like masses of variably chilled and pillowed gabbroic material. These layers all appear to dip steeply to the east. Silicic pipes that are approximately perpendicular to one gabbro layer extend upward from the underlying granite through its chilled base. The orientations of the pipes, as in the Pleasant Bay intrusion, indicate that the gabbroic layers were initially deposited roughly horizontally on incompletely solidified granite. Other gabbroic layers clearly grade upward (to the east) to granite through hybrid diorite.

Field relations in the gabbro-diorite unit closely resemble those in the Pleasant Bay intrusion (Fig. 9). Much of this unit can be mapped as macrorhythmic layers that grade upward from basally chilled gabbro to coarser-grained gabbroic, dioritic, or granitic rocks. Compared with the Pleasant Bay intrusion, silicic layers in the G-D unit more commonly approach granitic (instead of dioritic) compositions, and the up-section transitions from gabbro to granite are generally more abrupt. Layering dips moderately ( $20-50^\circ$ ) to the east in an arcuate pattern roughly conformable with the basinform shape indicated by gravity for both the granite and the gabbro-diorite unit. The orientations of silicic pipes in mafic layers suggest that the layers were originally deposited close to the horizontal and were subsequently bowed downward as the chamber evolved and solidified. The approximate average exposed thickness of the G-D unit is about 1.5 km. This is somewhat less than the thickness of gabbroic rocks inferred from gravity studies to lie beneath the main body of the CMG and suggests that the gabbroic material reaches a maximum thickness near the center of the basin beneath the CMG.

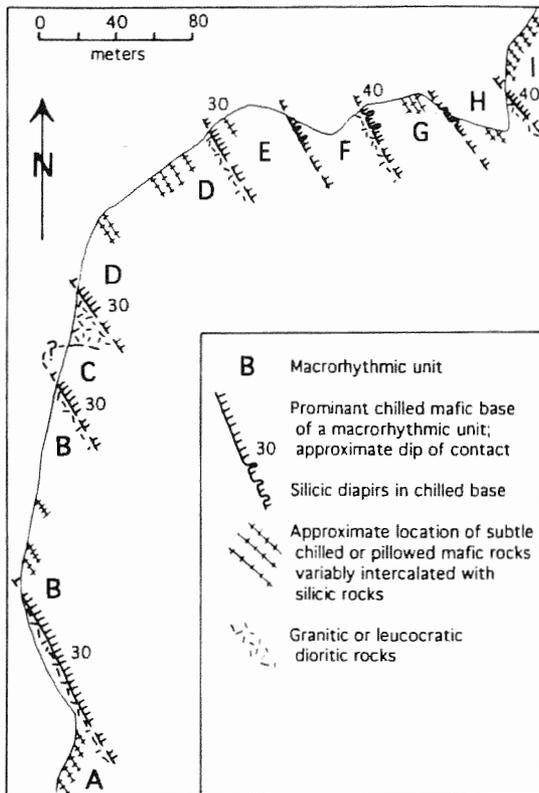


Figure 9. Simplified geologic map of the gabbro-diorite unit on Stewart Head (Stop #7 in Figure 8). Unpatterned areas along the coast are gabbro to mafic diorite.

### Petrography

**Cadillac Mountain granite.** The Cadillac Mountain granite is a relatively homogeneous, massive, medium- to coarse-grained, hypersolvus granite with CI less than 10. Both quartz and perthitic alkali-feldspar are equant and range in diameter from 2 to 7 mm. At lower elevations in the intrusion sodic plagioclase cores are common in alkali-feldspar, and many crystals show delicate oscillations of the two feldspars in the transition between core and rim. Hornblende is the dominant mafic mineral, minor biotite is common. Both minerals are typically interstitial. Opaque minerals, zircon and apatite are ubiquitous accessory phases; allanite, titanite, and fluorite occur sparsely. Some rocks near the eastern base of the intrusion close to the "shatter zone" are distinct from typical CMG in containing variable amounts of sodic plagioclase ( $An_{20}$ ), subhedral clinopyroxene and, rarely, fayalite.

Mafic clots (enclaves) commonly less than 1 cm in diameter are widespread throughout much of this granite (Seaman and Ramsey, 1992). They are dominated by small subhedral plagioclase and hornblende with subordinate opaque minerals and apatite. Their texture resembles some of the finer-grained mafic rocks in the gabbro-diorite unit. Disaggregation of these clots appears to explain scarce occurrences of small plagioclase and concentrations of hornblende crystals in the granite.

**Somesville granite.** The Somesville granite is a medium- to coarse-grained two-feldspar granite with a CI less than 10. It is dominated by large (up to 10 mm) equant alkali-feldspar with smaller (1-4 mm) equant quartz and subordinate, small (typically 0.3 to 1 mm), complexly-zoned plagioclase ( $An_{30-15}$ ). Biotite is the dominant mafic mineral; hornblende is scarce and commonly absent. Accessory phases include opaque minerals, apatite, titanite, zircon, and allanite. Fluorite is scarce. This granite also contains a variety of mafic enclaves (Seaman and Ramsey, 1992).

Several textural features suggest that this granite has been affected by hybridization with mafic material. Partly disaggregated mafic clots rich in hornblende and plagioclase occur sparsely, and hornblende commonly appears to be more abundant where small plagioclase is also more abundant. In many rocks, some plagioclase crystals have rounded, more sodic cores inside more calcic zones; rarely, plagioclase rims occur on irregular, cores of alkali-feldspar. Both types of zoning appear to record a corrosional event related to magma mixing.

Granitic dikes that sharply cut the gabbro-diorite unit have minerals and textures that closely resemble the Somesville granite.

**Gabbro-diorite unit.** Because of the the wide compositional range of rocks in this unit, petrographic description is broken down into three broad groups: gabbroic, dioritic, and granitic. All of these rocks are found interlayered and all are interpreted as having formed by sequential deposition on the floor of a magma chamber - in a manner comparable to the layered rocks in the Pleasant Bay intrusion. The gabbroic rocks range from chilled margins that texturally resemble basaltic dikes to medium-grained mafic cumulates. The dioritic rocks show the greatest petrographic evidence for hybridization.

Chilled margins of gabbroic layers typically have basaltic textures with radiating thin lathes of plagioclase comparable to the basaltic dikes that cut all units. Plagioclase is strongly zoned (An<sub>70-30</sub>). The mafic minerals in chilled margins of gabbroic layers are dominated by hornblende with subordinate augite, orthopyroxene, and opaque minerals - in contrast to the basaltic dikes which are generally dominated by augite with subordinate olivine. Biotite is scarce in the dikes and more common in the chilled layers. Xenocrysts of alkali-feldspar, sodic plagioclase, and quartz occur sparsely in dikes and chilled layers and are commonly rimmed by augite or hornblende.

Coarser-grained gabbroic layers are typically massive with randomly arranged plagioclase. Some have subhedral olivine; augite and/or hornblende are typically ophitic to subophitic. Plagioclase commonly shows strong normal zoning (e.g. An<sub>70-30</sub>); in many rocks it is complexly zoned with sodic patches, oscillations and strong reversals. Accessory opaque minerals and apatite are ubiquitous. Many of the gabbroic rocks can be described as extreme orthocumulates with cumulus plagioclase, olivine +/- augite and opaque minerals.

Gabbroic rocks commonly grade upward to intermediate dioritic rocks that display abundant petrographic evidence for hybridization between mafic and silicic magmas. Although most diorites are dominated by plagioclase and hornblende, many contain equant alkali-feldspar crystals or cores of alkali-feldspar in some plagioclase crystals. Crystals of intermediate plagioclase (An<sub>50-40</sub>) commonly contain rounded cores of more sodic plagioclase. Augite is common and typically has thick rims of hornblende. Blocky subhedral biotite crystals occur in some diorites and are commonly included by hornblende. Some rocks can be characterized as orthocumulates with subhedral cumulus plagioclase and hornblende. Accessory opaque minerals and apatite are ubiquitous; zircon, titanite and allanite are commonly present.

The most silicic layers in the G-D unit closely resemble the Cadillac Mountain granite.

## Geochemistry

**Introduction.** Approximately 120 rocks from the CMIC have been analyzed by XRF and ICP for major and trace elements. Representative compositions of the main units are given in Table 1.

**Cadillac Mountain Granite.** Except for very minor cumulates of sodic plagioclase +/- clinopyroxene near the eastern base, medium-grained rocks from the CMG range between about 72 to 75% SiO<sub>2</sub>. All rocks are low in CaO and have very low Mg#[=100Mg/(Mg+Fe)]. They show little compositional variation in major elements, but a relatively wide range in many trace elements (e.g., Rb, Ba) (Fig. 10). This compositional range and the petrographic

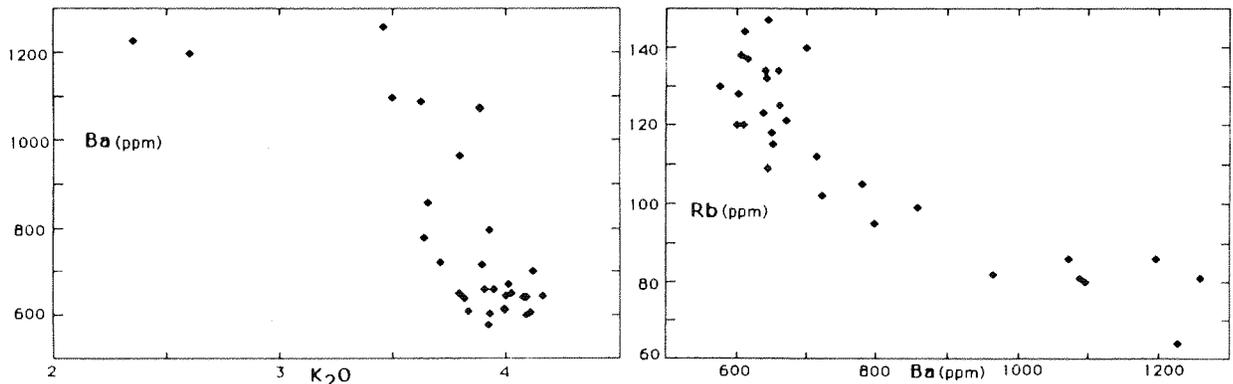


Figure 10. Plots of K<sub>2</sub>O vs Ba and Ba vs Rb for the Cadillac Mountain granite.

characteristics of the CMG are typical of many A-type granites. Samples taken along a traverse from the eastern margin of the body to the summit of Cadillac Mountain (approximately between field-trip stops 4 and 1 in Fig. 8) appear to indicate a significant compositional variation with elevation: granitic rocks near the eastern base are relatively enriched in Ba and Zr and depleted in some incompatible elements (e.g., Rb, Be) (Fig. 11). These relations suggest that the compositions of the lower granites have been affected by accumulation of feldspar and zircon.

Table 1. Representative chemical analyses of rocks from the Cadillac Mountain intrusive complex.

	Basaltic dike	Chilled gabbro	Gabbro-diorite unit (mafic to silicic cumulates)					Cadillac Mountain granite		Somesville granite	
	32C	6A	67Y	101A	126B	117C	112B	81	93	151	150
SiO <sub>2</sub>	48.85	46.63	46.31	50.58	56.83	66.40	70.30	72.94	74.43	76.56	74.04
TiO <sub>2</sub>	1.35	1.12	1.27	1.00	1.75	0.92	0.74	0.30	0.32	0.17	0.26
Al <sub>2</sub> O <sub>3</sub>	16.41	17.80	13.78	19.15	15.58	13.36	13.39	12.79	12.28	12.83	13.40
Fe <sub>2</sub> O <sub>3</sub>	1.99	1.33	2.05	1.75	3.57	2.46	1.59	1.96	1.47	0.40	0.51
FeO	6.98	8.50	9.70	7.65	6.15	3.57	2.43	1.62	1.54	0.81	1.00
MnO	0.18	0.15	0.19	0.16	0.16	0.09	0.07	0.12	0.09	0.03	0.04
MgO	8.35	8.47	12.89	3.52	2.54	1.45	1.26	0.15	0.23	0.14	0.33
CaO	10.57	9.44	8.72	10.15	5.96	3.30	2.03	0.80	0.84	0.93	1.35
Na <sub>2</sub> O	2.38	2.91	2.56	3.03	4.07	4.20	3.45	4.61	3.94	3.38	3.45
K <sub>2</sub> O	0.66	0.59	0.37	0.50	1.05	2.37	4.04	3.76	4.06	4.63	4.24
P <sub>2</sub> O <sub>5</sub>	0.13	0.13	0.17	0.09	0.54	0.22	0.14	0.03	0.05	0.00	0.03
LOI	2.25	3.80	1.24	1.47	2.06	1.00	1.35	0.67	0.70	0.37	0.53
TOTAL	100.10	100.87	99.25	99.05	100.26	99.34	100.79	99.75	99.95	100.25	99.18
Rb	28	23	6	10	26	54	130	82	120	198	177
Sr	219	200	207	441	349	172	143	48	53	29	69
Ba	85	91	97	142	219	511	572	964	601	324	444
Y	28	27	27	17	55	69	48	41	93	54	41
Zr	134	101	111	58	430	662	249	651	574	159	212
V	212	198	195	241	154	73	68	4	6	9	15
Ni	102	145	325	19	11	14	14	4	4	3	2
Cr	246	250	911	10	7	12	19	0	0	4	9
Be	1.2	1.7	0.9	1.4	1.2	2.3	2.3	1.8	3.3	2.9	2.9
Co	43	44	49	33	0	17	18	0	0	1	1
Sc	36	33	32	30	20	14	10	7	5	4	5
Ce	23	22	27	20	77	95	77	123	147	66	83
Yb	3.2	2.7	2.3	ND	5.2	6.9	4.7	5.0	9.2	6.5	4.7

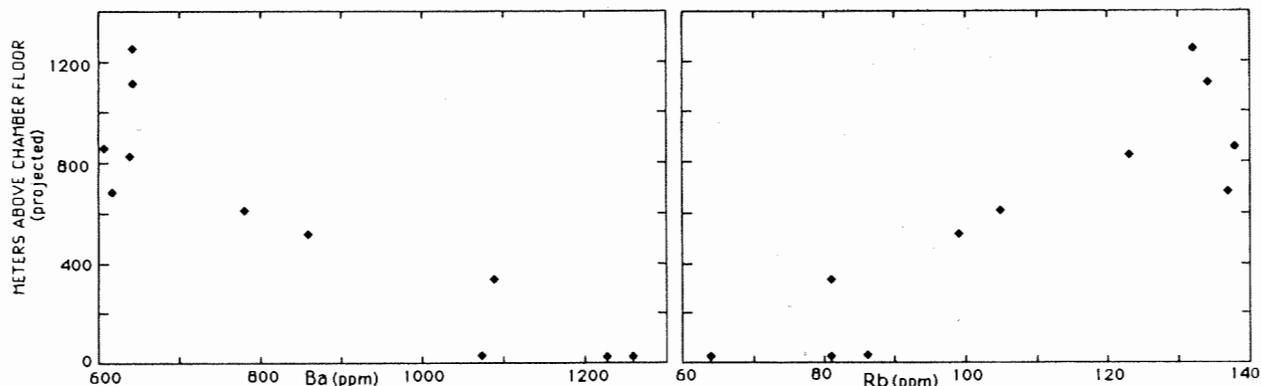


Figure 11. Plots of Rb and Ba against elevation above the floor of the eastern side of the Cadillac Mountain granite. The line of section is approximately from stops 4 to 1 (see Figure 8). The floor of the granite was projected along a dip of  $30^\circ$  to the west.

The silicic "blobs" that occur near the summit of the Cadillac Mountain and the transgressive granophyre zones have compositions that generally define smooth trends on plots of major and trace elements against  $\text{SiO}_2$ . They range in composition from about 70 to 78%  $\text{SiO}_2$ . Between 72 and 75%  $\text{SiO}_2$  their compositions are nearly identical to the CMG; at high  $\text{SiO}_2$  they trend to high values in  $\text{K}_2\text{O}$  and Rb and very low values in MgO, CaO, FeO,  $\text{TiO}_2$  and Sr (Fig. 12). These most evolved silicic blobs have compositions that closely resemble high- $\text{SiO}_2$  rhyolites associated with ash-flow tuffs.

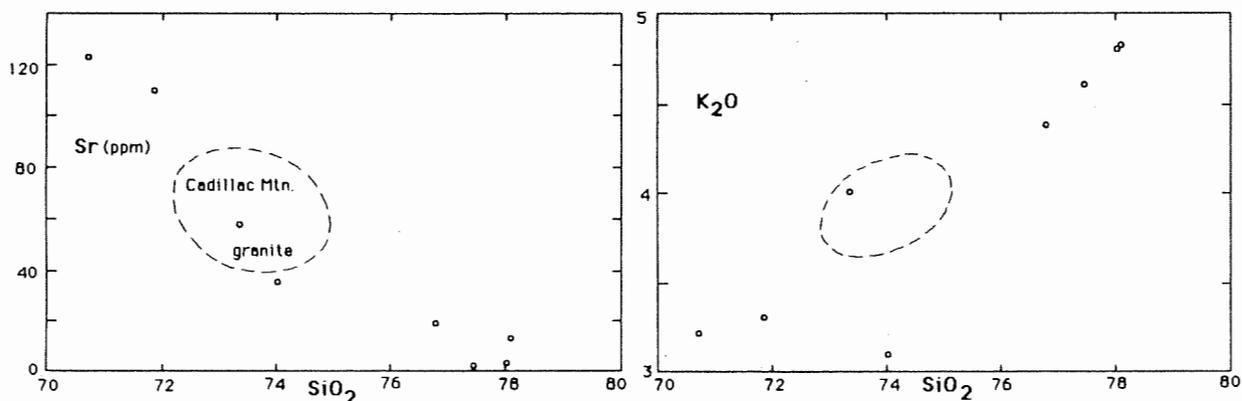


Figure 12. Plots of Sr and  $\text{K}_2\text{O}$  against  $\text{SiO}_2$  for chilled silicic "blobs" (stops 1 and 2 in Figure 8) and transgressive bodies of granophyre in the Cadillac Mountain granite. Dashed areas enclose compositional ranges of typical CMG.

**Somesville and other granitic rocks.** Relative to the CMG, the Somesville granite is characterized by higher  $\text{SiO}_2$ ,  $\text{K}_2\text{O}$  and Rb, and lower Ba, Zr, and Ce (Table 1). On plots of major and trace elements against  $\text{SiO}_2$  the Somesville granite is distinct from the CMG and, instead, appears to fall close to or within the compositional trends defined by smaller granitic plutons and dikes that cut the gabbro-diorite unit (Fig. 13). The most silicic samples of the Somesville granite appear to approach closely the compositions of the most silicic blobs within the CMG.

**Gabbro-diorite unit.** The chilled pillows and margins of mafic layers that occur as part of this layered sequence of rocks have compositions that appear to approximate mafic liquids and that are very similar to the N-S trending basaltic dikes (Table 1). Both suites show a similar range in MgO and comparable trends for both major and trace elements plotted against MgO (Fig. 14). These relations strongly suggest that the basaltic dikes represent feeders to the gabbroic component of the G-D unit. The wide compositional range and trends of the chilled liquids are consistent with variable fractional crystallization of plagioclase, olivine and augite at depth, prior to the emplacement of the basaltic magmas into the CMIC.

The compositions of medium-grained rocks from layers in the G-D unit range from about 45 to 75% SiO<sub>2</sub> with a scarcity of samples between 50 and 55% SiO<sub>2</sub> and between 7 and 4% MgO (Fig. 15). This compositional gap coincides with a gap in CIPW normative anorthite (between An<sub>55</sub> and An<sub>40</sub>). The rocks with lowest SiO<sub>2</sub> have compositions that are consistent with their being cumulates from the mafic chilled liquids and dikes. The rocks with highest SiO<sub>2</sub> closely approximate the average compositions of the CMG. The compositions of rocks with intermediate SiO<sub>2</sub> cannot be solely explained by bulk mixing between basaltic and granitic compositions. Instead, their compositions appear to reflect (as do their textures) some hybridization between mafic and silicic magmas, along with the accumulation of plagioclase feldspar and variable amounts of other phases, including pyroxene, hornblende, Fe-Ti oxides, and apatite.

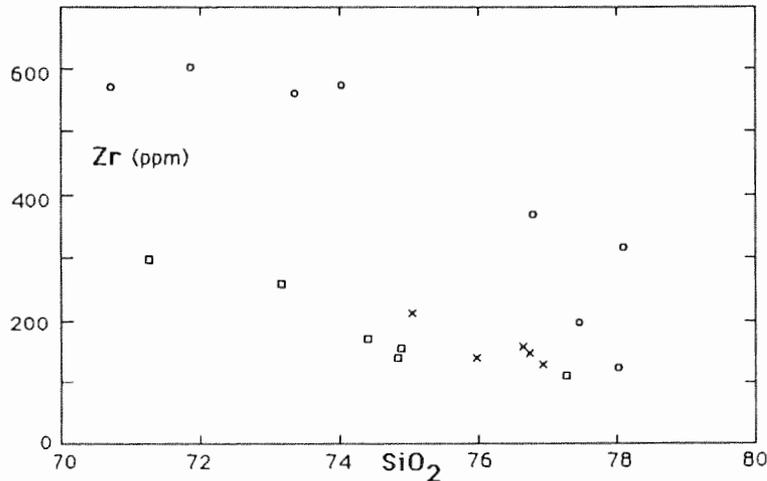


Figure 13. Plot of Zr against SiO<sub>2</sub> for silicic blobs in CMG (open circles), the Somesville granite (x's), and granitic dikes that cut the gabbro-diorite unit (open squares).

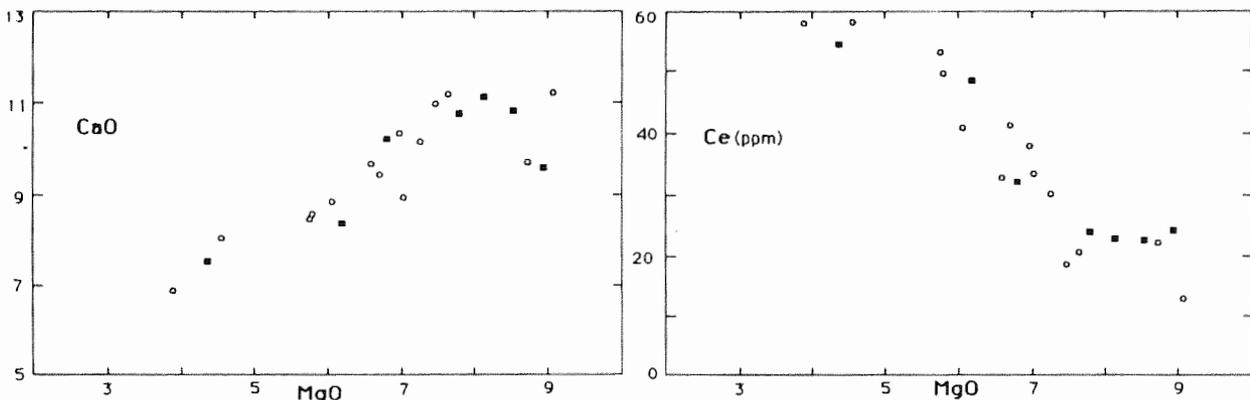


Figure 14. Plots of CaO and Ce against MgO for basaltic dikes (solid squares) and chilled gabbroic layers within the gabbro-diorite unit (open circles).

### Evolution of the CMIC

A tentative model for the evolution of the Cadillac Mountain intrusive complex can be discussed with the aid of a schematic cross-section which is reasonably well constrained by field relations and gravity studies (Fig. 16). The CMIC appears to have been initiated by injections of silicic magma that established a sub-horizontal, lensoid chamber at the unconformity between the Ellsworth schist and the Bar Harbor Fm. It is possible that injections of mafic magma had already occurred at the same level, establishing some of the mafic sills north, east, and south of the CMIC. Once established, the chamber of the CMG acted as a trap for later basaltic injections which ponded on its floor, leading at many times to strong compositional gradients in liquids near the base of the chamber. The solidification of these multiple injections of mafic magma ultimately produced the complexly layered gabbro-diorite

unit. Episodic turbulent events at the boundary between the liquids probably led to local mechanical mixing between rapidly solidifying mafic magma and silicic magma. The small mafic enclaves in the CMG probably originated during these events. Their distribution at all elevations in the CMG strongly suggests that silicic magma in the chamber underwent thorough convection.

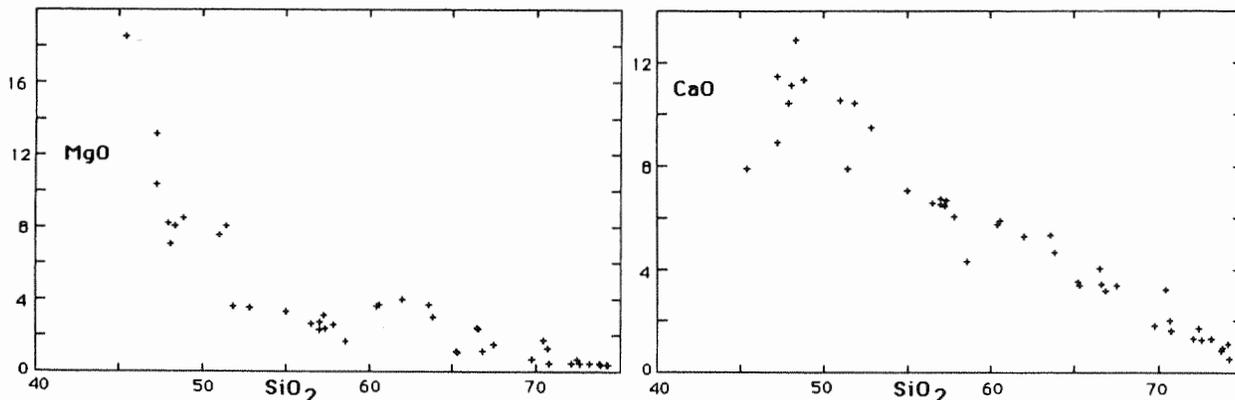


Figure 15. Plots of MgO and CaO against SiO<sub>2</sub> for medium-grained layers (gabbroic to granitic) in the gabbro-diorite unit.

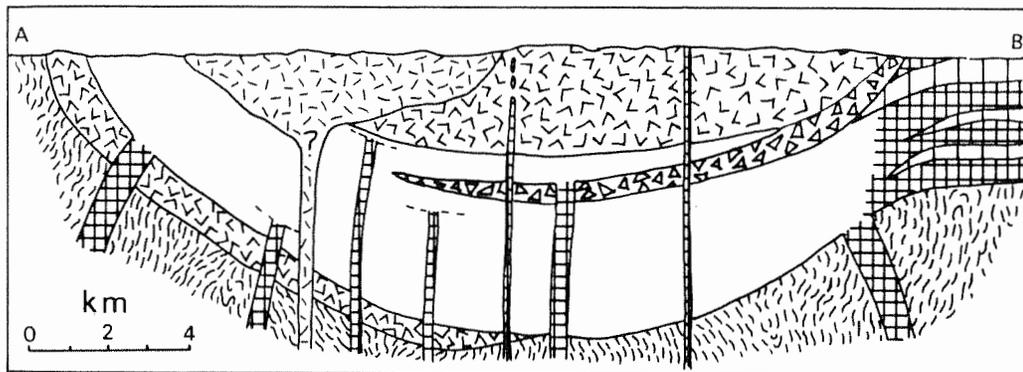


Figure 16. Schematic east-west section through Mount Desert Island drawn along a section line between A and B in Figure 8. No vertical exaggeration. Patterns as in Figure 8 except that here the hybrid gabbro-diorite unit is unpatterned.

The extensive intrusive breccia that forms the northern, eastern, and southern boundaries of the CMG may represent a single large event that resulted in eastward expansion of the silicic magma chamber. It is likely that some basaltic magma continued to enter the chamber after this event because basaltic dikes of comparable composition continued to be emplaced even after the upper portions of the CMG were solidified.

The strong compositional gradients in the lower elevations (in Rb, Ba, Be, Zr, etc.) suggest that, early on, while convection was still active, there was some tendency to accumulate early-formed crystals (e.g. alkali-feldspar, zircon) on the chamber floor and to concentrate residual liquid upward. The apparent lack of compositional gradients at higher elevations may indicate gradual stagnation of the chamber as infusions of basaltic magma waned.

The silicic blobs near the summit of Cadillac Mountain and the arcuate zones of intrusive granophyre appear to represent rapidly quenched samples of evolved silicic magma that was erupting out of the chamber after its roof was largely solidified. Field relations suggest that all of this granophyric material could have been approximately contemporaneous because the granitic mush, through which the evolved silicic melt rose, was relatively brittle (cooler and more crystallized) toward the margins of the CMG and relatively plastic (hotter and less crystallized) in the interior.

The Somesville granite may represent a late stage liquid related to the CMG. The chemical compositions of the most silicic Somesville granites closely resemble the most silicic chilled granophyric liquids within the CMG, and field relations suggest that the Somesville granite and the CMG granophyres could have been contemporaneous. Much of the geochemical and petrographic character of the less silicic Somesville granites can be explained by contamination with partly crystallized mafic magma. This interpretation is consistent with the fact that the SG is

restricted to an area immediately above the gabbro-diorite unit. Also, outcrops along the contact of these units indicate that mafic magma was available when the Somesville granite was emplaced.

### THE ISLE AU HAUT IGNEOUS COMPLEX

#### Introduction

The Granite-Gabbro-Diorite rocks of the Isle au Haut Igneous Complex (as named by Hogan and Sinha, 1989) are situated at the southernmost exposed sequence of the Coastal Maine Magmatic Province, on the island of Isle au Haut. The island is located about 8 km south of Stonington, on Deer Isle, Maine. The complex is composed mainly of a heterogeneous 390 Ma granite (28 km<sup>2</sup>) (Pb/U ages by R. W. Luce, personal comm., 1987) which intrudes older silicic volcanics to the west. The eastern third of the island consists of a layered 413 Ma gabbro-diorite-quartz monzodiorite complex (51 km<sup>2</sup>) (Fig. 17). Most of the mafic units on the island proper are diorites; however, a coarse-grained layered gabbro forms most of the small islands and ledges to the east. A homogeneous granite intrusion crops out on small islands to the east of the mafic complex and field evidence indicates that it has probably been emplaced contemporaneously with the layered gabbros of the mafic complex. We have found no direct field evidence for comagmatic interaction of the western Isle au Haut granite with the mafic rocks, and suspect that it may have a younger intrusive relationship given the preliminary ages mentioned earlier. However, because of the large uncertainty in ages ( $\pm 17$  Ma) it is not inconceivable that this granite could also be contemporaneous with the mafic rocks; possibly down-faulted to its present position. Most of the mafic rocks exposed on Eastern Head peninsula (the southern end of the island) are gabbro, locally interlayered with diorite and quartz monzodiorite (Fig. 18). Regional structural events have tilted these layers to a post magmatic attitude of N 10° E; 35° W which has exposed an almost complete and continuous cross-section of ten layers within the mafic complex (Fig. 19).

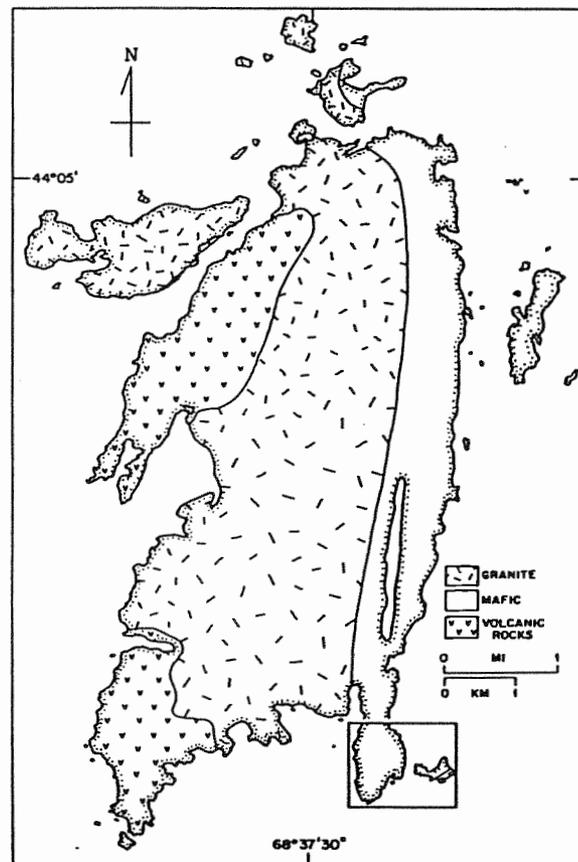


Figure 17. Generalized geologic map of the lithologic units which comprise the Isle au Haut Igneous Complex, taken from Smith et al. (1907) and Luce (1962). Box shows area of our concentrated research.

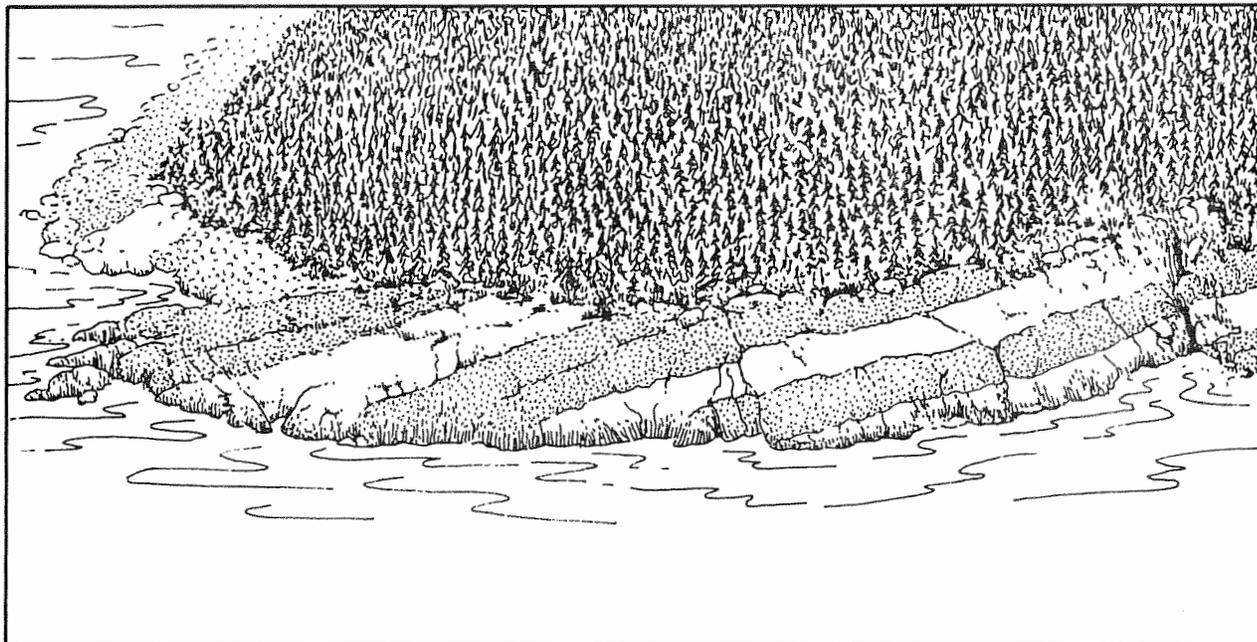


Figure 18. Sketch of aerial photograph shows alternating sequence of layers exposed along the southern shoreline of Eastern Head.

### Field Relations

The southernmost exposures on Isle au Haut reveal an alternating layered sequence of five diorites *sensu lato* and five gabbros (units AA - I) dipping to the west (Fig. 20). The gabbroic units are biotite hornblende gabbros and range in thickness from 7 to 106 m. At their bases, they have very fine-grained contacts, approximately 4-5 cm thick, which appear to have chilled against the underlying dioritic units. Immediately above this chilled rind, the grain-size progressively coarsens over 5 meters to a coarse gabbro with 6 mm plagioclase laths and large 11 mm equant hornblende crystals. Local wisps of coarse plagioclase-rich pegmatitic crystals occur randomly throughout the units.

At the chilled interface of the gabbro and the underlying diorite units are lobate, cusped structures (Fig. 21) which can best be described as "load cast structures" (Pettijohn *et al.*, p. 124, 1972; Reineck and Singh, p. 84, 1980). Others have found it necessary to adopt this sedimentological term for similar igneous structures (Wiebe, 1974, 1993a,b; Thy and Wilson, 1980; Parsons and Becker, 1986). Within the dioritic units, immediately below the chilled contacts, are randomly spaced, globular, mafic "pillow-like" structures ranging in size from a few centimeters up to a meter in diameter. Similar features have been described as "load ball structures" (Parsons and Becker, 1986); however, due to their chilled texture, we prefer to use "pillow" to convey both a structural and petrogenetic connotation. Although commonly found less than one half meter from the contact, some "pillows" are completely detached from the gabbro. They are very fine-grained, with chilled margins, and are texturally and chemically identical to the fine-grained margins of the gabbro from which they were detached. Both the "load cast structure" of the chilled gabbroic contact and the chemistry and texture of these "pillows" with their adjacent contacts provide strong evidence for the gabbro chilling against the dioritic units, and for the sinking of largely molten "pillows" into the underlying unconsolidated diorite.

At the same interfaces, fingers or pipes of a part of the dioritic unit penetrate upwards into the gabbro unit for several meters (Fig. 21). These fingers and pipes plunge  $53^{\circ}$ ;  $E 9^{\circ}$  S, a direction which coincides with the poles to the planes of the layered units, confirming that the magma layers were originally horizontal when emplaced.

The pipes are cylindrical in form and tend to be spaced about 1 - 1.5 meters apart. The gabbro is chilled against the pipes for tens of centimeters above the diorite-gabbro contact. The pipe diameter tends to increase from about 8 cm. to as much as 50 cm with increasing height in the gabbro (15 meters). Local large conduits over 1.6 meters in diameter have also been found. With increasing height in the gabbro unit, the frequency of the pipes decreases and the contact of the dioritic pipes with the gabbro becomes more diffuse. At higher levels, xenoliths of the gabbro occur within the dioritic pipes. These xenoliths include hybrid compositions as well as gabbros. The ultimate fate, and relationship of these dioritic pipes to the uppermost part of the thicker gabbro layers, is not fully understood.

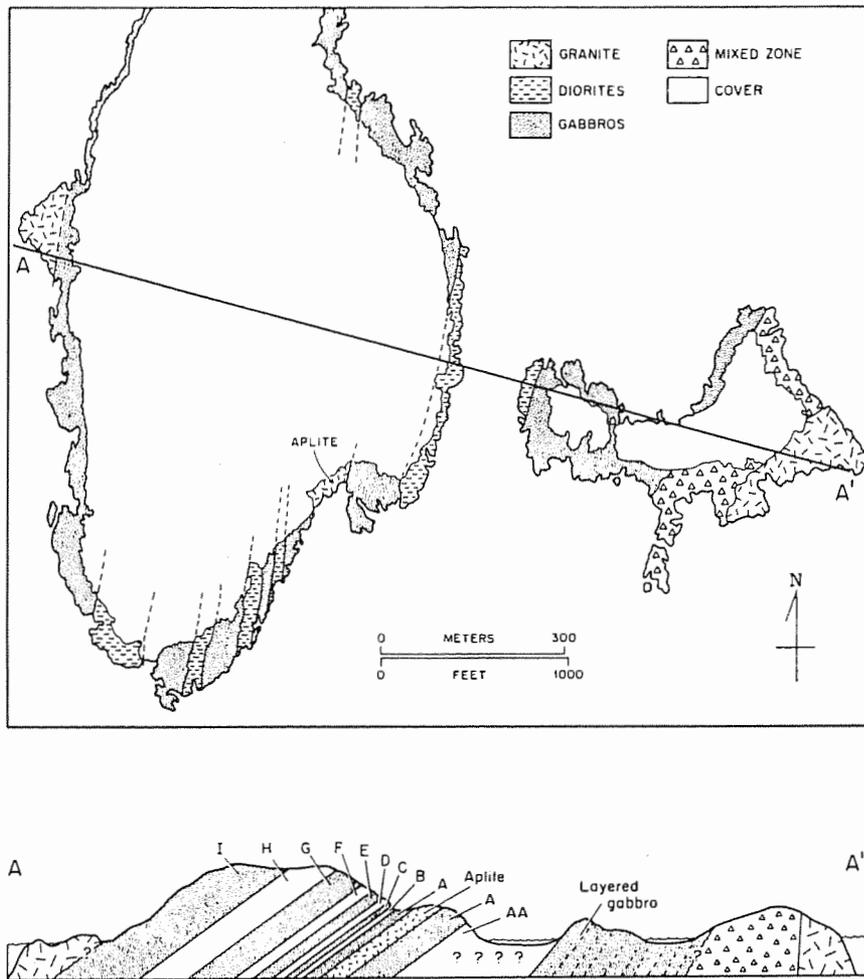


Figure 19. Geologic map and cross-section (vertical exaggeration 4x) of the composite layering of Gabbro-Diorite units and granites exposed on Eastern Head peninsula and Eastern Ear island.

With thinner gabbro layers (e.g. layers C and E), the pipes can be traced almost entirely to the overlying dioritic layer without any indication of tapering or mixing in the gabbro. We cannot tell from field relationships whether they amalgamate to form the overlying dioritic layer, breach the top of the gabbro magma to become resorbed into an overlying liquid, or eventually diffuse and mix with the gabbro.

Generally, the diorites form homogeneous units with little variation in texture or structure. Stratigraphically lower units are a biotite-hornblende quartz diorite changing upwards through the section to hornblende quartz monzodiorites. These range in thickness from 7 m to 24 m. At contacts where dioritic units overlie the gabbros, several types of relationships occur. At stratigraphically higher levels, the contacts are somewhat diffuse and wispy over a restricted area. As opposed to gabbro overlying diorites, no penetration of one unit into the other is found and the gabbros and quartz monzodiorites do not appear to interact away from the contact. With lower units, however, the contacts can be obscured by physical mixing and hybridization. In the case of the dioritic unit D and the underlying gabbro C, the contact between the units is an approximation and some samples assigned to unit D show more mafic and hybrid compositional characteristics. Figure 22 presents an idealized summary of the different relationships observed at the outcrop.

The conclusions to be drawn from these field and textural relationships are inescapable. The gabbros and underlying diorites represent coexisting, compositionally distinct, layered magmas. The fingers or pipes provide compelling evidence that the diorite, or some component of the diorite, was liquid and intruded into molten gabbros. The chilled margins of the gabbros reinforce this conclusion and indicate that the temperature contrast between the magmas effected the crystallization of the gabbro against the diorite. It is also evident that the magmas were gravitationally unstable and were attempting to overturn. Moreover, during crystallization of the gabbro, a

downward necking of gabbroic fingers or lobes into the underlying diorite eventually detached as sinking pillows. Owing to the more viscous nature of the diorite, the pillows were unable to descend farther than half a meter or so.

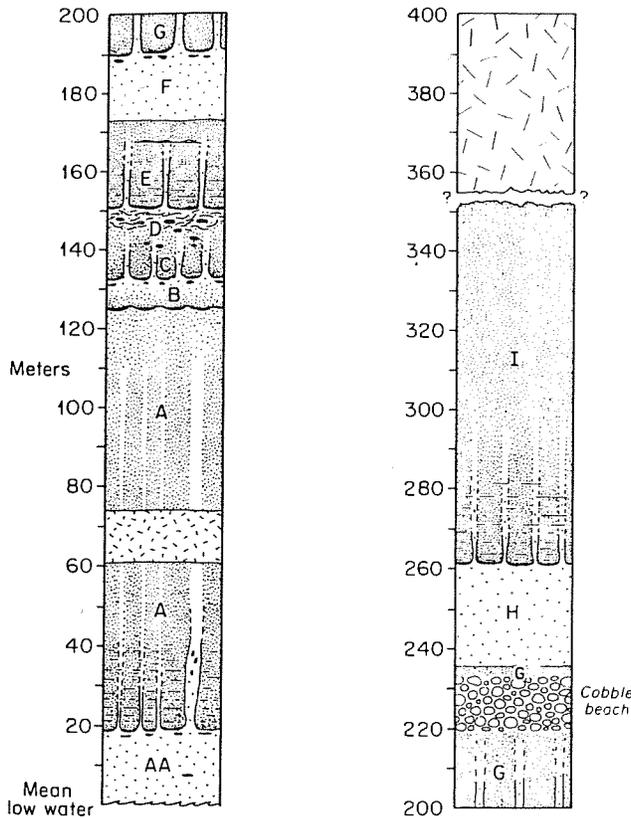


Figure 20. Schematic stratigraphic column across the southern coast of Eastern Head showing the exposed relationships of gabbroic (dark stipples) and dioritic (light stipples) layers and their thicknesses. An aplite sill at 60 m and the Isle au Haut Granite at 358 m are shown by fine and coarse hash marks respectively. All contacts are clearly exposed except where indicated.

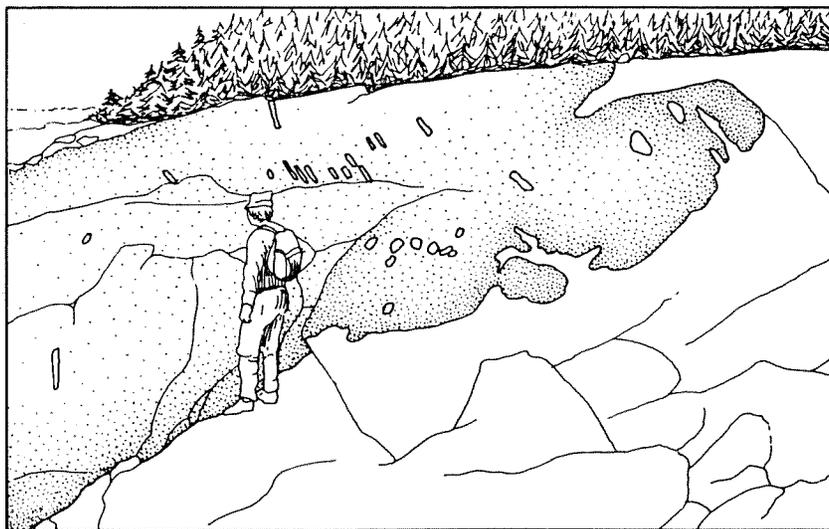


Figure 21. Sketch from a photograph of the contact between gabbro I overlying quartz monzodiorite H. Contact shows cusped "load cast" structures and pipes intruding into the gabbro.

**Petrography**

**Gabbros.** Rocks from the gabbroic units show little modal variation throughout the stratigraphic succession. All five layers are hornblende gabbros from the lowest layer through the uppermost layer (Fig. 23). They contain between 40 and 50 modal plagioclase which is compositionally zoned from An<sub>62</sub> cores to An<sub>35</sub> rims. Plagioclase

crystals near diorite contacts are clear laths from core to rim. Away from these contacts the crystals tend to be lightly to moderately corroded, or contain resorbed cores. The nature of these crystals, obviously in disequilibrium with their surroundings, may reflect exchange of volatiles or other components of the dioritic units with the gabbros. The rims are almost everywhere clear. Except for one sample from the lowest unit (A), none of the gabbroic units contains alkali feldspar or quartz.

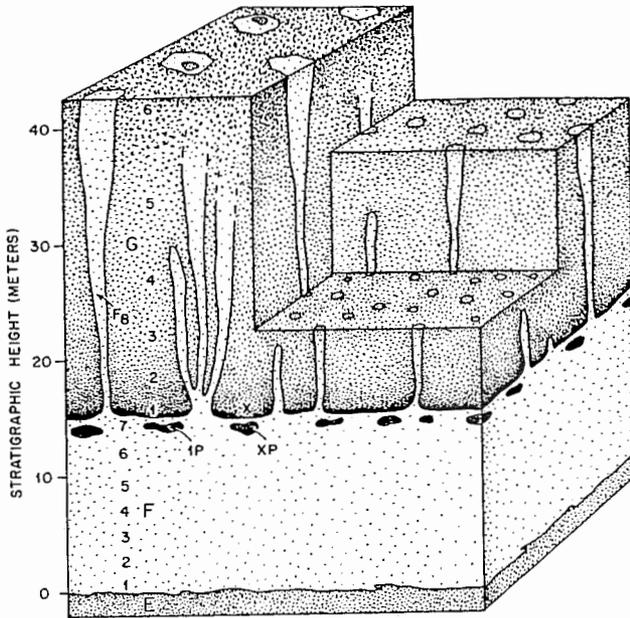


Figure 22. Block diagram showing schematic contact relationships between gabbros E and G and quartz monzodiorite F. This relationship is consistent for all contacts with the exceptions of diorite B overlying gabbro A and diorite D over gabbro C (which shows complex mixing and intermingling, as shown in Fig. 3). Sample location indicated by numbering.

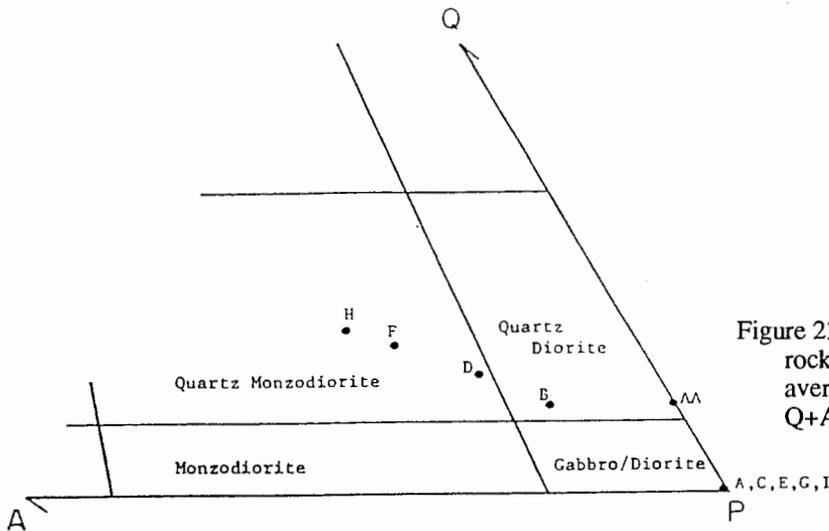


Figure 23. IUGS classification of igneous rocks (after Streckeisen, 1976) and the average modal analysis for each layer.  $Q+A+P = 100$ .

The predominant mafic mineral is a subhedral to anhedral pargasitic hornblende (nomenclature according to Leake, 1978), poikilitic with plagioclase, which ranges from 17 to 54 modal percent. The cores of many of the hornblendes are strongly altered and may indicate relict clinopyroxene. Biotite and titanomagnetite commonly rim the hornblende and occasionally are intergrown with it. Accessory minerals consist almost exclusively of apatite, but zircon and sphene are present, albeit infrequently, as very late-stage phases. Epidote and chlorite (which may be quite abundant) are sporadically present as alteration products.

At the lowermost contacts, the gabbros are fine-grained and have intergranular to intersertal textures. Away from the contacts, the texture changes from subophitic to ophitic and can best be described as coarse cumulate. A few of the units show fine rhythmic layering and one layer (A) shows a local collapse of obvious layering.

Compositionally, the gabbroic units are tholeiitic, subalkalic with little chemical variation. They resemble within-plate tholeiites using the classifications of Pearce and Cann (1973) and Meschede (1986) and are restricted to

the calc-alkaline series of Peccerillo and Taylor (1976) as shown in Figure 24. Plotted on an AFM diagram (Fig. 25), the gabbros are largely uniform in composition. Departures from the group can be ascribed to very local physical mixing and hybridization with the diorites. This appears to have been restricted to a thin region between gabbro I and quartz monzodiorite H and within the mixed zone comprising quartz diorite D. The gabbros are relatively uniform in composition for both compatible and incompatible elements within each unit and throughout the stratigraphic succession (Fig. 26).

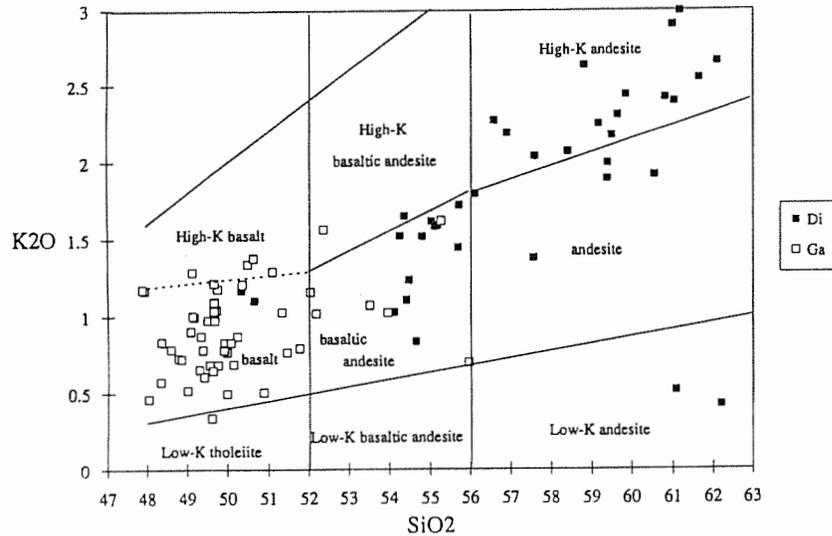


Figure 24. Composition of the gabbros and dioritic units on the  $K_2O$  and  $SiO_2$  diagram; boundaries of arc tholeiite, calc-alkaline, and high-K calc-alkaline series of Peccerillo and Taylor (1976).

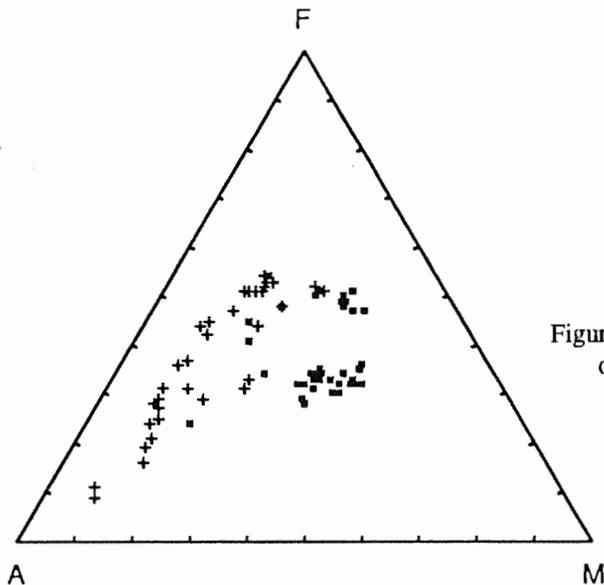


Figure 25. AMF diagram ( $Na_2O + K_2O - FeO^* - MgO$ ) of the gabbros (solid squares) and diorites (crosses).

**Diorites.** In contrast with the gabbros, the dioritic units are texturally homogeneous and show no structural and textural variations among the five layers. All are subophitic. Modally, the layers change upwards from quartz diorites (layers AA and B) to quartz monzodiorites (layers D, F and H) which increase towards higher quartz and alkali feldspar concentrations (Fig. 23).

Plagioclase makes up 44 to 63 modal percent of the dioritic units and shows normal zoning from  $An_{30}$  to  $An_6$ . The grains are not nearly as corroded as those from the gabbros, nor do core to rim variations appear as abrupt changes in composition.

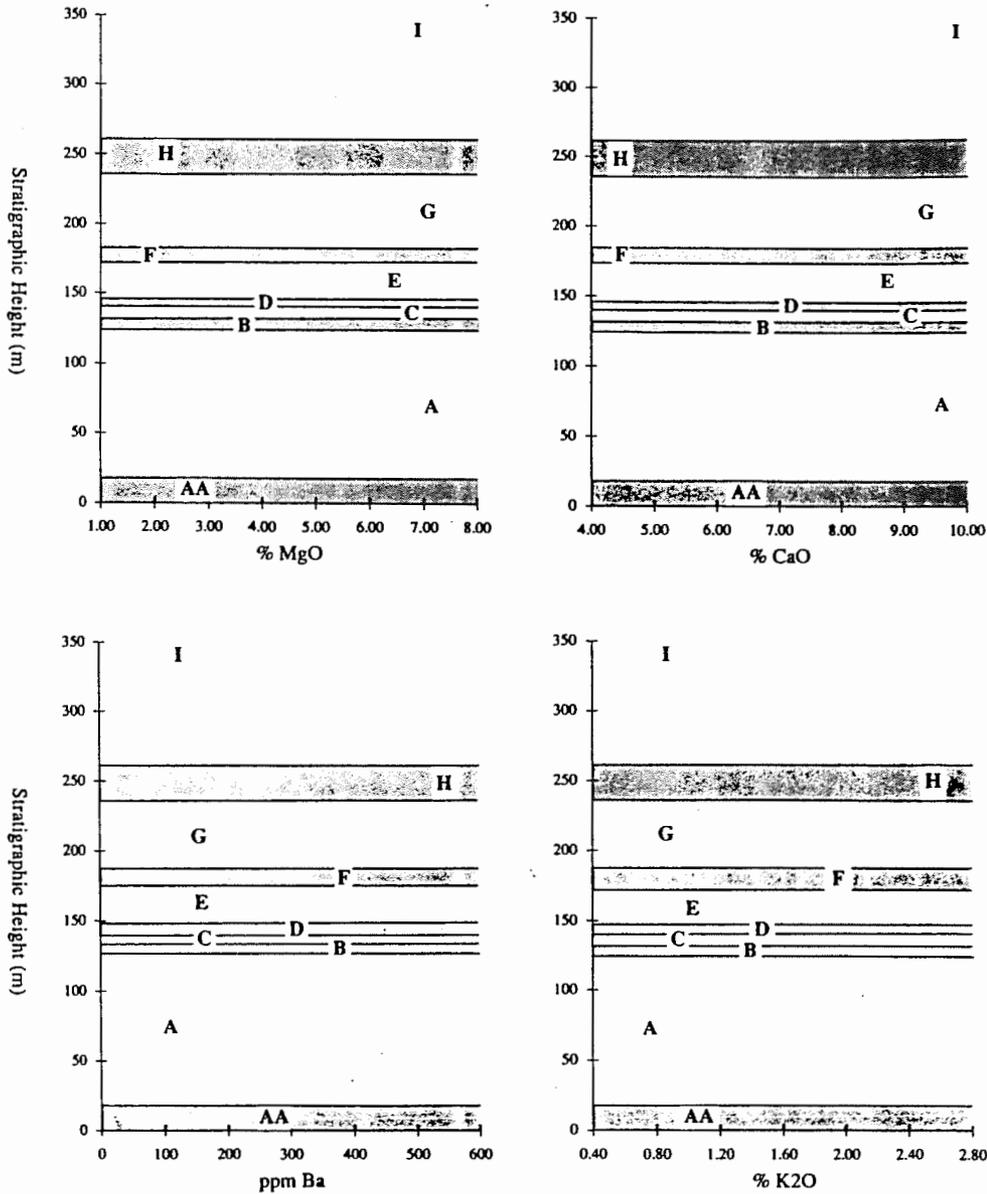


Figure 26. Average incompatible and compatible element compositions (wt. %) versus stratigraphic height for samples in the alternating layered series. Average composition ignore obvious effects of hybridization at some boundaries. Gabbros are layers A, C, E, G, and I; dioritic layers are AA, B, D, F, and H. Boundary between layers shown as solid horizontal line.

Alkali feldspar is interstitial and also present as a very late stage rim to plagioclase, but only in layers B and above. With increasing stratigraphic height, alkali feldspar becomes more abundant and appears to become a possible cumulus phase. Late-stage, interstitial, anhedral quartz increases in amount with stratigraphic height, though not as dramatically as alkali feldspar.

The dominant mafic phase is again hornblende, but it is edenitic in composition. Biotite is commonly intergrown with (or replaces) the hornblende, rims it, and also forms an interstitial member. Titanomagnetite rims both these mafic phases. There is also actinolite which may be the alteration product of pyroxene.

Accessory minerals include late-stage sphene, zircon, rutile, and substantial apatite in some samples. Epidote and iron oxide staining are present as secondary minerals.

At Eastern Head, the dioritic units show no obvious foliation, but are generally massive. This massive texture is in stark contrast to the diorite which makes up the northern two-thirds of the island and has a strong plagioclase foliation. Finally, xenoliths are rare; however, two dioritic units (AA and F) have one isolated and heavily resorbed xenolith apiece.

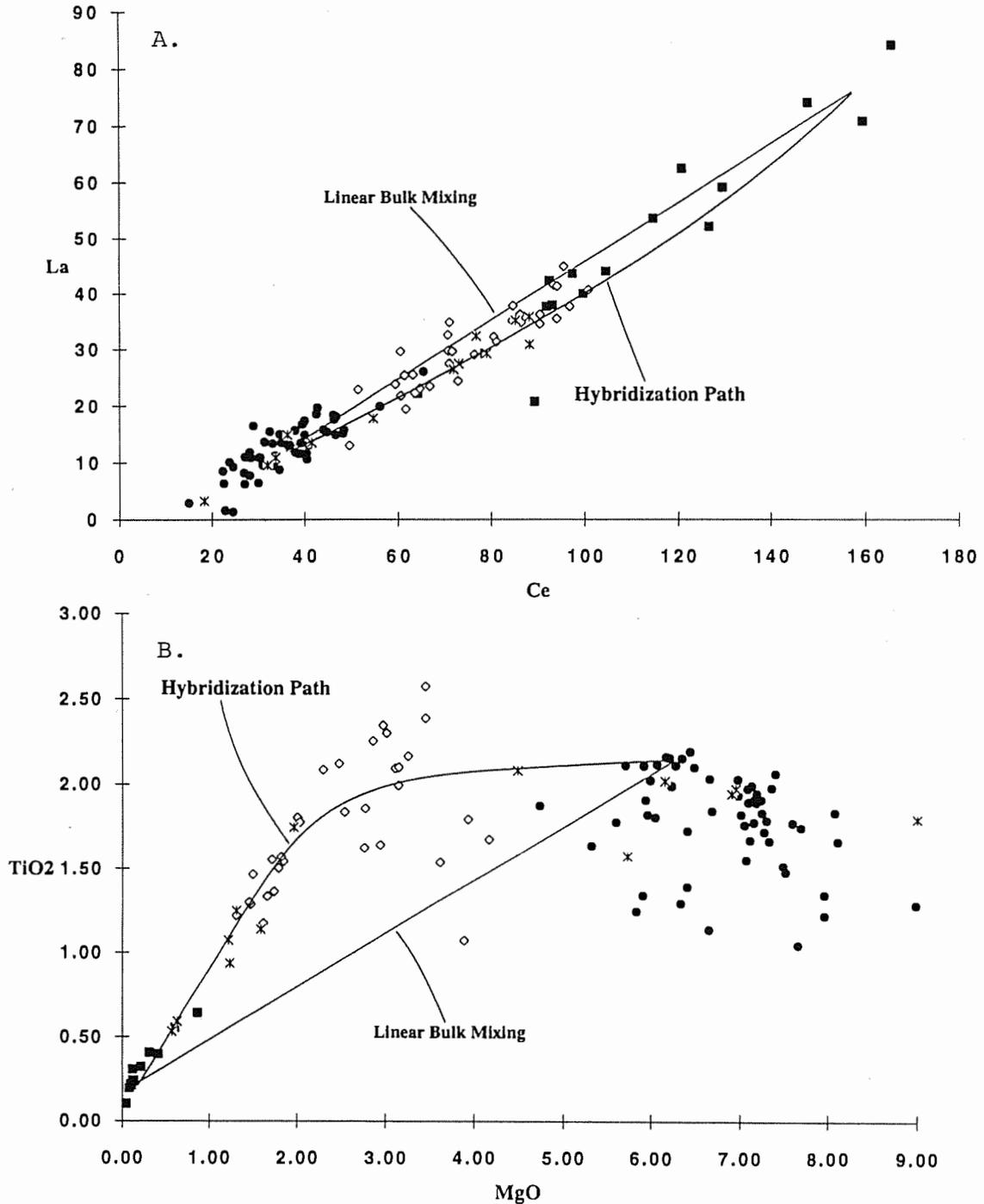


Figure 27. Compositional variations of granites (squares), gabbros (solid circles), diorites (open diamonds), and hybrids (crosses) with respect to: (A) La and Ce, showing similar diffusivities, and (B) MgO and TiO<sub>2</sub>, showing different diffusivities. Linear mixing trend is calculated from average granite to chilled gabbro compositions. Hybridization path is drawn to fit hybrid compositions.

Chemically, the dioritic layers are andesitic with an evolving calc-alkaline trend (Fig. 24). Incompatible elements increase in abundance, whereas compatible elements decrease, with stratigraphic height, in accordance with the change from quartz diorite proper to quartz monzodiorite (Fig. 26). There are also strong variations within each unit which are sympathetic to the overall stratigraphic trend. The pipes differ from their associated diorites in composition. They are strongly depleted in many elements that are abundant in the diorites (e.g.  $K_2O$ , Zr), and they have syenitic affinities. Occasional pipes terminate a short distance into the overlying gabbros. These concentrate actinolite and epidote at their terminus.

### On the Origin of the Diorites

Four possible models may be entertained for the origin of the dioritic units: (1) cumulates from the overlying Isle au Haut granite; (2) fractionation products of the underlying gabbros, with an assimilated granite component (AFC); (3) cumulates from a dioritic magma, with no genetic connection with either the gabbros or granites; or, (4) the product of mixing between the granite and gabbro. The geochemical and mineralogical data do not support the first two models. Using appropriate distribution coefficients for the phases which make up the diorites, neither fractionation of the gabbro, nor cumulate trends from the granite, produce the compositions of the dioritic units. The third model is plausible, but less satisfying. This leaves the possibility that the diorites are the products of a mixing event between the granite and gabbro in a stratified silicic magma reservoir as the most plausible solution.

Mixing has often been proposed to explain the interactions of bimodal magmas. In the plutonic environment, where commingling is common, linear arrays of major elements have been used to support mixing models between two endmembers. Typically, the major elements used to constrain the data are the network forming members (e.g.,  $SiO_2$ ,  $MgO$ ,  $Fe_2O_3$ ,  $CaO$ , etc.). Trace element data, however, are frequently scattered and non-linear. Recently, Leshner (1990) explored the possibility that mixing may not produce linear trends. Instead, due to differing diffusivities among the elements, non-linear hybridization trends tend to be favored over a linear bulk mixing trend. The dioritic units on Eastern Head are consistent with this interpretation.

In order to test this hypothesis, we have collected and analyzed hybrid samples from intimately mixed granite and layered gabbro to the east of the island. From elements with similar diffusivities (e.g., La and Ce), the hybridization trend closely matches a linear bulk mixing line (Fig. 27A). On the other hand, elements whose diffusivities are very dissimilar (e.g.,  $TiO_2$  and  $MgO$ ) show a marked bowing in the hybridization path away from a linear bulk mixing trend (Fig. 27B). Data from the dioritic units conform more closely with these curved hybridization paths, and not with a linear bulk mixing line. Therefore, we propose that the diorites are the cumulate product of a hybrid magma which formed from a mixing event between the granite and the gabbro.

### CONCLUSIONS

We believe that mafic-silicic layered intrusions (MASLI) like those visited on this field trip are extremely common and occur in a wide range of tectonic settings (Wiebe, 1993a, b). They represent the plutonic record of complex magma chambers comparable to those inferred to exist beneath many long lasting silicic volcanic systems. They provide compelling evidence for (1) multiple infusions of mafic magma into floored chambers of relatively silicic magma and (2) the existence of compositionally stratified magma chambers. Some MASLI provide a stratigraphic record of interactions between mafic and silicic magmas along double-diffusive boundaries. MASLI have great potential to provide new insights into magma chamber processes and into connections between the plutonic record and volcanic activity.

### ACKNOWLEDGEMENTS

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## ROAD LOG

## DAY 1 (See Fig. 2 for locations.)

**NOTE:** Stops 1, 3, and 4 (as well as Stop 7 on Day 2) require walking across private property in order to gain access to the shore. While in the past, landowners have readily permitted visits, access cannot be guaranteed at all times. It would be best to obtain permission from landowners when they are present.

**Mileage**

- 0.0 Road log begins at the western intersection of U.S. Route 1 with Maine Route 187 (about 1.2 miles east of Columbia Falls). Proceed south on Route 187.
- 6.5 Turn right on to paved road toward South Addison.
- 9.9 At point where paved road makes a sharp right turn, go straight ahead on to a gravel road. (There is a house with a satellite dish on the left.)
- 10.2 Road branches; take the left branch.
- 10.3 Road branches; take the left branch.
- 10.4 Continue straight ahead.
- 10.6 Park to the right of the drive. House on the left is owned by Mr. Don O'Brien. Path to shore is straight ahead. Upon reaching the beach we will take a traverse of roughly 250 m to the west along the shore.

**STOP 1. BICKFORD POINT SECTION** (See Fig. 3). Stop 1 provides a section through several macrorhythmic units. The lowest unit (1) grades upward from olivine gabbro to diorite and locally displays textural and modal layering. Units 2-5 represent small infusions of basaltic magma. The base of unit 6 is defined by strong chilling and well developed pipes structures. Preliminary Sr-isotope data indicate a systematic upward increase in initial Sr-isotope ratios in unit 1 from about 0.7030 at the base to 0.7035 at the top (Wiebe and Sinha, in prep.).

- 10.6 Return to vehicle and drive back to paved road.
- 11.2 Turn left (west) onto paved road.
- 13.7 Turn left onto paved road to South Addison.
- 15.3 Continue straight ahead (rather than taking the main road that turns to the right).
- 15.4 Pavement ends; continue on gravel.
- 15.6 Turn left (to Bar Island).
- 15.8 Turn left again (to Bar Island).
- 16.2 Stop and park in area to right. Walk down to shore where we will make a traverse of roughly 750 meters to the north along the shore.

**STOP 2. BAR ISLAND SECTION** (See Figs. 3 and 7). Stop 2 provides another section through several macrorhythmic units of a different character. The lower part of the section is characterized by many thin units representing small infusions of basaltic magma. Many small scale structures illustrate the gravitational instability of silicic material trapped beneath basaltic infusions. Both units E and F contain substantial thicknesses of highly silicic cumulates which contain some ghost-like mafic enclaves. Figure 7 illustrates the chemical variation in the upper 20 meters of unit F. Preliminary Sr-isotope data indicate a systematic upward increase in initial Sr-isotope ratios in the upper 10 meters of unit F from about 0.7035 to 0.7038 (Wiebe and Sinha, in prep.).

- 16.2 Return back towards South Addison.
- 16.6 Turn right at junction.
- 16.8 Turn right at junction.
- 17.1 Stop sign. Continue straight ahead.
- 18.7 At this T-junction, go left (north).
- 24.7 Continue across bridge into Addison.
- 25.3 In Addison, turn left at the Addison Post Office and cross bridge.
- 25.7 Take the left branch of this road that continues south along the west side of the Pleasant River.
- 26.0 Gravel surface begins.
- 27.6 Continue past llama farm (!) on left.
- 28.0 Road branches; keep to the left.
- 29.4 Road branches; keep to the left.
- 30.8 Road branches; keep to the left.

- 31.1 Stop and park along the road where possible. Access to the shore is a short walk across property owned by Mr. Mike Plummer. We will walk west and north along the shore for about 200 meters..

**STOP 3. GUARD POINT** The purpose of this stop is to examine excellent coastal exposures of large load cast structures of chilled gabbroic material that rest on the most evolved silicic cumulates in the Pleasant Bay intrusion. The silicic material has coarse subhedral alkali-feldspar and abundant interstitial quartz. Many prominent silicic pipes in gabbro occur in the northern half of this traverse; the upper portions of many pipes are pegmatitic and vuggy with epidote and low quartz.

- 31.1 Turn around and retrace route back to the north.  
 31.4 Road branches; keep to the right.  
 32.7 Take the left branch.  
 33.1 At paved road, turn left (south).  
 34.8 Gravel surface begins.  
 37.1 Road branches, keep to the right.  
 37.4 Stop along the road before passing the "Starkey" mail box on the right. There is a very short walk straight ahead to the shore. We will walk around the peninsula to the left (south) - about 600 meters.

**STOP 4. RIPLEY** This traverse displays many thin macrorhythmic units due to basaltic infusions. Pipe structures are locally well developed. These cumulate units are abundantly cut and locally rotated by many composite dikes consisting of chilled basaltic pillows in fine-grained leucocratic granite.

- 37.4 End of Day 1. Return to vehicle and return to U. S. Route 1 (at mile 46.1)

**DAY 2** (See Fig. 8 for locations.)

**NOTE:** Stops 1 through 5 are located within the Acadia National Park. Although sampling is normally prohibited, it is possible that we will obtain special permission for this trip for sampling at some of these stops. Sampling is permitted at stops 6 and 7.

#### Mileage

- 0.0 Road log begins at the far end of the parking lot at the summit of Cadillac Mountain.

**STOP 1. SUMMIT OF CADILLAC MOUNTAIN.** This locality provides an overview of Mount Desert Island and an initial look at the Cadillac Mountain granite.

- 0.0 Begin driving back down Cadillac Mountain.  
 0.4 Turn right, into the Blue Hill overlook and park in the parking lot.

**STOP 2. BLUE HILL OVERLOOK.** This locality provides a view of the western side of the island. The CMG at this locality contains many silicic "blobs" that range in composition from 70 to 78% SiO<sub>2</sub>.

- 0.5 Leave parking lot and continue driving down Cadillac Mountain.  
 2.0 Park cars on the right side of the road uphill from a hairpin turn.

**STOP 3. LOOP ON CADILLAC MOUNTAIN ROAD.** Road-cuts here provide excellent exposures of the granite and a wide variety of mafic enclaves. The granite is cut by a N-S trending basaltic dike.

- 2.0 Continue on down the mountain to the loop road.  
 3.7 Turn right onto loop road.  
 4.2 Take the right fork, continuing on the loop road.  
 8.3 Pull off to overlook on left across the road from a long road-cut.

**STOP 4. INTRUSIVE BRECCIA ON LOOP ROAD.** The breccia includes large blocks of the Bar Harbor Fm, as well as gabbroic and some metavolcanic rocks. There are excellent exposures of the basal, more mafic and plagioclase-rich phase of the CMG. These rocks grade rapidly to more characteristic CMG in road-cuts to the south.

- 8.3 Continue on loop road.

- 9.8 Pass tollgate.
- 12.3 Pull off into parking lot on right.

**STOP 5. INTRUSIVE BRECCIA AT OTTER POINT.** Large blocks of mafic and silicic metavolcanic rocks occur in the basal CMG. Mafic minerals in the granite include augite, hornblende, fayalite, and biotite.

- 12.4 Continue on the "Loop road" past Jordan Pond.
- 22.8 Continue on Loop road past the entrance to Cadillac Mountain summit road.
- 23.3 Take the left fork off of the Loop road.
- 23.8 Turn right onto entrance road to Route 233.
- 24.0 Intersection with Route 233. Turn right (west).
- 28.8 Pull off onto the side of the road just before the junction of Route 233 with Route 3.

**STOP 6. SOMESVILLE GRANITE.** These fresh road cuts are typical of the coarser-grained variety of the Somesville granite.

- 28.9 Turn right onto Routes 3 and 198.
- 30.3 Turn left (south) at T-intersection onto Route 102. Continue through the town of Somesville.
- 31.2 Turn right on road toward Pretty Marsh.
- 34.9 Continue straight ahead on road toward Pretty Marsh Picnic Area.
- 35.4 Continue straight past Pretty Marsh.
- 36.5 Turn right onto Cape Road.
- 36.6 Keep left on Cape Road (paved).
- 37.2 Continue on gravel road.
- 37.8 Turn right onto dirt road.
- 38.0 Take left fork (Cape Farm) and continue straight ahead.
- 38.3 Park in open grassy area away from house.

**STOP 7. STEWART HEAD.** We will walk down to the shore and take a traverse along the south and west sides of Stewart Head, returning in a loop back to the vehicles. The coastline here provides many fine examples of field relations within the gabbro-diorite unit. The very first exposures we encounter are of a large composite dike consisting of basaltic pillows in a matrix of leucocratic granite. This is one of two composite dikes studied by Chapman (1962b). Beyond these outcrops we will see a wide variety of mafic to silicic layers that have mainly formed by periodic infusions of basaltic magmas into the floor of the CMG magma chamber. A sketch map of part of this section is shown in Figure 9. This is the last stop of Day 2.

### DAY 3

**NOTE:** Transportation is not readily available on Isle au Haut, unless special arrangements are made in advance with a resident of the island. The best exposures on Isle au Haut are located on Eastern Head, which is approximately 7 miles by road (the only paved road) to the southeasternmost tip of the island. The Mail Boat schedules is strictly adhered to. Missing the boat may entail arranging for a special boat with an additional cost (~\$70). Accommodations and restaurants are also not available.

Assemble at Atlantic Avenue Hardware and the Isle au Haut wharf in Stonington, Maine at 0745 hours. The special boat for Isle au Haut will leave from Stonington promptly at 0800 hours. The boat arrives at Isle au Haut town landing at approximately 40 minutes after departing from Stonington. After arriving Isle au Haut, transportation will be provided to the southern end of the island.

**Mileage** (There are very few cars on the island with operating odometers!)

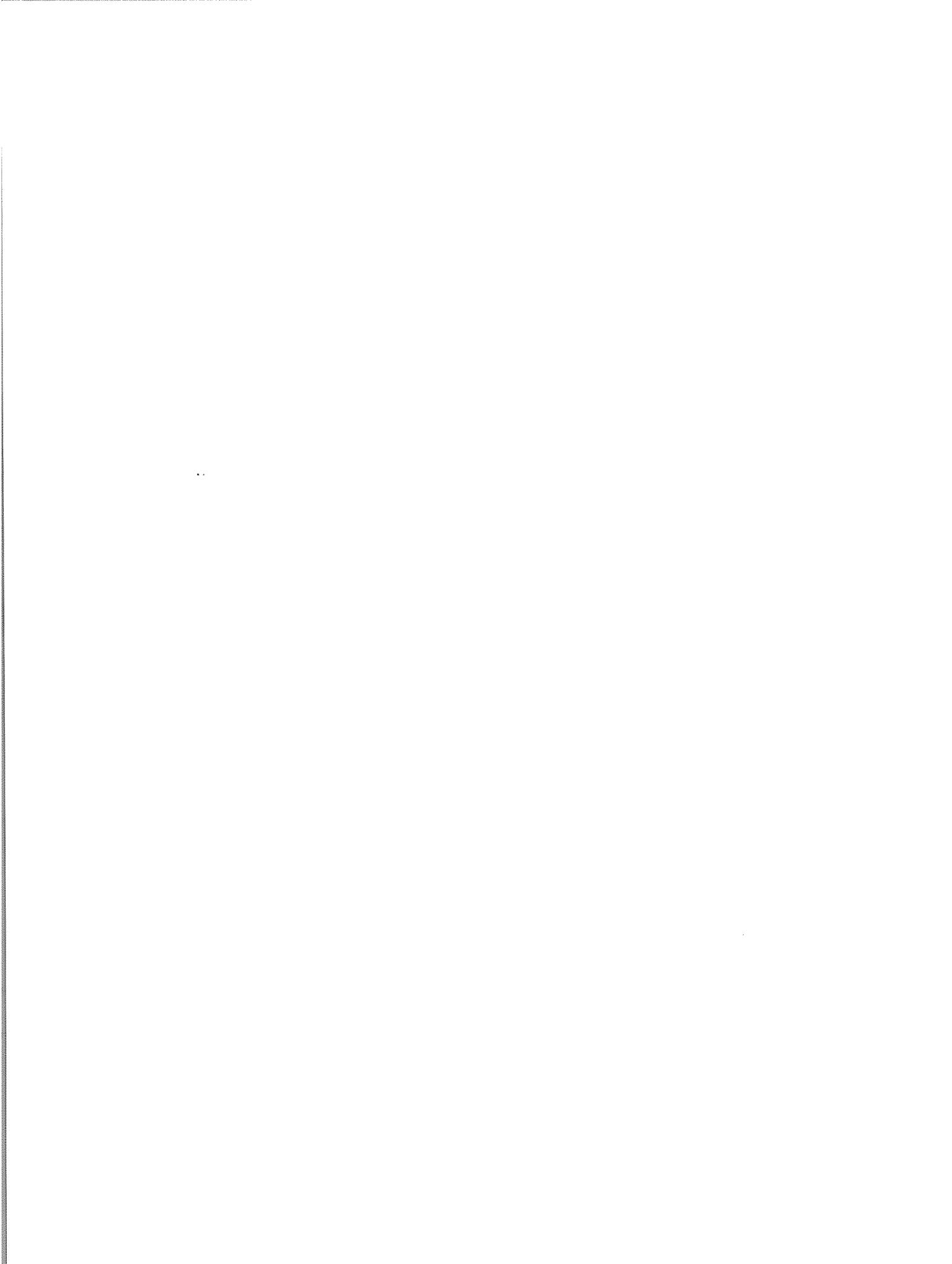
- 0.0 From the Isle au Haut Town Landing, take the paved road along the eastern side of the island to Head Harbor, the southernmost tip of the island. Paved road changes to dirt road for approximately one mile and returns to paved surface.
- ~6.0 Turn left onto dirt road past large yellow house (left). Road continues behind large red house and east of the Head Harbor. Park immediately south of dilapidated barn where road ends (more or less).
- ~6.3 Park trail (NO SAMPLING) to Eastern Head begins approximately 100 yards south of this spot, past a cabin and house on left. Discarded Model-T Fords are on the left.

**STOP 1. COMPOSITE LAYERING OF DIORITIC AND GABBROIC UNITS ON EASTERN HEAD (3 HOURS) (Isle au Haut Quadrangle, east).** The purpose of Stop 1 is to examine a series of alternating gabbros and quartz diorites to quartz monzodiorites which we believe to be five repeated intrusions of a basaltic magma onto the floor of an evolving silicic reservoir.

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Chapter B

# Sequence and Correlation of Tectonic and Metamorphic Events in Southeastern Vermont

By James B. Thompson, Jr., John L. Rosenfeld, and  
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## SEQUENCE AND CORRELATION OF TECTONIC AND METAMORPHIC EVENTS IN SOUTHEASTERN VERMONT

by

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### INTRODUCTION

The purpose of this trip is to introduce you to the intricately deformed metamorphic terranes of southeastern Vermont. Two of us (JBT & JLR) have been working here, for at least a part of most field seasons, for more than forty years, the third (CPC), and his students at Dartmouth College, are now under way in the exploration of further facets of this fascinating region. We shall present to you our current interpretation of its stratigraphic, structural, and metamorphic history. As in most complex terranes that are easily accessible, divergent opinions and interpretations have indeed arisen. As an example of a wholly independent investigation that has taken strong issue with us on many, if not most, fundamental aspects of our synthesis, see Ratcliffe and others (1992). Perhaps, when these, our most severe critics, have seen a bit more of the available outcrop, they may be less convinced of our error, -- but that is their problem, not ours! We shall unabashedly present things here as we now see them, and shall take you to as much of the key evidence as we can within the constraints of a three day excursion.

### GEOLOGIC SETTING

The areas to be visited during this excursion all lie within the Glens Falls two degree sheet (Thompson and others, 1990), the geologic map of Vermont (Doll and others, 1961), and also within the area of Continent-Ocean Transect E-1 (Thompson and others, 1993). All of the exposures to be visited lie between the Green Mountains, on the west, and the Connecticut River, on the east. Our first and most easterly stop will be a few meters from the bank of the Connecticut River, and our most westerly ones will be on the eastern slope of the Green Mountains. The Chester and Athens (pronounced Ay-thens) gneiss domes lie midway between, and are centered on the most westerly of the strong Paleozoic metamorphic highs of the New England Appalachians.

The Green Mountains of Vermont and the Berkshires of western Massachusetts, constitute a segment of the "Blue-Green-Long axis" of Rankin (1976), a series of external massifs that play a role in the Appalachians that is closely analogous to that of the Aar-Gotthard, Mont Blanc-Aiguilles Rouge and other external massifs in the western Alps (see Thompson, 1990, especially Figs. 1 and 2 for same-scale maps and further comparisons of the two regions). The Precambrian basement exposed in the core of the Green Mountains has been strongly modified by Paleozoic deformation and by the superposed effects of a relatively low-grade Paleozoic metamorphism. Where the later effects are least severe it is clear that the rocks bear a marked resemblance to those in the granulite-facies Grenville Terrane (ca. 1,100 Ma) that can be seen in a more pristine condition in the Adirondack Mountains of northeastern New York State. The rocks in the cores of the Chester and Athens domes are also believed to be in large part, Grenville correlatives, but much more strongly modified by later events than those in the Green Mountains.

The region between the Green Mountains and the Connecticut River, the focus of this excursion, has many features in common with the portion of the Pennine Alps lying between the Aar Massif and the Insubric Line. The geologic maps of both regions are locally characterized by intricate "marble cake" patterns suggesting poly-phase deformation of highly ductile materials that must have behaved almost like viscous liquids. Both regions have been interpreted as characterized by fold nappes, the lower ones basement-cored, and the whole modified by later arching or doming. Both regions also show evidence of a back flow or *retrocharriage* off the external massifs during or after doming. At the deepest tectonic levels (in the Chester and Athens domes in Vermont, and in the lower parts of Val d'Ossola and Val Leventina in the Alps) the cover rocks extend deep into the basement in nearly isoclinal fold-septa, the "mulden" of the Alpine geologists. The Alp-New England analogies can, in fact, be extended further to include the pre-metamorphic lithofacies of the cover rocks, and even to the metamorphic mineral assemblages and textures developed therein.

In Vermont the cover rocks in the eastern Green Mountains and in the Chester and Athens domes constitute what is known as the Eastern Vermont Sequence. Although non-fossiliferous in the area of this excursion there are sparse fossil localities in northern Vermont and Quebec (Konig and Dennis, 1964; Hueber and others, 1990) showing that it is, at least in part, mid-Paleozoic. The younger rocks in the Eastern Vermont Sequence are of Silurian to perhaps Early Devonian age and are separated from the older ones by a boundary known informally as the Richardson Memorial Contact (RMC) after C. H. Richardson, a geologist who worked in eastern Vermont in the early part of this century. The RMC is regarded by us as here an eastward-facing unconformity, though probably complicated by faulting farther north (Westerman, 1987). The rocks above the RMC constitute the outermost cover of the Chester and Athens domes. The structural and metamorphic features in them are necessarily Acadian or later. Some of the lower units, below the RMC, in the Eastern Vermont Sequence may be as old as Late Proterozoic, but the bulk of the lower section is believed to be Cambrian through mid-Ordovician.

The Bronson Hill terrane, just to the east in western New Hampshire also has its own distinctive suite of cover rocks. These constitute the New Hampshire Sequence, ranging in age from mid-Ordovician through Early Devonian. Biostratigraphic control is considerably tighter in the New Hampshire Sequence than it is in the Eastern Vermont Sequence. We shall see New Hampshire Sequence rocks only at the first two stops of this excursion, one of them a fossil locality.

The boundary between the above two sequences corresponds roughly to the course of the Connecticut River, and is here known informally as the Chicken Yard Line (CYL) from a once aromatic locality in Dummerston, Vermont, just north of Brattleboro, a few kilometers south of the southeast corner of Plate I. We believe that the CYL is an eastward-facing unconformity, extensively offset by later folding and faulting, but others have regarded it as a westward-verging thrust fault.

The following discussion is necessarily brief. Readers interested in a broader overview may find some of what they want in the text accompanying Continent-Ocean Transect E-1 (Thompson and others, 1993), also Thompson (1990), or, for a dissident view, Stanley and Ratcliffe (1985). Other references deal with a finer scale. For still further detail we have tried to select outcrops that can speak for themselves!

## PLAN OF THE EXCURSION

### Friday

Our first stop, on Friday, will be at a fossil locality in the New Hampshire sequence and the second stop will be at the CYL. The rocks at and near the CYL are at the lowest metamorphic grades to be encountered in this excursion. Pelitic rocks in the New Hampshire Sequence adjacent to the CYL have been quarried locally for roofing slate.

The remainder of Friday's excursion will be devoted to the rock units and to structural and metamorphic features that can be seen between the CYL and the RMC, and in the immediate vicinity of the latter. Proceeding westward, from the CYL toward the RMC and the Chester dome, we shall encounter a progressive rise in metamorphic grade. (Pelitic rocks to be seen on Sunday in the deeper mulde in the Chester and Athens domes include coarsely crystalline staurolite-kyanite-garnet schists.) Special features of the day will include large snowball garnets and the transformation of low-grade greenstones into coarsely crystalline hornblende-garnet gneiss.

### Saturday

On Saturday we shall concentrate on the older rocks of the Eastern Vermont Sequence as displayed between the basement-cover contact on the eastern flank of the Green Mountains and the RMC, mainly in the northwestern corner of the map (Plate I), in the towns of Bridgewater, Plymouth and Ludlow. Metamorphic grade here will *decrease* as we proceed westward, down-section, toward the Green Mountains. Special features of the day will include the basement-cover unconformity, highly strained, polymictic conglomerates at the base of the cover sequence, and an alpine-type ultramafic.

### Sunday

Sunday will be devoted to the mantles and cores of the Athens and Chester domes. The first stop will be an extended one to take advantage of the excellent highway and spillway exposures in the mantle of the Athens dome.

These exposures date from the construction of the Townshend Dam on the West River, near the southwest corner of the map (Plate I), and display a condensed version of the stratigraphic sequence seen on Saturday. Special features will include small ultramafic pods, more snowball garnets (including the first to have had their growth rate measured), and the migration of fluid species during metamorphism as revealed by stable isotope studies.

The rest of the day, the last of this excursion, will be spent in the domes. The last stops of the day will be to see the schists and associated marbles and calc-silicate rocks in the lowermost septa of cover rock in the Chester dome. Special features will include augen gneisses, still more snowball garnets, and epitaxial intergrowths of kyanite and staurolite.

## DESCRIPTIONS OF MAP UNITS

### White Mountain Plutonic-Volcanic Series

Mount Ascutney is the dominant topographic feature of the Connecticut Valley between Springfield, Vermont and Hanover, New Hampshire, and, though not to be visited, will be visible, weather permitting, on each day of the excursion. It is part of a Mesozoic (ca. 120 Ma) magmatic complex that cuts across the northeastern margin of the Chester dome (Daly, 1903; Chapman and Chapman, 1940; Foland and Faul, 1977). The eastern part (Ascutney proper) is mainly syenite with minor granite and felsite. The western part (Little Ascutney) is mainly gabbro or diorite but includes a remarkable breccia (Schneiderman, 1989) containing blocks apparently derived from relatively low-grade (though now hornfelsed) metasediments that covered the still-buried Chester and Athens domes during the early Cretaceous. Unmetamorphosed mafic dikes, to be seen at several of our stops, are probably all Mesozoic

### Other Intrusive Rocks

Small bodies of granitic rock (trondhjemite to granite) cut all of the metasedimentary units in eastern Vermont. Some, at least, are therefore Acadian or younger. Most are too small to show on Plate I, but larger bodies occur both to the north (Barre Granite) and to the south. One, east-northeast of Gassetts, contains exotic fragments apparently derived from a nearby mulde. Others (at Proctorsville and Newfane) contain flattened micaceous orbicules. Some are foliated, and some are not.

Ultramafic rocks are found in most of the cover units beneath the RMC, but not above. The smaller bodies are either serpentinite with a rim of talc-carbonate rock or wholly talc-carbonate. Larger bodies contain cores of relatively unaltered harzburgite or dunite. All are presumably pre-Taconian, some, deep in the Chester dome, but not shown on Plate I, may be Precambrian. Certain thin occurrences that can be traced for a mile or more, in areas of amphibolite or greenstone, may be metamorphosed komatiites. Others may have been intrusives of various shapes and sizes into ocean floor sediments and basalts.

### New Hampshire Sequence

The units of the New Hampshire Sequence within or near the area of Plate I are (in descending order) the Littleton, Fitch, Clough and Partridge Formations. These have not been differentiated on Plate I, except for a dashed line surrounding the area of Clough that includes Stop 1. The Littleton Formation, the most extensive, is here a gray, non-calcareous slate or phyllite, locally with thin sandy beds that may show grading and tops. Along the CYL the Littleton contains quartzite, conglomerate, and slate-matrix conglomerate which, in a few occurrences, show channeling and graded beds indicating tops east into the main mass of the Littleton. The Littleton Formation contains Lower Devonian fossils in the vicinity of Littleton New Hampshire, its type locality (Billings, 1937; Boucot and Arndt, 1960).

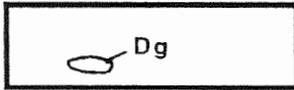
The Fitch, Clough, and Partridge Formations crop out east of the CYL with which they do not come in contact, but rather occupy the western hinge regions of two major nappes (the Bernardston and Skitchewaug nappes) that root 15-20 km to the east on the crest and eastern flank of, a series of gneiss domes constituting the the Bronson Hill anticlinorium. The nappes are essentially fold nappes, but with extreme attenuation and local thrusting on the inverted limbs. The Fitch Formation, to be seen only in driving past, contains granulite or "pit rock" with distinctive calcareous pits, impure marble, amphibolite, and minor mica schist. Recognizable fossils near Littleton, New Hampshire show that the Fitch is Late Silurian, Ludlow to Pridoli (Harris and others, 1983), extending, at least locally, into the earliest Devonian (Elbert and others, 1988). The Clough Formation has a basal conglomerate, mainly a quartz conglomerate, but locally polymictic, overlain by a thin zone of gray to slightly rusty mica schist,

# Map Units, Plate I

(Descriptions in text. Some implied ages are specified in text, others are estimated.)



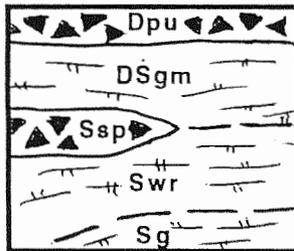
White Mountain Plutonic-Volcanic Series



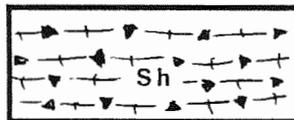
Granitic intrusives



New Hampshire Sequence



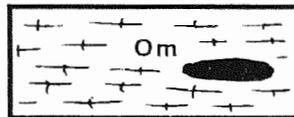
"Califerous Schists" and their associated Volcanics:  
 Dpu Putney Volcanics  
 DSgm, Gile Mountain Formation  
 Ssp, Standing Pond Volcanics  
 Swr, Waits River Formation  
 Sg, Goshen Formation



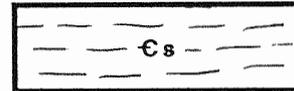
Hawley Formation,  
 quartz conglomerate dotted.



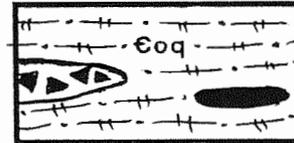
Barnard Gneiss



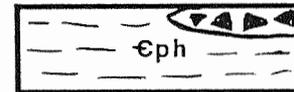
Moretown Formation,  
 ultramafic rocks in solid black.



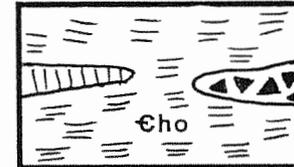
Stowe Formation  
 (Green Mountains only, not mapped around domes)



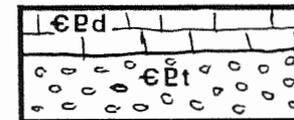
Ottauquechee Formation,  
 amphibolites and greenstones in black triangles, ultramafic rocks in solid black.



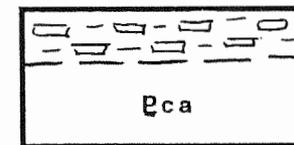
Pinney Hollow Formation,  
 amphibolites and greenstones in black triangles.



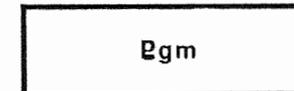
Hoosac Formation,  
 amphibolites and greenstones in black triangles, quartzite and dolomite of Plymouth Member in cross-strike lines.



Tyson Formation, EPt, dolomite member, EPd, in brick wall pattern.



Precambrian basement in Chester and Athens domes. Slabby gneisses, augen gneisses, etc. of marginal zone are patterned.



Precambrian basement in Green Mountains.

# Geology of the Chester & Athens Domes, Vermont

Geology by  
J. B. Thompson & J. L. Rosenfeld,  
1947-1992

Portions modified from maps by Brace (1953),  
Chang & others (1963), Chapman (1952),  
Davidson (1981), Downie (1982),  
Hepburn & others (1984), Karabinos (1984a),  
Kruger (1946), Lyons (1955), and Skehan (1961).

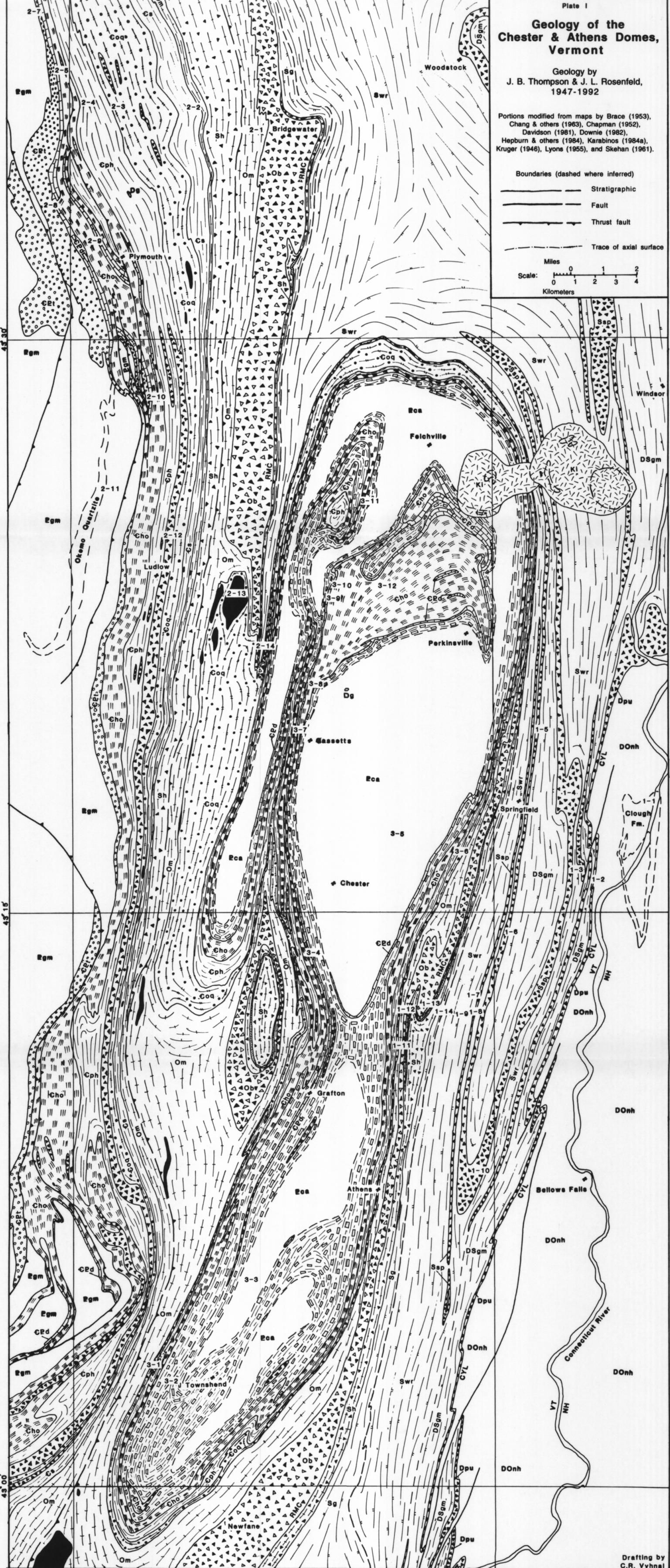
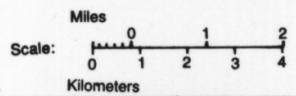
Boundaries (dashed where inferred)

— Stratigraphic

— Fault

— Thrust fault

— Trace of axial surface



and that, in turn by clean quartzite with minor calc-silicate and calcareous zones near the top. Numerous fossil localities in the upper part (as at Stop 1) have Lower Silurian, late Llandovery, faunas (Boucot and Thompson, 1963).

The Partridge Formation, to be seen only in driving by (between Stops 1 and 2), is a gray to black, highly sulfidic phyllite or schist with minor amphibolite. It is probably correlative with mid-Ordovician graptolitic slates in the Rangeley Lakes region of western Maine (Harwood and Berry, 1967). Farther east, in the central Bronson Hill terrane, the Partridge overlies the Ammonoosuc Volcanics and its associated intrusives that are believed to be remnants of a pre-Taconian island arc. For more on the nearby part of the Bronson Hill terrane see Thompson and others (1968; reproduced, with additions and emendations in Robinson and others, 1979). In the Littleton area of New Hampshire the Silurian formations truncate structures in the older units, marking a clear-cut Taconian unconformity. The Littleton Formation also rests on all units older than itself (including the Eastern Vermont sequence, along the CYL) indicating at least a major erosional interval prior to its deposition. It is clearly, to us, the youngest unit in the area, and may, in its upper part, be synorogenic with the Acadian orogeny.

### Eastern Vermont Sequence above the RMC

Much of eastern Vermont is underlain by units that were grouped collectively in the nineteenth century as the "Calciferos Schists" (Edward Hitchcock and others, 1861). The Calciferous Schists have since been subdivided, but we have shown them in one pattern on Plate I, except for two thin, mafic metavolcanic units, the Standing Pond and Putney Volcanics, that are undoubtedly more reliable time markers than the gradational boundaries that separate the major subdivisions.

The principal metasedimentary subdivisions are, in the center the Waits River Formation, on the west the Goshen (or Northfield) Formation, and on the east the Gile Mountain Formation. The boundaries between these are indicated by dashed lines on Plate I, except where arbitrarily defined by the presence of the Standing Pond. The dominant rock types are impure gray ankeritic limestones and calcareous schists weathering to a characteristic brown punky crust, gray phyllite or schist, and gray micaceous quartzites, that may also be calcareous. Carbonate rocks are dominant in most of the Waits River, pelites in the Goshen, and schistose quartzites in the Gile Mountain, but each type is found in each unit, indicating that these units may be, to a large extent, lateral facies of one another. Tops evidence is sparse and rarely near enough to key contacts to be of much help. Our best evidence is that tops are east across the RMC and also east across the Putney Volcanics toward the CYL. We currently favor the Gile Mountain as here being younger than the Standing Pond, but have no proof of this, and also can not rule out the possibility that the Standing Pond and Putney may be equivalent. The matter is troublesome because reversal of tops across the Standing Pond means a reversal of vergence of the nappes outlined by it! If you see something convincing that we have missed it would be much appreciated! Fisher and Karabinos (1980) have shown, in at least one area, that Gile Mountain lithic types are ~~there~~ younger than the adjacent Waits River types, but there remains the nagging doubt that this may not be widely applicable. The boundary in question is clearly not a time line and hence tops across it can easily reverse if tongues of one lithofacies penetrate the other laterally.

The Calciferous Schists are strikingly similar in all variants to the rock types found in the Schistes Lustrees or Bundnerschiefer, the youngest preorogenic rocks of the western Alps. We suggest that the Calciferous may have been deposited in an extensive interarc basin during the Taconian-Acadian interval, possibly floored by oceanic crust although no record of this now survives. The Acadian orogeny may then represent the closure of this basin welding an island arc and an Avalonian terrane to the North American craton on eastward-dipping sutures, one probably buried beneath the Littleton formation east of the CYL and the other east of the Bronson Hill terrane.

The unit that separates the Calciferous from the RMC on the west is here assigned to the Hawley Formation, originally defined by B. K. Emerson (1898) in western Massachusetts. It is also probably equivalent in part to the Russell Mountain Formation of Massachusetts and to the Shaw Mountain Formation and the lower part of the Northfield Formation as first defined by Currier and Jahns (1941) in the Northfield area, 30 km to the north of Plate I. Still farther north, near Hardwick, Vermont (Konig and Dennis, 1964), the Shaw Mountain has yielded Silurian faunas. The dominant rock types in the Hawley are volcanics, almost exclusively mafic, interlayered toward the south of Plate I with highly carbonaceous and sulfidic schists. Conglomerates occur at two positions in the Hawley, the more extensive one close to the base, and the other near the contact with the Calciferous Schists. The lower conglomerate is associated with gray, to slightly rusty weathering schists, and, but rarely found, a thin, buff-weathering, muscovitic limestone a short distance above the conglomerate, but separated from it by a schist containing cotecule nodules and some amphibolite. Cotecules (fine-grained granulites made mainly of manganif-

erous garnet and quartz, commonly with cummingtonite) are found throughout the Hawley, but are best developed in the upper part where they may be as much as forty feet thick and grade locally into iron formations. These cotucules are probably Fe-Mn cherts related to marine vulcanism. The boundary between the Hawley and the overlying Calciferous Schists is sharp and, in Massachusetts, has been interpreted as a decollement (Hatch, 1975). In any case there is an abrupt change in sedimentary lithofacies at this boundary suggesting at least a break in sedimentation. The presence of two distinct conglomerate zones in the rocks here assigned to the Hawley suggests further that there may have been considerable tectonic activity during the time interval that the formation represents.

### Eastern Vermont Sequence below the RMC

The map units below the RMC are, in descending order, the Barnard Gneiss, and the Moretown, Stowe, Ottauquechee, Pinney Hollow, Hoosac, and Tyson Formations. On Saturday we shall see these rocks on the eastern flank of the Green Mountains, then see most of them again on Sunday in the mantles of the Chester and Athens domes. The RMC is believed to represent at least a considerable erosional interval inasmuch as rocks assigned to the Hawley in the mantle of the Chester dome cut down across the three upper units to rest directly on the Ottauquechee.

The Barnard Gneiss is made of felsic gneisses and amphibolites derived from bimodal volcanics and what are probably their comagmatic intrusives. The most recent radiometric dating (Karabinos, personal communication, 1992) indicates an Early to Middle Ordovician age of ca. 480 ma. An earlier, younger date of ca. 418 ma (Aleinikoff and Karabinos, 1990) is probably anomalous, but still worrisome. The underlying Moretown Formation is made of mica schists and schistose quartzites, the latter characteristically having a "pinstripe" texture in which granulitic layers commonly on the order of a centimeter thick are separated by thinner laminae of micaceous material. The pinstripe quartzites of the Moretown are much like rocks mapped by Alpine geologists as metaradiolarites. Slates, more or less on strike to the north near Magog, Quebec, and probably correlative, contain Middle Ordovician graptolites (Berry, 1962).

The Stowe Formation is typically a pale green mica schist or phyllite containing biotite and garnet, but also includes amphibolite of probable volcanic origin. It is thin but persistent along the east flank of the Green Mountains, but is absent or too thin to show in the mantles of the Chester and Athens domes. The next unit, the Ottauquechee, is dominantly a dark gray or black, highly sulfidic phyllite or schist. Fine grained quartzites, some of which are highly carbonaceous are also characteristic, especially near the base. There are also minor, non-carbonaceous schists and some amphibolites. The Pinney Hollow Formation is similar to the Stowe but much more extensive, and will first be seen at a lower metamorphic grade, with little or no garnet and abundant chloritoid. Greenstones, grading into amphibolites are abundant in the upper part, especially in the more southerly exposures and in the mantles of the domes.

The dominant rock type in the Hoosac Formation, beneath the Pinney Hollow, is a mica schist containing abundant albite porphyroblasts. North of Ludlow, in the northwest corner of Plate I, the upper part includes a Plymouth Member consisting of quartzites overlain by dolomite and dolomite breccia. These last resemble parts of the Lower Cambrian Cheshire Quartzite and Rutland (Dunham) Dolomite both of which crop out about 24 km to the west, in and near the city of Rutland (see Thompson, 1972). The base of the Hoosac is locally rich in magnetite, especially where it rests on dolomites of the underlying Tyson Formation. In places the Hoosac, containing a few pebbly beds, rests directly on the basement of the Green Mountain core. Mafic metavolcanics (Turkey Mountain) occur in the Hoosac farther south on the east flank of the Green Mountains and in the mantles of the Chester and Athens domes.

The Tyson Formation is, where present, the basal unit of the cover sequence on the east flank of the Green Mountains. It rests with profound unconformity on the gneisses and metasediments of the Mount Holly Complex (Grenville equivalents). The lower part includes conglomerates and greywackes overlain by quartzite and carbonaceous schists, and the upper part, where present, is a massive dolomite. Iron ores near its upper contact with the Hoosac were mined until about the time of the Civil War. These are associated with a distinctive, coarse dolomite breccia, and are believed to represent a *terra rossa* developed during an erosional interval prior to the deposition of the Hoosac. The basal clastics, in the south part of Ludlow, contain a thin layer of augen gneiss that appears to be an integral part of the sequence. More extensive augen gneisses, farther south in Jamaica, have been dated at ca. 960 Ma (Karabinos and Aleinikoff, 1990) and may be akin although we would be surprised if the Tyson were that old. The carbonate rocks of the Tyson have been recognized outlining the Hoosac Schists in the deeper infolds of cover rock in the Chester dome, but the basal clastics have not been identified in the domes with any

certainty. It is possible that some of the marginal, slabby gneisses and augen gneisses in the domes, discussed more fully below, may be lower Tyson equivalents.

The main part of the cover sequence, from the Hoosac through the Moretown, on the east flank of the Green Mountains has, though at a higher metamorphic grade, many features in common with the Cambro-Ordovician sequence found in the Taconic klippen west of the Green Mountains (see Thompson, 1972, for possible correlations). The principle differences are in the presence of abundant mafic metavolcanic material, and in the presence of ultramafic rocks in the Eastern Vermont Sequence (but only *beneath* the RMC). We see evidence for a pervasive east-over-west movement throughout the Eastern Vermont Sequence below the RMC, so that the highest units in the section have experienced the greatest net translation. It is interesting that the lowest units (Tyson and locally the Hoosac) have carbonates suggesting affinities with the parautochthonous carbonate bank sequence west of the Green Mountains, and that there are also resemblances to the continental slope sequence of the Taconic klippen. Mafic volcanics and ultramafics in the upper Pinney Hollow and Ottauquechee suggest an ocean floor environment, as would possible radiolarites in the Moretown. The metavolcanic rocks of the Barnard Gneiss are possibly related to an island arc. Other investigators (Stanley and Ratcliffe, 1985) show discrete thrusts at most formation boundaries, but we regard this as an oversimplification that fails to recognize the pervasive movement throughout the section. Although we show one through-going thrust, the evidence for it is less than wholly convincing. One unit in which the internal movement may be extreme is the Ottauquechee Formation much of which is lithically (or protolithically) like the Alum Shales of the Scandinavian Caledonides. In the mantles of the Chester and Athens domes the Ottauquechee shows a remarkable, large-scale pinch-and-swell and is greatly thickened in the hinge region of the basement-cored nappe of the west side of the Chester dome. The Ottauquechee in the dome mantles also contains most of the ultramafic rocks in the area of the domes, as well as what appear to be detached slabs of both overlying and underlying units.

#### **Precambrian Basement, Green Mountains**

The "Mount Holly Complex," is a name often applied to the basement rocks in the core of the Green Mountains. Radiometric dating (Karabinos and Aleinikoff, 1988, 1990), however, and lithologic correlations, make it virtually certain that the Mount Holly is essentially a continuation of the Grenville of the southeastern Adirondacks, strongly reworked by Paleozoic tectonic and metamorphic events. The major rock types are felsic gneisses, amphibolites (some with large, partially chloritized garnets), and a variety of metasedimentary rocks including coarsely crystalline marbles and calc-silicate rocks, mica schists (many with coarsely crystalline graphite), and quartzites, some of which, as on Okemo Mountain west of Ludlow, are massive and vitreous. The Paleozoic metamorphism has produced retrogressive features that may be seen throughout, but are most pervasive and conspicuous toward the western margin of the massif. These features include chloritization of mafic minerals, saussuritization of plagioclase, and blue, opalescent quartz, a light-scattering effect, due, in most instances, to the exsolution of fine rutile needles (Niggli and Thompson, 1979). These retrogressive features are relatively inconspicuous on the east side of the Green Mountains, particularly to the south, near Jamaica, where the Paleozoic metamorphism reaches its highest grade (staurolite-kyanite zone). Structural trends in the Green Mountain core are strongly discordant to those of the overlying Paleozoic cover.

#### **Precambrian Basement, Chester and Athens Domes**

The deeper cores of the Chester and Athens domes are mainly felsic gneiss with minor amphibolite. Metasedimentary types are best seen in the Athens dome where there are marbles, calc-silicates, and schists containing coarsely crystalline graphite. In both domes an inner core has been separated on Plate I from an outer one, either by a thin zone of marble, calc-silicate, and schist, or by a dashed line indicating that this boundary is obscure at best. One of the more extensive rocks in this outer core region is an augen gneiss (Bull Hill Gneiss) dated at ca. 960 Ma (Karabinos and Aleinikoff, 1990). Other types are amphibolites (minor) and slabby, phengitic gneisses much used in local construction, most notably at Chester Depot. These last are quite similar to the Alpine "Maggia Lappen" that are quarried for the same purpose just north of Locarno, Switzerland. It is possible that some or all of the outer core rocks may actually be equivalent to the lower part of the Tyson Formation. There is clearly room for further work on this!

## TECTONO-STRATIGRAPHIC HISTORY

### Mesozoic to Recent

Features recording the glacial and post-glacial history are abundant in the area of Plate I, but are of little concern to this excursion, although we shall point some of them out in passing. The earlier Mesozoic and Cenozoic history of the New England Appalachians has been summarized in Thompson and others (1993, p. 11-18).

During the Triassic-Jurassic rifting the area of Plate I was almost certainly buried beneath red beds, lavas, and lake sediments in the down-dropped region between the Ramapo fault system on the west and the Connecticut Valley fault system to the east. The nearest sediments of the Hartford-Deerfield basin are less than 20 miles south of the southeast corner of Plate I, and the nearest faults of the Connecticut Valley system cut through the Bronson Hill just over a mile east of the CYL. Deep erosion of the Paleozoic rocks had taken place before the rifting began. The erosion surface beneath the Mesozoic sediments in the Hartford-Deerfield basin dips easterly, and that beneath the sediments of the Newark basin dips westerly toward the Ramapo fault, showing that the central region between the two fault systems has also been uplifted and deeply eroded *since* the rifting. The absence of Mesozoic sediments in the breccia at Mount Ascutney (Schneiderman, 1989), just west of the Connecticut Valley fault system, suggests that post rifting erosion had already been extensive by the mid-Cretaceous when the Ascutney volcanic edifice was formed, although the Chester dome was then still covered by its metasedimentary mantle. Further erosion, by the end of the Cretaceous produced what is now a high-level erosion surface, of which there are still remnants in the area of Plate I. This surface (by no means a peneplane) is found between 2000 and 2500 ft in the Green Mountains, and between 1200 and 1500 ft in the hills in the vicinity of the Connecticut Valley. Subsequent etching of this erosion surface had, by the Middle Oligocene, produced a topography not very different from what we see today.

### The Acadian- Alleghanian Interval

There is abundant evidence that this area was deeply eroded immediately following the Acadian Orogeny. The materials removed went in part into the Catskill delta and into areas farther west on the North American craton. Much of this erosion had probably been accomplished by the time of deposition of the Pennsylvanian coal measures farther west, and on the New Brunswick platform to the northeast. The erosional history outlined here and above is entirely consistent with the Al-silicate triple point isobar (Thompson and Norton, 1968), as now exposed at the erosion surface in western New England.

The effects of the Alleghanian Orogeny are mainly confined to southern and coastal New England (see Thompson and others, 1993, p.18-22, for a summary of recent literature). It is possible, however, that the sinistral (east-side-north) movements recorded by the folds and faults along and near the Chicken Yard Line (stops 1-2 and 1-3 below) are of this vintage. They are the youngest structural features seen in the area of Plate I except for some sub-horizontal kink bands possibly associated with Mesozoic faulting. These sinistral movements may thus be complementary (see Thompson and others, Fig. 3) to the more conspicuous dextral movements of coastal New England and adjacent Maritime Canada.

### The Taconian-Acadian Interval

This interval is better documented in the Bronson Hill terrane immediately to the east (see Thompson and others, 1968; Robinson and others, 1979, 1991) than it is in eastern Vermont. This is due in part to better biostratigraphic control, and in part to unambiguous facing at key contacts. In brief the main features in the Bronson Hill are the Siluro-Devonian sedimentation on and east of the remnants of a Taconian island arc, followed by collision with a land mass from the east resulting in westward verging nappes carrying hot-rocks westward over colder ones, the axial surfaces of these nappes being deformed by subsequent doming.

In eastern Vermont there are also early Acadian nappes followed by later doming but there are uncertainties here caused by fewer fossil localities and the ambiguous facing across the Standing Pond. We envision the Siluro-Devonian sedimentation here as being in a broad basin bordered on the west by the newly formed Taconian orogen, in which an island arc had been recently attached to the craton, and bordered on the east by a second Taconian island arc. This last is the one involved in the Bronson Hill terrane, probably above an eastward-dipping subduction zone, that may have caused the Siluro-Devonian volcanism in the Bronson Hill area. We suspect that the basin in which the Hawley and the overlying Calciferous were laid down was floored by oceanic crust but have no proof. Where we

see it the Hawley, as now understood, rests on the Barnard Gneiss which we believe to be a remnant of the western arc. To the south, near the Massachusetts line, the Hawley cuts down to the Moretown, and to the north, along the east side of the Chester dome, it cuts down as far as the Ottauquechee. Volcanism and the occurrence of conglomerates at at least two levels within the Hawley, and the sharp break in sedimentary style at the base of the Calciferous suggest an early period of unrest followed by the relatively placid accumulation of the Calciferous, interrupted briefly by the Standing Pond volcanism.

We believe that the break at the Chicken Yard line represents a flooding of Littleton-type clastics from the east across the Bronson Hill at an early stage of the Acadian Orogeny, with the Littleton soon overridden by the hot-rock nappes from the east. So far so good. The problem is that the Acadian nappes developed *within* the calciferous are eastward verging, at least if our currently favored topping across the Standing Pond is correct! One possible explanation for this is that the relatively plastic Waits River may have been extruded westward due to the onloading from the east by the overriding nappes from the Bronson Hill terrane (Rosenfeld, 1968; Fletcher, 1972).

Later stages of the Acadian orogeny are marked here, as in the Bronson Hill terrane, by the final closure of the sedimentary basin followed by the partly forced and partly buoyant rise of the gneiss domes and emplacement of granitic rocks in the collisional region. We also believe that the collisional stage was accompanied by a renewal of the Taconian thrusts west of the Green Mountains (Ando and others, 1984) causing the Green Mountain massif and the Chester and Athens domes to override the craton with attendant uplift. We suggest further that this uplift caused a *retrocharriage* or back-flow eastward off the massif and the domes as a final stage. The evidence for this back-flow in the area of this excursion is best shown by the late stage reversal of rotation shown by the outer rims of the garnets in the lower part of the Calciferous on the east side of the Chester and Athens domes. Actual back *folding*, however, has been suggested for the area immediately to the south of Plate I by Rosenfeld in Hepburn and others, (1984, p. 93-100), and farther south, in central and eastern Connecticut by Rosenfeld and Eaton (1985).

### **The Late Proterozoic-Taconian Interval**

The Taconian Orogeny and the events leading up to it are quite well understood, at least as far as the area west of the Green Mountains is concerned (see Thompson and others, 1993, p. 27-31, for a recent summary), but there is some divergence of opinion concerning what went on east of the Green Mountains. There is general agreement that the Tyson, and locally the Hoosac, both rest with profound unconformity on the deeply eroded Grenvillian rocks of the Mount Holly Complex. The *terra rossa* at the top of the Tyson, however, implies a substantial hiatus between the deposition of these two units. The relation of the Tyson to the 960 Ma augen gneisses and related rocks farther south in the Green Mountains, and in the Chester and Athens domes, however, is uncertain and needs further radiometric work. We would recommend starting such a study with the thin augen gneiss that appears to be within the Tyson near Ludlow.

Cross-mountain correlations suggest that the Hoosac is probably Lower Cambrian and that its Plymouth Member is probably correlative with the Cheshire Quartzite and the overlying Rutland (Dunham) Dolomite. We regard the Hoosac as mainly a continental slope facies, containing an eastern feather edge of the carbonate platform sediments to the west. The Pinney Hollow has much in common with some of the more abundant rocks of the Taconic allochthon, and may represent a lower slope to ocean floor facies. The Ottauquechee, Stowe and Moretown have many features, including ultramafics, suggesting an ocean floor origin. We believe that the Taconian Orogeny was caused by the docking, on an eastward-dipping subduction zone, of an island arc of which the Barnard Gneiss is a remnant. We suggest further that the docking of this arc caused a pervasive east-over-west movement throughout the rocks between the Barnard and the Mount Holly basement, but with a lion's share of this movement concentrated in the Ottauquechee. The large accumulation of Ottauquechee in the western hinge region of the basement-cored, westward-verging nappe west of the Chester dome is consistent with such behavior. We believe that the net effect of these east-over-west movements was to cause the accumulation of an accretionary wedge in the western Green Mountain region and beyond, and that the Taconic allochthon is a remnant of it.

A great amount of loading, probably of arc material, must have then been present above eastern Vermont to account for the Taconian blue schist metamorphism recorded farther north (Laird and Albee, 1981; Laird and others, 1984). Large garnets in the Hoosac and Pinney Hollow in the mantles of the Chester and Athens show textural; and rotational unconformities, as seen at stop 3-8, below. The rims are clearly Acadian, but the cores are Taconian or even older. These will receive further study.

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### ROAD LOG

Because this guidebook will probably have a life after the 1992 GSA meeting, we have written the road log for the needs of persons who may wish to follow the route later by private car. The GSA group will be traveling in a bus or vans and therefore will not need a mileage log. We anticipate that these vehicles may have to take more roundabout routes, in several instances, owing to such factors as bridge capacities, road widths, and the like. We also should warn the GSA group that not all of the stops listed below may actually be visited, either for the reasons implied above, or for time limitations, foul weather, or unforeseen emergency. Our reasoning is that it is better to have the information in here, and to have it available if needed, than to be sorry that it was omitted.

#### Day One, Friday

The starting point for mileage is at the junction of Vermont Routes 11 and 143 (in downtown Springfield, about 0.3 miles SW of the Hartness House). We will proceed east on the "Skitchewaugh Trail" (Route 143).

#### Mileage

- 0.9 Outcrops (initially on left) of Gile Mountain, continues for 0.3 mi.
- 1.7 Standing Pond Volcanics in pasture on left.
- 1.9 Waits River Formation on right.
- 2.3 Intersection, go right.
- 3.0 Quartz conglomerate of Clough Formation in pasture on left. We are now east of the Chicken Yard Line (CYL), in rocks of the New Hampshire Sequence. The valley here is Spencer Hollow, once occupied by the early post-glacial Connecticut River, now about 300 ft below and 1.5 mi farther east.
- 3.2 Just beyond sharp left turn, on the right, is another outcrop of the quartz conglomerate with large potholes.
- 3.5 Rusty weathering, pyrrhotitic schist of the Partridge Formation on the right. This Partridge is in the core of the Bernardston nappe. Ledges of the overlying Clough are in the woods to the northeast and east.
- 4.2 Enter road cuts in the Littleton Formation on the inverted limb of the Skitchewaugh nappe. Near the underpass beneath Interstate Highway I-91 there are outcrops high on the right, and in the cuts on I-91, of the "pit-rock" of the Fitch Formation, here atop the younger Littleton.
- 5.8 Turn sharp right onto Route US 5, and follow it south. Connecticut River is on left.
- 7.4 Amphibolite and pit-rock of the Fitch Formation.
- 7.7 Ditto.
- 8.2 Stop at outcrops on right near intersection.

**STOP 1-1. FOSSILIFEROUS QUARTZITE IN THE UPPER PART OF THE CLOUGH FORMATION.** Rusty-weathering calcareous zones (coquinas) in the quartzite, contain numerous highly deformed fossils and fossil fragments (Boucot and Thompson, 1963). Recognizable remains are chiefly crinoid columnals and corals, but brachiopods, a cephalopod, and a possible trilobite have also been found. Just south of the principal road cut, a large mafic dike, metamorphosed to epidote-amphibolite, cuts across the quartzites, which here dip to the northwest. This locality is on the upper limb of the Skitchewaugh nappe.

- 9.5 Garnet-bearing schists and pebbly quartzites in central part of the Clough.
- 9.8 Basal conglomerate of the Clough, mainly vein quartz or quartzite cobbles, but some parts polymictic.
- 10.1 Better exposures of ditto. Drive by slowly.
- 10.6 Rusty schist and amphibolite of the Partridge in core of the Skitchewaugh nappe. Plaque just beyond marks southeastern terminus of the Crown Point Road, dating from the French and Indian War.
- 11.4 Junction with Route 11, go right.
- 11.9 Underpass, I-91, outcrops of Littleton just beyond on left.
- 12.2 Park immediately beyond Howard Johnson's.

**STOP 1-2. CHICKEN YARD LINE.** Bus will park in the Howard Johnson's lot. The CYL (see Trzcieski and others, 1992), midway in the big cut on the north side of Route 11, is a nearly vertical contact between gray phyllites and rare sandy beds of the Littleton on the east, and greenstones and related rocks of the Putney Volcanics on the west. Several boudins of sheared, pebbly quartzite can be seen along the contact. Is this a fault, or is it a zone of extreme thinning and attenuation with consequent boudinage of the relatively brittle quartzite? We then walk south on Route 5 to an outcrop in a clump of trees on the south side of a driveway, leading to a YMCA camp. This goes off to the right several meters south of the bridge. The exposure is of a fold plunging steeply south in quartz conglomerate and quartzite. The fold opens northward to include some gray phyllite of the Littleton. Return to Route 5 and walk a short distance south to an outcrop of greenstone on the right. This is in the core of a south-opening fold complementary to the north-opening one in the clump of trees. These define one of the steeply plunging, east-side-north folds, that offset the CYL from here south. There are also (Plate I) several faults that produce offset in the same sense. A few meters farther south the Littleton reappears, east of the greenstone, and includes some phyllite-matrix conglomerates. Return to vehicles and continue west on Route 11 past some low outcrops of Putney on the right.

- 12.5 Large cut on right has nearly vertical contact between gray calcareous phyllites of the Gile Mountain on the east and greenstones of the Standing Pond Volcanics on the west.
- 12.7 Contact on the left between Standing Pond, on the east, and phyllites and punky-weathering gray limestones of the Waits River Formation, on the west.
- 13.1 Park on right opposite cut in punky-weathering limestones of the Waits River Formation.

**STOP 1-3. WAITS RIVER-STANDING POND CONTACT.** Since Stop 1-2, we have crossed the Putney, Gile Mountain, Standing Pond, and a fold axis in the Waits River. We are now at the east edge of a second appearance of the Standing Pond. This belt of Standing Pond is part of the eastern "gunwale" of the canoe-shaped outcrop pattern shown by the Standing Pond on Plate I. These cuts give a complete section through the Standing Pond to its contact with the Gile Mountain to the west. Near the west contact is what appears to be a dike that has yielded a zircon age of 423 Ma (Aleinikoff and Karabinos, 1990).

Convincing primary evidence on the direction of tops at either contact would be most welcome! There is, however, one feature at the Standing Pond-Waits River contact that may have a bearing on this. The Waits River and the crystal tuff in the Standing Pond at the contact are presumably both marine. A large oriented thin section cut across this contact shows the tuff to have no observable apatite except for a single grouping of three apatite crystals sitting exactly on the contact. One is almost 2 mm long with its long dimension parallel to the contact and its c-axis normal to the contact. Another is much smaller and shares an interface with the first. The third is smaller yet and slightly detached from the second. This grouping is probably derived from a detrital grain or grains as are the adjacent feldspars. Apatite porphyroblasts of this size are not known in rocks of so low a grade. It is thus possible that the higher density of the apatite caused it to reach bottom before less dense feldspars of comparable size, suggesting that the Waits River is older than the Standing Pond.

A steep, northeasterly striking crenulation cleavage is probably related to the steeply plunging, east-side-north folds seen at stop 1-2. The Standing Pond here also contains lapilli tuffs, but these are indistinct on the fresh

surfaces, and the weathered surfaces, at the tops of the cuts, are not safely accessible. Board vehicles again at the east end of the cuts.

- 13.2 Turn right and cross river on iron bridge.
- 13.4 Waits River on left.
- 13.5 Turn left on road climbing steeply away from river.
- 13.9 Outcrops on right of Standing Pond continue for about 0.4 mi.
- 14.7 Littleton phyllites.
- 14.9 Bear left at fork, Littleton on left just beyond. This is a section, still in use, of the Crown Point Road.
- 15.8 Turn left on Route 143.
- 16.0 Turn right onto Eureka Street in what was once downtown Springfield.
- 16.2 Park in farmyard (with permission of owner!) and hike west along north line of property 0.4 miles to northwest corner of partly overgrown clearing.

**STOP 1-4: LAPILLI TUFF AND CRYSTAL TUFF IN THE STANDING POND.** No hammers, please! We are here in part of a large, north-plunging fold in the east gunwale. Note especially the extreme elongation of the lapilli, *parallel* to the fold axes! "Balanced" cross sections would encounter grave difficulty here if such strains were not properly taken into account. On the hill to the west the Standing Pond, now offset westward, contains both garnet and hornblende. Return to vehicles via same route.

- 16.7 Waits River on right.
- 17.7 Go left at intersection.
- 18.3 Waits River on right.
- 18.4 Go right at intersection.
- 18.6 Waits River in low outcrops on right.
- 18.9 Standing Pond, still in east gunwale, containing amphiboles.
- 19.2 Gile Mountain on left. Sag in partly grown up pastures to right has exposures of crystal tuffs and related rocks of the Standing Pond. These contain both amphibole and garnet. Gile Mountain continues north along road.
- 19.6 Turn left at corner.
- 20.1 Turn left again onto Weathersfield Road, then right at sign "Koledo" and park with permission of landowner.

**STOP 1-5. STANDING POND, HIGHER METAMORPHIC GRADE.** Outcrops in the fields are Gile Mountain, the hill to the west is in the Standing Pond of the west gunwale. The Standing Pond, here at a higher metamorphic grade than yet seen, includes a spectacular display (no hammering, please!) of coarse garbenschiefer with large garnets and sprays of hornblende. The clearing at the top of the hill provides a panoramic view across the Chester dome to the Green Mountains beyond. Return to vehicles and drive south toward Springfield.

- 20.9 Glacial erratic on right is syenite from the Mesozoic stock at Mount Ascutney.
- 22.6 Springfield, turn left on Route 11.
- 23.0 Cross Black River and continue straight ahead up hill, leaving Route 11.
- 23.4 Bear left at top of hill.
- 25.4 Hardscrabble Corner, turn right.
- 26.3 Go left at intersection, Waits River on right just beyond.
- 27.5 Turn left at sign "Yoder" and park in driveway turnaround with permission of landowner.

**STOP 1-6. COBBLE HILL.** Again, with owner's permission, walk south across fields into overgrown former pasture with outcrops of amphibolite showing flattened lapilli. To the east, on the west slope of Cobble Hill, are crystal tuffs, in north-plunging, west-side-up folds. Higher on the slope are some fine garbenschiefer (no hammering please!). Large garnets show west-side-up rotation. Note that a fine bedding is preserved only inside the garnets.

- 28.1 Height of land, Standing Pond of the narrow band west of the "canoe" on right.
- 28.6 Standing Pond amphibolites in pastures to left.
- 30.1 Park, as best one can, on the right.

**STOP 1-7. SNOWBALL GARNETS REVEALED.** A no-hammer outcrop! This is locality T9e of Fig. 14-4 of Rosenfeld (1968). Proceed westerly across north end of field, past small cottage, to contact between garnetiferous phyllite of Waits River Formation on west, and coarse garbenschiefer (fasciculitic schist) of the Standing Pond on the east. The contact is on the west limb of a very tight nappe in the Standing Pond immediately west of the somewhat broader "canoe" nappe whose limbs we explored at stops 1-3, 1-4, 1-5, and 1-6. This locality, which could easily be missed, is probably the best locality found for seeing evidence of both of the two main Acadian events within a single rotated garnet. This is because the garnet to be observed is broken open along a weathered ankeritic lamina passing through its center, and thereby exposes in three dimensions the two rotational stages. Because of scale, observation will involve close-up, guided looking by small groups of no more than five at a time. Also the exposure is next to the cabin owner's flower garden. Please try not to trample things too much. A 1:1 stereoscopic photograph of the principal rotated garnet you will see at this locality appears as figure 14-3 in Rosenfeld, 1968, p. 192. It rotated clockwise  $87^\circ$  about a north-pointing, nearly horizontal axis during the the first event, possibly a westerly extrusion of the Calciferous structurally beneath it, and rotated  $88^\circ$  clockwise about an axis plunging  $40^\circ$  northeasterly during the later doming and back-flow. A discussion of the generation of the central surface of a garnet at this locality is found in Rosenfeld, 1970, p. 40.

Although the back-flow affected the crest of the Chester dome it is not shown by a late growth of the garnets there, and must be inferred from other evidence. There is a simple explanation for this contrast between the sides of the dome and the crest. It comes from ideas first applied to geology almost simultaneously, more than 150 years ago by mathematician-geologist Charles Babbage and by Sir John Herschel (Babbage, 1989, v. 9, p.84-87; 94-97), both evidently inspired by their mutual friend, Charles Lyell, and the then recent pioneering work on heat transfer by Fourier. This contrast probably results from the arching upward of isothermal surfaces during doming and the resulting thermal relaxation that caused cooling near the crest and simultaneous heating on the flanks of the domes during and after doming. The concepts of Babbage and Herschel, had been almost completely ignored since the last century until revived by Sleep (1979, p. 584-585). Sleep's work has played an important role in prodding us to look for examples of this in the rocks.

- 30.2 Gile Mountain on right.
- 30.4 Sharp right turn, Waits River on right.
- 30.5 Park to left.

**STOP 1-8. STANDING POND IN BLACK RIVER.** A no-hammer outcrop! The locality is on the north bank of the Williams River 300 m south of the road south of the pond by the elbow in the road to Springfield east of Brockway Mills. Go south from the road along the east edge of the field, cross the railroad tracks, and continue south along the west side of the north-south stone wall and barbed wire fence down the steep embankment to the contact that was seen at stop 1-7. Large garnets show only a single large clockwise rotation associated with the first Acadian event, in contrast to those at stop 1-7. Here it is possible to see a number of garnets broken open along weathered ankeritic laminae passing through their centers, and thereby to determine easily their axes of rotation. A photograph of a rotated garnet from the same zone 1.3 km south, showing similar rotation around a south plunging axis of the first event, appears as figure 14-6 in Rosenfeld (1968, p. 195). The outcrop presents surfaces of enough different orientations that cross sections parallel to the garnet rotational axes can also be observed and their axial orientations determined. Evidence of the second event at both localities appears only as crinkles. Also observe other types within the Standing Pond, which carries the axis of the tight nappe mentioned in the description of stop 1-7. Recent detailed mapping, still in progress farther south, shows that rocks of the Calciferous track the axis of this tight nappe in a series of axial culminations. The nappe plunges south east of Westminster West probably to reappear and arc southwesterly over the northwest part of the Guilford Dome west of Brattleboro as shown by Hepburn and others (1984, Plate 1). This fold in the Standing Pond is probably also the one whose crest "porpoises" with low-angle plunges as far south as the latitude of Greenfield, Massachusetts.

- 30.8 Turn left, cross bridge, and park on left just before railroad tracks.

**STOP 1-9. BROCKWAY MILLS.** This is the main zone of limestone, calcareous schists and subordinate phyllites of the Waits River Formation. This used to be one of the best exposures of the Waits River Formation in southern Vermont before recent construction of a small hydro-power plant. Sprays of zoisite in quartz veins and pods, isoclinal folding, easily observed rotated garnets at lower end of falls. Also a mafic dike with multiple chilled borders. Euhedral calcite phenocrysts have been observed in thin section in one of the borders.

- 31.2 Dangerous intersection. Go left on Route 103.

- 31.4 Turn right (south) onto Pleasant Valley Road.
- 31.8 Gile Mountain on right.
- 33.2 Standing Pond in pastures on left.
- 33.7 Waits River in septum between west gunwale of "canoe" and narrow westerly belt of Standing Pond.
- 33.8 Amphibolites of the Standing Pond. Clockwise (west side up) folds. Road continues southerly along the Standing Pond-Gile Mountain contact.
- 36.1 Dangerous intersection with Route 121. Go straight ahead (east) on 121.
- 36.4 Intersection in Saxtons River village. Park vehicles.

**STOP 1-10. SAXTONS RIVER.** Exposures are in the Saxtons River below and to the east of the bridge on the Pleasant Valley Road just south of the intersection. The purpose of this stop is to observe southward plunging minor folds in the Standing Pond along the axis of the "canoe" nappe whose limbs we observed at stops 1-3, 1-4, 1-5, and 1-6. Folds with counter-rotating garnets on their limbs appear along the north side of the river, 0.3 miles to the west (Rosenfeld, 1970, p. 85-86). Return to vehicles and head west on Route 121.

- 38.4 Waits River on right.
- 39.3 Amphibolites of Hawley Formation on right.
- 39.4 Turn right (north) at intersection, leaving Route 121. Low outcrops of Barnard gneiss on right just beyond.
- 39.8 Bear left at fork.
- 42.0 Ascend through unnamed "gulf" (Vermontese for narrow, more or less on-strike defiles). The course of this gulf is determined by the easily weathered, sulfidic schists of the Ottauquechee Formation.
- 42.1 Park on right at head of gulf.

**STOP 1-11. MORETOWN FORMATION AND BARNARD GNEISS.** The exposures are on the steep slope east of the road. A glimpse of the carbonaceous, sulfidic schists of the Ottauquechee followed by quartzose schists and "pinstripe" quartzites of the Moretown, showing some interlayering with felsic gneisses and amphibolites characteristic of the overlying Barnard Gneiss

- 42.8 Moretown Formation on the right.
- 43.5 Park on right.

**STOP 1-12. BARNARD GNEISS AND HAWLEY FORMATION.** Just east of the road are outcrops of Barnard gneiss from which Aleinikoff and Karabinos (1990) obtained a zircon age of ca. 524 Ma, but more recent work by Karabinos (personal communication 1992) has revised this figure to ca. 485 Ma. To the east, across the Barnard Gneiss, are some rusty and gray schists containing coticules and amphibolites, here assigned to the Hawley Formation. Farther east, on the steeper slope, is quartz conglomerate and quartzite overlain by a thick series of amphibolites, all part of the Hawley.

- 43.8 Barnard Gneiss on right.
- 44.9 Road turns sharp left then right immediately.
- 45.0 Turn right on Route 103.
- 46.0 Turn left on side road with bridge over railroad tracks.
- 46.2 Covered bridge.
- 46.4 Park as best you can.

**STOP 1-13. COTICULE WITH CUMMINGTONITE PARTINGS.** On the north side of the road are some of the distinctive rocks found associated with the mafic metavolcanics in the upper part of the Hawley. Variants not exposed here include thin beds of iron formation. These probably are metamorphosed Mn-Fe cherts related to ocean floor volcanism. Amphibolites of the Hawley crop out on the bend to the east.

- 46.7 Park on left.

**STOP 1-14. DOUBLY ROLLED GARNETS.** Outcrops are ledges in woods north of road. Sieve texture garnets in calcareous schists of the lower Waits River Formation show early counterclockwise rotation (east side up), followed by a later and much less conspicuous clockwise rotation. *Unfortunately*, in spite of published pleas to conserve good field trip stops for future geologists to enjoy, the key, field-visible part of this exposure has been removed. This is the locality of the specimen "discussed in detail" in Bell and Johnson (1989), upon which, Bell and others (1992) have based their questioning of the view that sigmoidal, "spiral-shaped inclusion trails" in

snowball garnets signify rotation during growth. They prefer to consider the patterns as formed from "overprinting of near-orthogonal foliations in schist," evidence of which they claim to find as growth discontinuities in the garnets. A cut and etched oriented specimen collected close to that locality will be displayed on this trip so that field trip participants may make up their own minds as to the general validity of these discontinuities.

47.4 Brockway Mills (Stop 1-9). Turn left on paved road and return to Springfield.

### Day Two, Saturday

On Saturday we shall drive west on Route 11 then north on Route 106 via North Springfield to Woodstock, where the road log begins. The early part of the trip is in the gneisses of the Chester dome. At about 7.0 miles, just north of Perkinsville, the highway enters the Hawks Mountain mulde, remaining in it for about 3.0 miles. Dips are northerly or northeasterly. We then cross more gneisses, dipping northerly, to Hammondsville where, at about 16.0 miles, we pass into the mantle of the Chester dome, crossing the RMC at about 16.5 miles. We then ascend through the Calciferous, crossing a pelite-quartzite zone in the Waits River, that marks the axial surface region of the "canoe" in the Standing Pond. At the height of land, at about 19.0 miles, are exposures on both sides of the road of more typical Waits River. Note that dips have been consistently north for the last twelve miles! At Woodstock we start the road log at the junction of Route 106 with U. S. Route 4, and turn west following the Ottauquechee River.

7.3 Watch for oncoming traffic and park on left.

**STOP 2-1. MORETOWN FORMATION.** Outcrop is on right. Watch out for traffic! Quartzose schists, with faint indication of a pinstripe texture, and with layers of amphibolite and hornblende gneiss.

8.2 Bridgewater Corners.

8.7 Cuts on left are in carbonaceous and sulfidic schists currently assigned to the Hawley Formation.

9.4 Moretown again on left.

9.5 Stowe formation on right.

9,7 Park.

**STOP 2-2. STOWE FORMATION, GARNET ZONE.** The outcrop has the common garnet-zone assemblage: quartz-muscovite-chorite-biotite-garnet. The abundant quartz lenses were probably a byproduct of silica-releasing reactions at the biotite isograd. The metamorphic grade is here dropping westward, toward the Green Mountains. The Stowe is a thin unit here between the Ottauquechee and the Moretown and is not separately mappable in the Chester and Athens domes.

10.2 Rusty weathering, carbonaceous, sulfidic schists of the Ottauquechee Formation on right. For the next two miles, following the Ottauquechee River, we will remain in this distinctive unit. Other rock types are carbonaceous and non-carbonaceous quartzites (in some outcrops layered black and white), quartzose (non-sulfidic) garnetiferous schists, and greenstones or amphibolites.

10,6 Ottauquechee right

11.2 Quartzose, garnetiferous schists on left.

11.6 Ottauquechee, both sides.

12.2 Ditto.

12.4 Park.

**STOP 2-3. OTTAUQUECHEE FORMATION.** These carbonaceous, sulfidic schists and quartzites are in the lower part of the formation. Though mainly carbonaceous here the quartzites may also be free of carbon or contain alternating black and white layers. The schists contain biotite and small garnets.

13.4 Low outcrops of Pinney Hollow Formation on right.

13.5 Park on right.

**STOP 2-4. PINNEY HOLLOW FORMATION.** The cut is in green phyllites containing abundant chloritoid peppering the foliation surfaces. The principal assemblage is: quartz, muscovite-paragonite-chlorite-chloritoid with hematite abundant enough to give some layers a purplish color (best seen, if possible, in bright sunlight). The lower outcrops to the east have the a less aluminous assemblage: quartz, muscovite, biotite, chlorite

and garnet, with magnetite, rather than hematite as the ferric phase. Note crenulations and associated "slip" cleavage, also a down dip streaming of platy mineral aggregates.

- 14.0 Junction with Route 100 at West Bridgewater. Exposure in gravel bank, high on right, is carbonaceous albitic schist in the upper part of the Hoosac Formation. Bear right on Route 4.
- 14.9 Quartzitic zone in the Hoosac on left.
- 15.0 Turn right on side road crossing Ottauquechee River.
- 15.2 Park on left.

**STOP 2-5. TYSON AND HOOSAC FORMATIONS, TERRA ROSSA.** Overhanging ledges On right are albite-rich schists of the Hoosac Formation, here non-carbonaceous, and with a high concentration of magnetite, especially near the base. Upper part of Tyson Formation is massive dolomite, passing upward into a coarse breccia in the last two meters below the contact. The blocks in the breccia are surrounded by a filling of micaceous material and iron oxides, believed to represent a fossil soil. Elsewhere, at this position, iron oxides (both magnetite and hematite) are present, and were mined as iron ore a few miles to the south, near Plymouth Union, early in the last century. These ores are interpreted as a metamorphosed *terra rossa*. These occurrences indicate a major erosional interval between the deposition of the Tyson and Hoosac Formations. To the south the Tyson is locally missing so that the Hoosac rests directly on the basement.

- 15.6 Park on right (time permitting).

**STOP 2-5A (OPTIONAL, TIME PERMITTING).** Magnetite ore at top of Tyson dolomite.

- 15.8 Rejoin Route 4 and turn right (north). Outcrops passed will be described in more detail on our return after Stop 2-6.
- 18.3 Sherburne. Turn vehicles around (watch traffic!) and head back south.
- 18.4 Park on right.

**STOP 2-6. POLYMICCTIC CONGLOMERATE AT BASE OF TYSON FORMATION.** This stop is just off the northwest corner of Plate I. Cobbles of quartzite, gneiss, pegmatite, and rare blue, opalescent vein quartz are derived from underlying "Mount Holly" basement. Note strong elongation in down-dip direction. These coarse conglomerates occur at or near the unconformity with the basement but are not everywhere present. Where Route 4 turns west, just north, it crosses the unconformity at a point where there are a few pebbly zones but no conglomerate.

- 18.6 Outcrops on right until next stop are gneisses and schists in the Mount Holly Complex of the Green Mountain basement.
- 19.2 Park on right.

**STOP 2-7. UNCONFORMITY AT BASE OF TYSON.** In a natural outcrop, just north of cut, we see folds in Mount Holly gneisses that are truncated by metagreywackes at base of Tyson. Note graded beds in metagreywackes. Unconformity is also visible in cut, but not as obvious. Next cuts south are basement gneisses. Highway is very nearly along the unconformity. The valley to the left marks the course of the dolomite member of the Tyson.

- 20.4 Long road cut on right is in interbedded dolomite and quartzite of the Tyson, also some carbonaceous schist. The conspicuous quartz vein fills a nearly horizontal tensional opening that is consistent with the pebble elongation at Stop 2-7.
- 22.7 West Bridgewater. Turn right on Route 100.
- 24.1 Interbedded quartzite and dolomite of Tyson on left. Quarry in hill behind is in the main dolomite.
- 25.6 Dolomite of Tyson Formation on right.
- 25.8 Park on right opposite dolomite quarry.

**STOP 2-8. TYSON FORMATION.** Quarry to left (east) is in main dolomite of Tyson. On the right the basal part of the dolomite contains graded, pebbly beds indicating tops east.

- 26.3 Dolomite of the Tyson crops out over the next mile, both in cuts on the right and in the brook to the left.

- 28.2 Plymouth Union. Turn left (east) on Route 100A. A small adit on the left in the top of the Tyson dolomite, beneath the overhanging ledges of Hoosac schist, is a relic of the iron mining days. The road climbs steeply through the albite schists of the Hoosac. These schists are progressively more quartzitic toward the top of the ascent.
- 28.7 Clean quartzite on left. This is in the Plymouth Member of the Hoosac.
- 29.1 Turn left on side road.
- 29.3 Plymouth. Turn left again (north), passing Coolidge birthplace and cheese factory.
- 30.3 Turn around in driveway and obtain permission of landowner for next stop.
- 30.4 Park to visit outcrops in pasture on right.

**STOP 2-9 (LUNCH). DOLOMITE BRECCIA IN UPPER PART (PLYMOUTH MEMBER) OF HOOSAC FORMATION.** The dolomite here includes an intraformational breccia that is quite unlike that associated with the terra rossa at the top of the Tyson. It rests on quartzites that can be seen in the woods just west of the fields. The dolomite closely resembles a dolomite breccia in the basal beds of the Rutland (Dunham) Dolomite in the vicinity of Rutland, approximately 15 miles west-northwest across the Green Mountains. The quartzite, though here thin, is not unlike the Cheshire Quartzite that underlies the Rutland Dolomite on the west side of the mountains. Both the Cheshire and the Rutland carry Early Cambrian fossils (Olenellus zone). The quartzites here grade downward into the albite schists of the lower Hoosac. More albite schists occur above the dolomite. Retrace route to Plymouth Union.

- 32.4 Plymouth Union. Turn left (south) on Route 100. The valley continues along the Tyson dolomite.
- 34.3 Valley here starts to offset eastward owing to folding and some high-angle thrust faults (see next stop).
- 35.8 Valley straightens out again and follows the western shore of Lake Amherst. The Crown Point Road crosses the valley just north of the end of the lake, which takes its name from General Amherst.
- 36.1 Prominent outcrops on the right are highly deformed conglomerates of the lower Tyson.
- 36.7 Echo Lake is now on our left.
- 37.1 Park in grassy area to the right.

**STOP 2-10. CONGLOMERATES OF DRY HILL.** This is the most strenuous stop of the excursion. It involves about 2 miles of walking and a climb of about 500 feet over rough terrain. If you feel up to it we strongly recommend it. We shall see excellent exposures of highly strained, polymictic conglomerate of the Tyson in slivers separated by east-dipping, high-angle thrusts. We shall also see some of the underlying basement and overlying dolomite. The cobbles and boulders exhibit extreme strain, rarely equaled elsewhere, if at all. Initial east-west elongation has been followed by gently north-plunging folds. The polymictic nature of the conglomerate provides an excellent opportunity to study the effects of both mineral composition and grain size on the response of the clasts to the stresses producing the deformation. For early descriptions of this locality see Hitchcock (1861) and Van Hise (1896). Van Hise, who did not visit here, apparently did not read Hitchcock as carefully as he should have! For more recent studies see Brace (1955) and Evans and others (1980).

- 37.6 Tyson (formerly Tyson Furnace).
- 38.8 Rescue Lake now on left. Valley cuts upward across section, south of here, to follow the dolomite in the Plymouth Member of the Hoosac.
- 41.3 Junction with Route 103. Turn right (west).
- 41.5 Albite schists of the Hoosac on left. We bear right on side road (the old 103).
- 42.1 Gneiss of Mount Holly to right in brook.
- 42.4 Schists of Mount Holly on left.
- 43.0 Park and turn around.

**STOP 2-11. BUTTERMILK FALLS.** The outcrops on both sides of the bridge are of a quartzite, part of the Mount Holly Complex, that crops out extensively on Okemo Mountain just to the south. The quartzite is highly recrystallized and contains pegmatites and quartz-tourmaline veins. The quartzite strikes northeasterly and intersects the unconformity at the base of the cover sequence near Dry Hill. It is obviously the source of the cobbles and boulders in the conglomerate there. Return to junction with Route 103.

- 44.5 Bear left on Route 103 toward Ludlow.
- 45.6 Quartzitic schists, just below Plymouth Member of Hoosac on right.
- 46.2 Road to Okemo ski area. Continue on Route 103.
- 46.5 Entering Ludlow, turn left.

- 46.6 Traffic light at junction with Route 100 south. Continue straight ahead on 103.  
 47.0 Turn left (north) on Commonwealth Avenue, and ascend through schists of the Pinney Hollow.  
 48.5 Park and obtain landowner permission.

**STOP 2-12. AMPHIBOLITE (CHESTER AMPHIBOLITE) IN PINNEY HOLLOW FORMATION.** The amphibolites and greenstones in the upper part of the Pinney Hollow are less well-developed to the north where we crossed it this morning, and did not crop out there. Just east of the road near the house to the north are carbonaceous schists and black and white quartzites of the Ottauquechee. Immediately west of the road are gametiferous pelites of the Pinney Hollow. The knoll to the west in the pasture is epidote amphibolite. Farther west, downslope in the spruces, are ankeritic greenstones with little or no amphibole. In northern Vermont and southern Quebec equivalent rocks contain relict blueschist assemblages of Taconian vintage (Laird and Albee 1981; Laird and others, 1984; Trzcienski, 1976). These are wiped out southward by a combination, probably, of too long burial and later, superposed, Acadian heating. Turn vehicles around an return to Route 103.

- 50.0 Route 103. Turn left (east) toward Chester.  
 51.0 Schists on left, now assigned to the Hawley, contain black weathering lenses filled with small garnets. These are probably metamorphosed and flattened manganese nodules.  
 51.5 Smithville. "Pinstripe" of the Moretown on left.  
 52.2 Park on right for outcrops on left. Watch traffic!

**STOP 2-13. ULTRAMAFIC ROCKS.** The outcrop across the highway is serpentinite with sharply defined crosscutting zones in which it has been altered to talc-magnesite rock. The rusty-weathering outcrops to the east are of relatively fresh ultramafic rock, altered along joints to serpentinite. This ultramafic body is one of the larger ones in southern Vermont, and was initially dunite and harzburgite (Davidson, 1981). The country rock is Ottauquechee, continuous with that in the mantles of the Chester and Athens domes. This Ottauquechee is in the thickened hinge region of the basement-cored nappe on the west side of the Chester dome.

- 53.9 Waits River on left. We are here in the narrow syncline on the west side of the Chester dome.  
 54.2 Park on right at head of Proctorsville Gulf. Outcrops are across highway. Watch for traffic!

**STOP 2-14. WAITS RIVER FORMATION, PROCTORSVILLE GULF.** Exposure of basal limestones, calcareous schists and gametiferous phyllites of the Waits River Formation on the east limb of the narrow, nearly isoclinal, Proctorsville Syncline. This exposure has been subject to frequent damage due to both natural and human causes. If we are fortunate, we will see gently north plunging, west-side-up folds associated with the later stage of the Acadian orogeny. Highly rotated garnets in the punky-weathering gametiferous calcareous schist show an opposite rotational sense, associated with the earlier Acadian event, on the lower side of a possible westerly extrusion of the Waits River.

### Day Three, Sunday

Drive west and south from Springfield, about 28 miles, via Chester, Grafton and Townshend to the parking lot for the Townshend flood control dam on the south side of Vermont Route 30, about 1.9 miles northwest of Townshend village, toward West Townshend. Road log starts here.

**STOP 3-1. TOWNSHEND DAM.** We will examine the large road cut opposite the parking lot and other exposures about the dam and spillway. The spillway will be entered from a maintenance road at its southeast end after examining the roadcut. We have here a virtually complete section through the slightly overturned units between the marginal phase of the gneisses of the Athens dome and the Moretown. The Hoosac has a few gaps, but the Pinney Hollow and Ottauquechee are in continuous outcrop. The attitude is about N20°E, 78°E. Laird and others (1984) and Karabinos and Laird (1988) have obtained <sup>40</sup>Ar/<sup>39</sup>Ar ages of 376 Ma and 380 Ma for hornblendes here and nearby. The spillway provides the only complete exposure through the Ottauquechee Formation on either the Chester or Athens dome. This locality is noteworthy for its display of snowball garnets in the cut just opposite the bridge crossing the spillway. These show counterclockwise (east side up) rotation of about 180° as one looks northerly. The rotational axes plunge southerly about 20 degrees and lie within the schistosity. These will be seen later in sections parallel to the rotation axis at a point directly below in the spillway beneath the bridge. Contrast the relative constancy of shear sense indicated by the rotated garnets with the variability displayed by the folds. This contrast probably results from the fact that the garnets record an earlier nappe stage of deformation before

development of the highly heterogeneous rheology — reflected in boudinage here — that accompanied the higher temperature metamorphism.

Graphical “unfolding” of the strata and their contained garnets to an approximate pre-doming horizontality orients the rotational axes of the garnets here to about  $N40^{\circ}E$ , almost the same direction as that of the leading edge of a nappe in the Standing Pond Volcanics that now arcs northeasterly across the Guilford dome to the south-southeast as shown by Hepburn and others (1984, Plate 1). This garnet rotational sense around northerly oriented axes, after elimination of the effect of the late deformation, persists laterally in this unit continuously on strike around the south end of the dome (Hepburn and others, p. 93-100). Both the rotated garnets and nappe doubtless reflect a synchronous northwesterly flow. The same garnet rotational sense around northerly oriented axes, after elimination of the effect of the late deformation, persists laterally in this unit continuously on strike around the south end of the dome (Hepburn and others, p. 93-100). This contrasts with a better documented and convergent west-southwesterly flow a few miles north, at a different structural level (Rosenfeld, 1968, p. 199-200; p. 193). Convergent flow may explain the deformation of Pebbles in a conglomerate near the top of the Hawley on Windmill Mountain to the east (Rosenfeld, 1968, p. 197-198). These pebbles are greatly elongated in the plane of the schistosity parallel to the possible direction of extrusion of the overlying Waits River and are shortened, also in the plane of schistosity, in a direction at right angles to the elongation.

An “oligoclase isograd” can be located in matrix assemblages at lower grades, some distance northwest of Townshend Dam, which is located close to the staurolite isograd. Though not of low enough variance to be uni- or divariant the isograd corresponds to the whole rock reaction: muscovite + chlorite + clinozoisite + albite to garnet + biotite + oligoclase with quartz to balance. In matrix assemblages the assemblage on the left occurs only to the northwest of the isograd. It is interesting that we can see coexisting albite, oligoclase, and clinozoisite here as a relict assemblage preserved in the garnet interiors but not outside where there is a weakly zoned oligoclase. The contrast between these relict and matrix assemblages would seem to be good evidence of growth of the garnets over a considerable range of temperatures.

We then enter the spillway and pass through plagioclase-rich gneisses and amphibolites of the Hoosac Formation into the interbedded schists and amphibolites of the Pinney Hollow Formation. On the south side of the spillway in that unit we will observe a rusty-weathering paragonite schist (formerly misidentified as Ottaquechee) having large garnets up to 3 cm in diameter. Some of these are the first to have had their growth duration measured (Christensen and others, 1989). They took about 10.5 million years to grow as determined from rubidium-strontium decay. Precision is about 4.2 million years. One of these garnets, having the characteristic snowball pattern with rotation of almost 230 degrees, thus gives some idea of the strain-rate during prograde metamorphism.

Chamberlain and Conrad (1992; 1993) also used this exposure in their study of the zoning of oxygen isotope ratios in garnets. The utility of the method of Christensen and others (1989) rests largely on the assumption that the evolution of the  $87/86$  ratio of strontium in the matrix during garnet growth is due solely to decay of Rb and is not influenced by a transport of Sr by infiltrating fluids. It is, therefore, of interest to determine whether their outcrops were infiltrated by fluids during metamorphism and, if so, whether the Rb-Sr systematics were affected by the infiltration. With this in mind, Chamberlain and Conrad used the carbon dioxide laser extraction system at Dartmouth to determine if the garnets were zoned in oxygen isotopes. Their studies show that the garnets are indeed strongly zoned in oxygen isotopes. The magnitude and nature of the isotopic zoning depends upon the garnet's location in the outcrop studied. In the schists of the Pinney Hollow Formation “delta O-18” is about 9 parts per thousand in contrast to values of 12.5 parts per thousand in the schists of the adjacent Ottaquechee Formation. Garnets from the Pinney Hollow within 10 meters of the contact with the Ottaquechee have relatively homogeneous values ranging from 9.5 parts per thousand in the core to 10.5 parts per thousand in the rim. Garnets in the Pinney Hollow 85 meters from the contact are more strongly zoned, with values ranging from a low of about 6.0 parts per thousand in the cores to a high of about 9.0 parts per thousand in the rims. These zoning patterns were produced by continuous infiltration of relatively high delta O-18 waters derived from the subjacent schists into the paragonitic schists during garnet zone metamorphism.

It is possible to determine the time-integrated fluid fluxes by comparison of observed isotopic zoning profiles in garnet with those calculated from the equation describing combined advection-diffusion of a tracer. Using this method, Chamberlain and Conrad calculated time-integrated fluid fluxes that have several interesting implications. First, using these fluid fluxes they estimated that Sr was transported 1.5 cm to 1.5 m during fluid flow. Therefore, the assumption that the matrix is closed to Sr during metamorphism (see Christensen et al., 1989) is a reasonable

approximation for most of the rocks at Townshend Dam. Second, the fluid fluxes that we calculate are consistent with the fluxes that Ferry (1992) determined from petrologic studies of rocks adjacent to the Chester and Athens domes. Ferry (1992) proposed that the gneiss domes were centers of regional-scale hydrothermal systems during the Acadian metamorphism. The garnet isotopic data, in part, support this hypothesis and suggest that flow was directed toward structurally deeper rocks now exposed in the cores of the gneiss domes.

Having finished with the Pinney Hollow garnets we may now cross into Ottauquechee Formation and observe the zone containing snowball garnets that we saw earlier in the highway cut. Note that the sections through some of them are parallel to the rotational axis, and may thus be compared with those observed in equivalent orientation in the Standing Pond Volcanics at the first day stop on the north bank of the Williams River. *Please — don't spoil this exposure by hammering on it; exposed axial sections are hard to find!*

Farther up the spillway, at the Ottauquechee-Moretown contact, are small boudins of ultramafic material, now talc-actinolite rock. When done please return promptly to the vehicles, and proceed southeasterly on Route 30, there is much more to see today!

- 0.4 Scott covered bridge on right, amphibolite of the border phase of the dome gneisses on left. These rocks may be correlative, wholly or in part with the Tyson Formation of the eastern Green Mountains.
- 0.7 Park with care.

**STOP 3-2. CONGLOMERATE GNEISS AND AUGEN GNEISS.** The highly deformed conglomerate gneiss is in contact with an augen gneiss (Bull Hill Gneiss). Both are parts of the border phase of the gneisses of the Chester and Athens domes. Recent work on zircons of the Bull Hill Gneiss (Karabinos and Aleinikoff, 1990) yields a post-Grenville age of about 960 Ma. Because of the granitic composition of the Bull Hill gneiss and its tabular, conformable nature, it is possible that it represents a metamorphosed stack of rhyolitic ignimbrites. In the southern part of the Athens dome, it has not been possible to delineate accurately the boundary between the Bull Hill gneiss and what are believed to be older, but lithically similar, Precambrian granitic augen and flaser gneisses in the core of the dome. Note counterclockwise folds in the gneiss, possibly a result of buoyant upthrusting of the gneissic core of the dome.

- 1.9 Townshend. Turn left onto Route 35 and proceed northerly.
- 2.3 Magnetite-bearing, flaser gneisses to west across field. Continue north through a variety of gneisses.
- 3.8 Simpsonville.
- 5.3 Bear left off Route 35 onto the Grafton road.
- 5.6 Park on right.

**STOP 3-3. GRAPHITIC CALC-SILICATE ROCKS IN CORE OF ATHENS DOME.** Road cut on northeast side of Townshend-Grafton road northwest of intersection with Route 35. Calc-silicate rocks, characterized by coarse graphite flakes and pyrrhotite, now largely altered to a rusty, yellowish, odoriferously sulfurous mess. Mineral assemblages include diopside-zoisite-microcline-plagioclase-biotite-tremolite-sphene-calcite-graphite and altered pyrrhotite. The strata strike northeasterly. These rocks are similar to units in the Grenville of the southeastern Adirondacks, and also to rocks to the west in the core of the Green Mountains, and in the Berkshires farther south.

- 6.5 Our ascent since Stop 3-3 has brought us into a narrow valley that we shall follow north-northeasterly toward Grafton. The course of the valley is probably determined in large part by dolomites along the axial trace of the Hawks Mountain mulde, for which we will see some evidence at Stop3-4, farther north, between Grafton and Chester. Dip slopes on east side of valley are mainly Bull Hill Gneiss.
- 12.0 Grafton. Turn right on Route 121. Grafton is a carefully preserved Vermont village. "Funny Farm" starring Chevy Chase was recently shot here.
- 12.1 Route 35. Turn left (north) toward Chester.
- 13.8 Dolomite of Tyson on left, dipping west beneath albite schists of the Hoosac. The valley here and on to the north is aligned with that south of Grafton.
- 16.3 Enter Grafton Gulf, a narrow defile parallel to strike. The dip slopes to the east are Bull Hill Gneiss. The west side of the gulf has overhanging ledges of albite schist of the Hoosac and slabby gneisses typical of many in the border phase of the domes.
- 16.8 Chester-Grafton town line.

16.9 Park.

**STOP 3-4. DOLOMITE IN GRAFTON GULF.** A pillar of dolomitic marble here supports an overhanging ledge of albite schist. An interesting feature of the dolomite is the presence of both oligoclase and microcline, the latter largely encased in a selvage of phlogopite. Solution pits indicate that much of the calcite present has been dissolved and otherwise reacted with water and suggests why the dolomite is characterized by a topographic groove. We explain the relatively smooth dips of the rocks flanking the dolomite as a result of its ease of deformation and thus its "lubricating" quality. Strain was undoubtedly concentrated in the dolomite. This also has practical applications. The large flagstones used for construction in this part of Vermont (a fine display will be seen later today in Chester) are associated with this zone.

17.4 Summit of Grafton Gulf. From here the highway cuts northeasterly into the core of the Chester dome.  
 19.4 Chester. Turn right on Route 11.  
 19.9 Intersection. Bear left on Route 11.  
 22.7 Park.

**STOP 3-5. GNEISSES OF CHESTER DOME.** Core of dome is mainly two-mica gneiss with minor amphibolite, a few wisps of mica schist and quartzite, and some small ultramafic bodies (not shown on Plate I). Gneisses here are cut by pegmatites containing large crystals of epidote.

23.5 Turn right on Pleasant Valley Road.  
 23.8 Turn left on Breezy Hill Road.  
 24.0 Border phase gneisses on left.  
 25.1 Turn around and park.

**STOP 3-6. MANTLE SEQUENCE ON EAST SIDE OF CHESTER DOME.** Rocks on the Stellafane Observatory grounds are garnetiferous schists and amphibolites of the Pinney Hollow. Feldspathic schists of the Hoosac crop out in the woods to the west. The carbonaceous schists and black and white quartzites in the fields to the east belong to the Ottauquechee. Schists of the Moretown and Barnard Gneiss can be seen on the knoll to the southeast. We are here on the southeast side of the Chester dome, whereas at Townshend Dam we saw a similar section on the southwest side of the Athens dome. Return via Breezy Hill Road and Pleasant Valley Roads to Route 11.

26.7 Turn left (west) on Route 11.  
 29.8 Turn right on unpaved road toward North Chester.  
 30.6 Junction with Route 103. Proceed straight ahead on Route 103 through the "Stone Village" of North Chester. Many of the houses here, one a converted school, and also a church, are made of the slabby phengitic gneisses characteristic of the dome margins.  
 32.7 Low outcrops of gneisses and amphibolites of Chester dome on right for next half mile.  
 33.8 Start of road cut on right in core gneisses of the Chester dome.  
 34.6 Gassetts. Bear left on Route 103.  
 34.9 Gneisses on right in border phase of Chester dome. Covered zone just beyond is at position of marbles on east side of Hawks Mountain mulde.  
 35.1 Start of 0.2 mile cut through Hoosac Formation in axial region of Hawks Mountain mulde. The schists, initially albitic, pick up garnet and then paragonite, staurolite, and kyanite toward the north end of the cut. These last are the "Gassetts Schist" beloved by metamorphic petrologists (Thompson and others, 1977a,b; Eugster and others, 1972). We will not stop here now, but do not be disappointed. We will see the same sequence later, in more revealing natural exposures at Stop 3-8.  
 35.4 Park on right beyond bridge, and walk to small quarry south of the highway.

**STOP 3-7. SLABBY GNEISS, MARBLE AND CALC-SILICATE.** The slabby, phengitic gneisses, some of them augen gneisses, were quarried to build the new bank, post office, and Catholic church in Chester. Of more interest are the marbles and calc-silicates, that have been described in some detail by Alan Thompson (1975). Minerals of interest include diopside, tremolite, clinzoisite and phlogopite. We have now crossed the Hawks Mountain mulde, a southeasterly-closing, eyelid-shaped mass of Hoosac schist across the north-central part of the Chester dome, outlined by marbles and calc-silicates of the Tyson.

- 35.7 Turn sharply right on side road that leaves Route 103 and soon swings left to proceed north along the marble valley on the west side of the Hawks Mountain mulde. On the left are slabby gneisses. The ridge on the right is Hoosac (Gassetts) schist.
- 36.6 Turn right (east) on side road, then turn around and park just west of a small wooden bridge.

**STOP 3-8. HOOSAC (GASSETTS) SCHIST.** Walk north, just east of bridge, ascending southwest ridge of Hawks Mountain to east-west power line about 300 ft. up. At crest is a view southeast over the lowlands in the central portion of the Chester dome. The rocks immediately below are albite schists, with gneisses on the flats beyond the base of the steep slope. The marbles and calc-silicates between are again covered here, but can be seen farther east along the break in slope at the base of the mountain. Westward along the power line the schists are increasingly garnetiferous, culminating in the *piece de resistance*, the "high roller" outcrop, about two thirds of the way down. The essence of the assemblage is quartz, muscovite, paragonite, garnet (snowballed), staurolite, and kyanite. Note the unconformities separating the outer rims from the cores of the garnets! The slot below to the west is in the marble and calc-silicate zone on the west side of the mulde. Return to vehicles and rejoin north-south road.

- 36.8 Turn right (north) toward Duttonsville Gulf.
- 37.2 Small body of two-mica granitic rock on right.
- 37.8 Bear right at fork.
- 38.2 Entering narrow part of gulf. Slabby gneisses on left, Hoosac schists on right.
- 38.4 Narrow underpass, slabby gneisses overlying marbles (inverted section) on left just beyond.
- 38.7 Right bend. Slabby gneisses ahead overlie marble in small quarry just west of railroad tracks. Road then turns left again, cutting into the Hoosac for about 0.5 mi.
- 39.2 Abandoned dolomite quarries through trees on left. Odd stone structure is a lime kiln, not a relic of the early Vikings. Road turns left just beyond.
- 39.7 Turn right (north) at intersection, passing through dangerous underpass. Continue across Black River.
- 39.8 Cavendish. Turn right on Route 131, passing more slabstone houses, some with gingerbread, also a stone church on left. (Alexander Solzhenitsyn lives in the north part of town.)
- 40.4 Turn right on unpaved side road.
- 40.6 Turn around and park at power station.

**STOP 3-9. CAVENDISH GORGE.** Exposures in gorge on west side of Black River, just upstream from the power station, show contacts between schists to the east and marbles, calc-silicates, and slabby gneisses to the west. Be careful, these rock are very slippery when wet! Note magnetite zone (the terra rossa) at Hoosac-Tyson contact, also folding in calc-silicates and associated quartzites. Return to highway.

- 40.9 Turn right on Route 131.
- 41.4 Whitesville. Turn right on side road, then turn around and park near red house just west of bridge.

**STOP 3-10. RECUMBENT FOLD AND REACTION SKARNS.** With permission of occupant, walk north from house to projecting ledge on bank of river.. Ledge is albite schist with abundant magnetite, enclosing the nose of a fold in dolomite. The contact is here marked by a reaction skarn producing a layer of diopside next to the marble and actinolite next to the schist. Ledges in river are garnetiferous albite schists of the Hoosac.

- 41.5 Turn right (north) on Route 131.
- 42.1 Staurolite schist of Hoosac. The marbles pass through a saddle in the hill to the west.
- 42.2 Marbles and calc-silicates have returned east and appear in bed of black river to right.
- 42.5 A tributary brook enters Black River. Turn left on side road (Hardy Hill Road) immediately beyond.
- 43.2 Turn right on Field Hill Road, then left immediately, just beyond Crown Point Road marker.
- 43.9 Bear right at fork.
- 44.2 Coarse garnet schist on left for about 0.2 miles.
- 44.6 Marble below slabby gneiss on left.
- 44.9 Turn around at road corner and return south.
- 45.3 Park near marble outcrops on right.

**STOP 3-11. GASSETTS SCHIST ON STAR HILL.** This is a no hammer stop! Walk northeasterly over exposures of coarse garnet schist with abundant staurolite and kyanite to cliff top (be careful!) with

fine display of the schist. Staurolite and kyanite are here in epitaxial intergrowths with (010) of staurolite against (100) of kyanite. Ledges are broken by collapse over marbles beneath. Then walk west to road, just west of which is an exposure of the marbles and calc silicates, and overlying slabby gneisses on the top side of this mulde (not the Hawks Mountain mulde). The sequence is repeated again, but with calc-silicates very thin, some distance up the hill to the west. A detailed study of the Star Hill area has recently been made by E. A. Downie (1982). The detailed structure is a multiply folded mulde at so fine a scale that it had to be generalized on Plate I. Return to Route 131.

- 47.6 Turn left (east) on Route 131, calc-silicate and marble on right in river, schists on slopes beyond.
- 47.8 Slabby gneisses on left. These overlie (inverted) the marbles in the river.
- 48.3 Intersection. Bear right on Route 131.
- 49.3 Park for outcrop on left.

**STOP 3-12. GASSETTS SCHIST AT BLACK RIVER.** We are now back in the Hawks Mountain mulde. The essence of the assemblage is quartz, muscovite, paragonite, garnet, staurolite, and kyanite (best seen in woods at north end of outcrop). The garnets are highly snowballed but this is best seen in an appropriately cut thin section owing to the nature of the exposure. To see snowball garnets clearly on natural outcrops in the field requires a glaciated outcrop providing central sections of the garnets, and oriented so that the rotation axis is roughly normal to the surface of the outcrop. A garnet from this locality has been studied isotopically by Chamberlain and Conrad (1991), and one from either this or a nearby locality has been studied by Young and Rumble (1993). Both studies show delta O-18 values decreasing from core to rim.

This is the end of the excursion. Route 131 crosses Vermont Route 106 about two miles east, and passes under Interstate Highway I-91 at an interchange 7 miles beyond. Route 131 crosses the Connecticut River just east of I-91 and becomes New Hampshire Route 103. The shortest route to Boston is to follow Route 103 via Claremont and Newport, New Hampshire to an interchange with I-89 near Warner (about 36 miles). Warner, New Hampshire is about two hours from Boston via I-89 and I-93.

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Chapter C

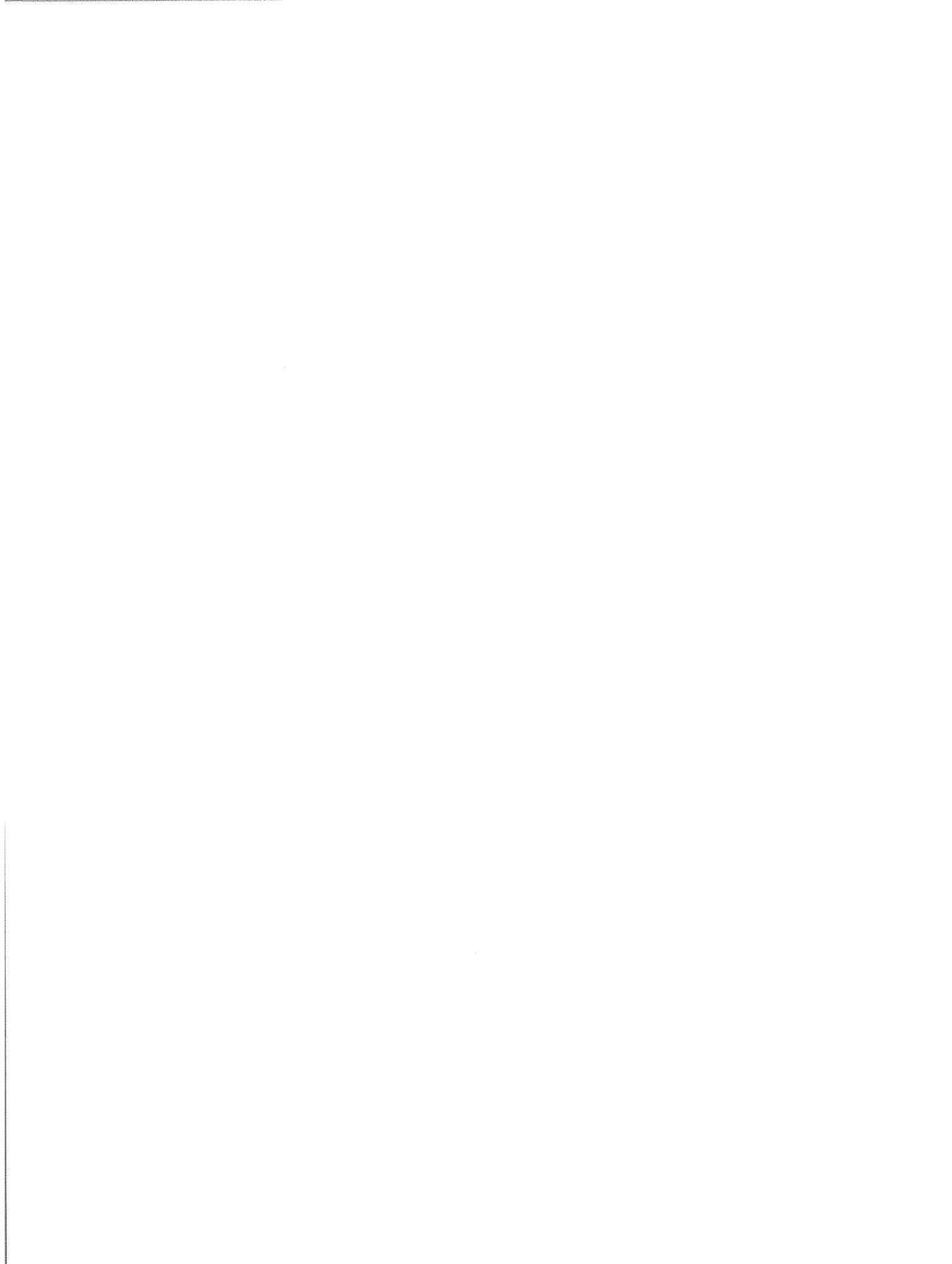
# A Tectonic-Stratigraphic Transect Across the New England Caledonides of Massachusetts

By Peter Robinson, Nicholas M. Ratcliffe, and J. Christopher  
Hepburn

Field Trip Guidebook for the Northeastern United States:  
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## A TECTONIC-STRATIGRAPHIC TRANSECT ACROSS THE NEW ENGLAND CALEDONIDES OF MASSACHUSETTS

led by  
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Nicholas M. Ratcliffe, U. S. Geological Survey, Reston  
J. Christopher Hepburn, Boston College

### INTRODUCTION

The State of Massachusetts and adjacent New York contain a nearly complete section across the northern Appalachians (Figure 0.1). From the Prudential Center in Boston to Mount Greylock (3491') in the northwestern part of the State is a distance of 178 km. On a clear day both of these points can be seen from the summit of Mount Wachusett (2006') in the middle part of the State. This is a distance equivalent to that from Como south of the Alpine suture to Zurich in the Swiss Plain. As compared to northern New England, where the distance across the orogen is more than 450 km, in southern New England the belt narrows to barely 200 km. As might be expected with such a drastic narrowing, this is accompanied by a great increase in the intensity of tectonism, plutonism and metamorphism. This narrowing also makes it possible to examine a representative selection of highly varied rock types and tectono-stratigraphic terranes, within the scope of a single three-day field trip and small geographic range.

This trip takes advantage of the 1983 Bedrock Geologic Map of Massachusetts (Zen et al.; 1983). The map compilation team organized by E-an Zen contained a group of geologists with various interests and very different geographic backgrounds in the State, who shared their expertise and worked hard on the various compromises needed to portray such diverse geology on a single map sheet. The exercise helped to place the State's geology within a broader tectonic framework as outlined by Hatch et al. (1984), and set the stage for additional work that has been in progress for nearly a decade, producing several significant revisions based on stratigraphic relations, on new radiometric ages, and on quantitative petrology. The trip has been organized by two of the four original compilers, Ratcliffe and Robinson, and by Hepburn, who made significant individual contributions to the map and has since been active in the region.

In its broadest outlines Massachusetts appears to be composed from rocks related to three tectonic plates active in the early and middle Paleozoic, and its structural and metamorphic features are mainly the product of three orogenic events, the Late Ordovician Taconian orogeny, the Devonian Acadian orogeny, and the Pennsylvanian-Permian Alleghanian orogeny. These features were then modified during Mesozoic extension associated with the opening of the modern Atlantic, involving faulting with vertical displacements as great as 8 km.

The western part of the State consists of rocks formed on or near the margins of Laurentia in Proterozoic through Late Ordovician time. Ample evidence indicates that Laurentia occupied an equatorial position in the early Paleozoic. Our trip begins here, although this choice is more the result of logistical convenience than scientific choice. Some of us would rather have begun at the east end and worked toward the North American craton to avoid the impression of North American chauvinism that is induced conceptually by the particular location of the Atlantic as a result of Mesozoic continental rifting.

The eastern part of the State consists of rocks formed on or near the Avalon plate. In earlier work it was supposed that this mass collided with Laurentia during the Ordovician Taconian orogeny. However, more recent paleomagnetic work as well as lithostratigraphic work suggests it was part of Gondwana which remained in high southern latitudes through the Ordovician and only arrived at the latitude of Laurentia in the Devonian.

The central part of the State contains rocks formed on or near "craton X" as described by Zen (1983). This contains the Ordovician volcanic and intrusive rocks of the Bronson Hill magmatic arc as well as late Proterozoic rocks of general Avalon affinity, but uncertain structural position. The closing of an ocean between craton X and Laurentia and its collision with Laurentia are generally associated with the Late Ordovician Taconian orogeny. The closing of a second sedimentary and oceanic belt between the amalgamated Laurentia - craton X and the Avalon plate is generally associated with the Devonian Acadian orogeny. A number of structural and metamorphic events in the late Paleozoic are thought to be related to the collision of Africa with the amalgamated Laurentia-craton X- Avalon. Most, but not all, Appalachian geologists believe this last collision zone lay outside the New England area, though its effects were strongly felt in local areas.

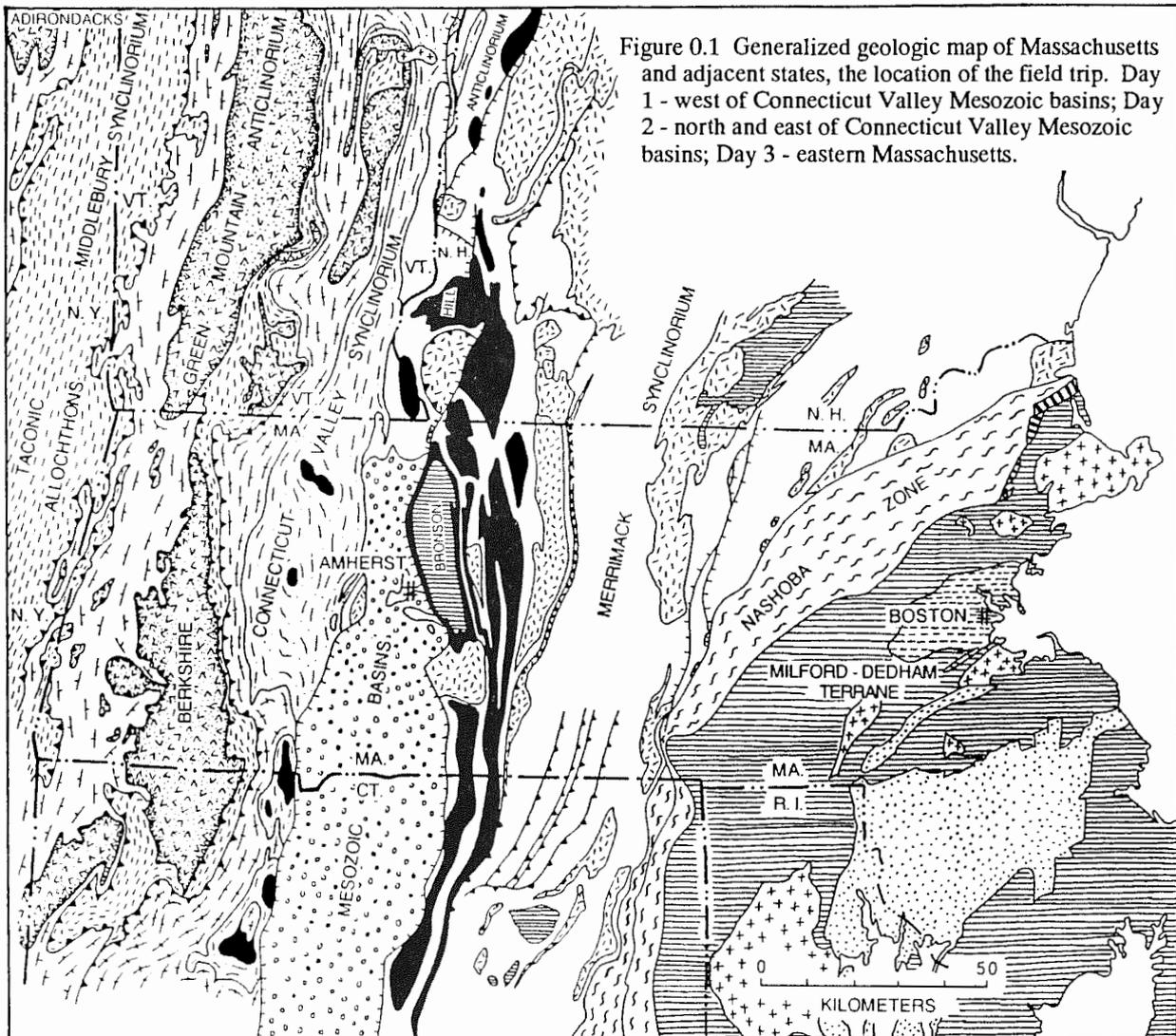


Figure 0.1 Generalized geologic map of Massachusetts and adjacent states, the location of the field trip. Day 1 - west of Connecticut Valley Mesozoic basins; Day 2 - north and east of Connecticut Valley Mesozoic basins; Day 3 - eastern Massachusetts.

**Key to patterns -- Day 1, western Massachusetts:** Laurentian Middle Proterozoic basement - dots and v's; Cambrian-Ordovician carbonate bank and shelf - dip symbols; Cambrian-Ordovician Taconic facies, Hoosac Formation and lateral equivalents - fine dashes; Cambrian-Ordovician accretionary prism - crossed dashes; Gneisses of magmatic arc in cores of domes - black; Silurian-Devonian Vermont sequence - dash-dot. **Day 2, central Massachusetts:** Late Proterozoic in Pelham dome, Massabesic Gneiss in N. H., and Willimantic dome in Ct. - horizontal ruling; Ordovician plagioclase gneisses in cores of domes - black; Ordovician, Silurian and Lower Devonian of Bronson Hill anticlinorium and Merrimack synclinorium - not patterned; Late Ordovician to Pennsylvanian intrusives - random dashes; Triassic-Jurassic sedimentary rocks and basalts. **Day 3, eastern Massachusetts:** Late Proterozoic of eastern New England - horizontal ruling; Late Proterozoic of Boston basin - dashed horizontal ruling; Ordovician to Devonian alkalic plutons - crosses; Nashoba zone - curvy lines; Late Silurian to Early Devonian Newbury Volcanics - dark cross stripes; Pennsylvanian Narragansett, Norfolk, and Worcester basins - dots; Triassic of Essex County basin - circle.

Our first day in the field will be mainly concerned with features formed on the Laurentian margin. The principal structural effects are those of the Taconian orogeny, though as we work eastward there will be a progressive increase in Acadian tectonic and metamorphic effects. The principal tectono-stratigraphic units, described in a more or less tectonically upward progression include the following:

- 1) North American Grenvillian and post-Grenvillian basement, which we will see only in its highly deformed state in the Berkshire thrust sheets and not in its pristine state as in the nearby Adirondacks;

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- 2) The Laurentian eo-Cambrian, Cambrian, Lower Ordovician passive margin sequence consisting of early rift clastics, earliest Cambrian littoral sands, and the Cambrian-Ordovician carbonate bank and back reef facies;
- 3) Unconformably overlying Late Ordovician graywacke-shale sequence of the Taconian foreland basin;
- 4) Eo-Cambrian, Cambrian, Lower Ordovician continental slope-rise clastics contemporaneously deposited oceanward of 2) but now largely preserved in the Taconian allochthons thrust over previous units;
- 5) Metamorphosed clastic rocks similar in part to 4) but characterized by abundant rift related volcanics and considered to have been deposited on thinned Laurentian continental crust or oceanic crust. Based on analogies with adjacent Quebec, young oceanic crust and mantle were obducted onto this sequence in the Late Cambrian to Earliest Ordovician, and are now preserved in Massachusetts as deformed lenses of mafic and ultramafic rocks;
- 6) Vestiges of fore-arc basin deposits and volcanics formed in front of the overriding Bronson Hill magmatic arc and now constituting the uppermost part of the Taconian accretionary prism;
- 7) Gneisses constituting the leading edge of the Bronson Hill arc exposed within Acadian gneiss domes;
- 6) The Silurian-Lower Devonian sequence of the Connecticut Valley belt that overlaps the Taconian suture.

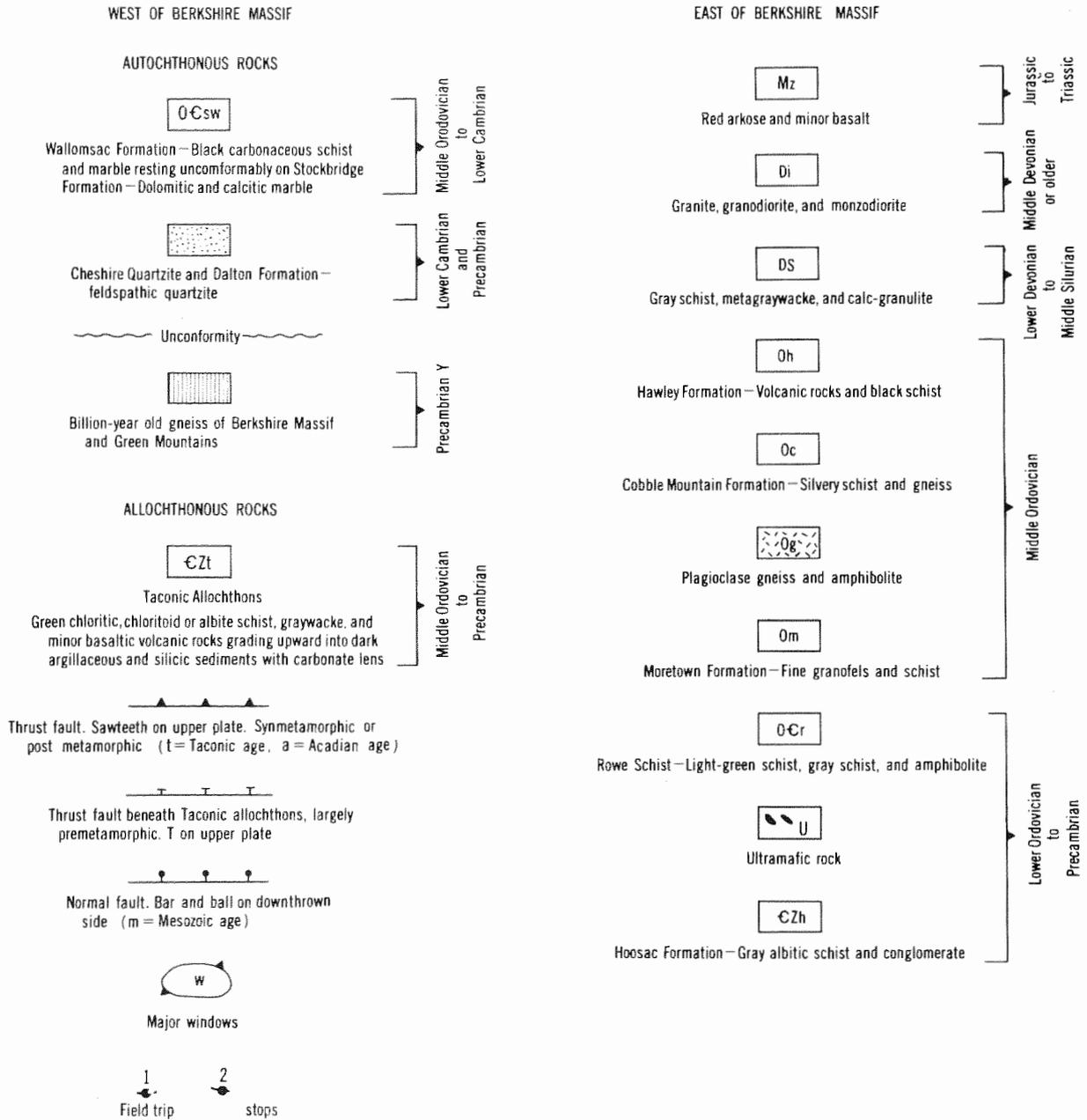
Our second day in the field will be concerned with features formed on craton X and its Ordovician and Silurian-Devonian cover. Being on the upper plate of the Taconian collision, Taconian effects aside from magmatism and volcanism are restricted to modest unconformities. The dominant effects are Acadian, ranging from fold and thrust nappes to granulite facies metamorphism, but we will also examine the results of newly discovered Pennsylvanian ductile deformation. The major tectono-stratigraphic units from deepest to shallowest include the following:

- A) Late Proterozoic (613 Ma) alkalic igneous rocks and quartzites of Avalon affinities with controversial contact relations with overlying rocks;
- B) A Late Ordovician (455-443 Ma) intrusive complex of felsic to mafic calc-alkaline rocks that may be the roots of a magmatic arc,
- C) A Late Ordovician cover sequence of tholeiitic arc volcanics ranging from basalts through andesites to rhyolites and overlying black shales;
- D) A Silurian-Devonian cover in which the Silurian is in two facies, a thin western conglomerate and calcareous facies overlying the ruins of the Taconian orogen, and a thicker eastern shale-graywacke facies forming continental slope-rise deposits on the post-Taconian North American margin.

The third day will be devoted to fault systems and strata at the junction between craton X and Avalon, and to the characteristics of the Boston Avalon zone. Critical to Acadian plate interpretations is the contrast between the Silurian continental slope-rise clastics of paleo-North America and the poorly exposed fossiliferous Late Silurian - Early Devonian coastal volcanics of arc affinities attached to Avalon. The fault-bounded Nashoba terrane of Ordovician or older rocks has been variously assigned to craton X or Avalon. Once on Avalon proper, the sequence includes:

- I) Late Proterozoic stratified rocks and volcanics,
- II) Cross-cutting Late Proterozoic granites,
- III) Vendian spore-bearing strata of the Boston basin,
- IV) Early and Middle Cambrian shelf strata with Baltic faunal affinities,
- V) Late Ordovician through Devonian alkalic plutons,
- VI) Pennsylvanian continental clastic basins.

EXPLANATION



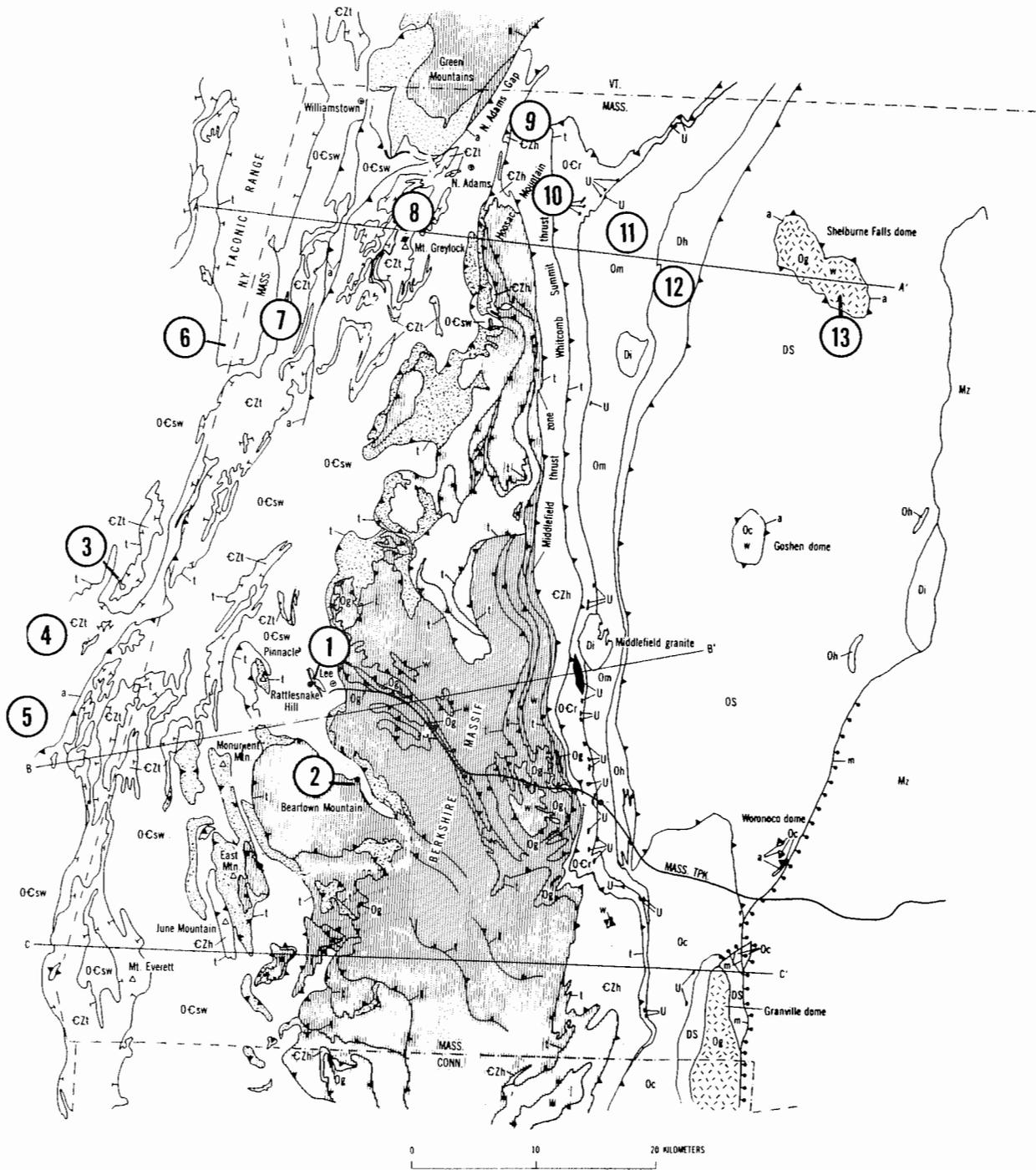


Figure 1.1. Generalized geologic map of western Massachusetts showing field trip stops and section lines for figure 1.3.

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**DAY 1: THE TACONIDE ZONE OF WESTERN MASSACHUSETTS**

by Nicholas M. Ratcliffe

**INTRODUCTION**

Modern plate-tectonic interpretations (Stanley and Ratcliffe, 1985) of the Taconide Zone of western New England stem in large part from a program of systematic mapping conducted in western Massachusetts from 1960 to 1980 which culminated in the publishing of the Bedrock Geologic Map of Massachusetts; 1:24,000 geologic maps of nearly all of the area of the trip are available (see list of references in Zen et. al., 1983). This work was conducted principally by Donald B. Potter, Rolfe Stanley, David Harwood, E-an Zen, and Norman L. Hatch, Jr. and Nicholas Ratcliffe. Since the publication of the State map, studies of the regional metamorphism and geochronology by Sutter et. al. (1985) and by Hames et. al. (1991) have helped define the extent and timing of the Taconic metamorphism and of the subsequent Acadian remetamorphism. The general content of the trip on Day 1 follows closely a field trip lead by Norm L. Hatch, Jr. and myself (Ratcliffe and Hatch, 1979) across this same area. The extensive work of Hatch and of Rolfe Stanley in the Rowe-Hawley zone (Hatch and Stanley, 1988; Stanley and Hatch, 1988) forms the basis for the discussion of that area.

The newly published geologic map of the Williamstown area (Ratcliffe et. al., 1993) will be relied on extensively to complement the text discussion of the stops on Mount Greylock and in the Hoosac Range. The work and cooperation of my colleagues Don Potter and Rolfe Stanley are gratefully acknowledged.

**LITHOTECTONIC UNITS**

The bedrock of western Massachusetts (fig. 1.1) belongs to four major lithotectonic units (1) the parautochthonous basement gneisses of the Berkshire massif (1.3 to 1 By) and their Late Proterozoic and early Cambrian clastic cover (Middle Proterozoic gneisses and the Dalton Formation and Cheshire Quartzite); (2) the autochthonous Lower Cambrian through Middle Ordovician shelf carbonate sequence (Dalton Formation, Stockbridge Formation and exogeoclinal Walloomsac Formation); (3) the composite slices of the Taconic allochthon (Biddings Brook, Chatham, Everett and Greylock slices); and (4) the rocks of the eugeoclinal belt Rowe-Hawley zone exposed east of the Berkshire massif.

The distribution and origin of these rocks is the result of the late Proterozoic and Cambrian rifting of the Laurentian margin and subsequence Cambrian? through Middle Ordovician closure of the Iapetus Ocean.

Despite the fact the entire area has structural and metamorphic overprint of Acadian age the Acadian structures and remetamorphism is weak enough in westernmost Massachusetts to clearly discern the Taconic events and structures.

**SEQUENCE OF TACONIC TECTONIC EVENTS**

Based on detailed geologic mapping (references in Zen et. al., 1983), geochronology (Sutter et. al., 1985) following sequence of events in Taconic orogeny have been established, following the interpretations of Stanley and Ratcliffe (1985). (1) Late Proterozoic to Early Cambrian rifting and deposition come clastic and basaltic volcanics in the Chatham and Rensselaer Plateau slices of the Taconic allochthon, and basaltic volcanics in the Hoosac Formation in the basal cover of the Berkshire massif (Ratcliffe, 1987); (2) deposition of a westwardly transgressing pre-shelf clastic section of the Hoosac Formation, Dalton Formation and Cheshire Quartzite; (3) building of the shallow water carbonate shelf and contemporaneous deposition of slope and rise sediments of the lower Taconic slices; (4) formation of an eastern accretionary wedge and oceanic tectonic mélange in the Rowe-Hawley zone, and contemporary volcanic arc of the Bronson Hill; (5) emplacement of the Taconic

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allochthons (Taconic  $D_1$  deformation) first as nonmetamorphic but hard rock slices was followed by increasingly more metamorphic thrust slices (Taconic  $D_2$  metamorphic structures) as deformation moved onto the shelf; and (6) widespread imbrication of the Taconic slices, and basement rocks of the old cratonic edge as the Berkshire massif was emplaced as a series of nested sialic flakes ( $D_3$  Taconic deformation).

The climactic Taconic  $D_3$  event produced a distinctive and widespread fold-thrust fabric (Ratcliffe and Harwood, 1979) that is recognizable from central Vermont to as far south as Dutchess County, New York (Stanley and Ratcliffe, 1985; Ratcliffe, 1992). Everywhere this event was synchronous with attainment of the maximum grade of Taconic Barrovian metamorphism, and extensive tectonic shortening of the old Laurentian margin. In the north, this imbrication is responsible for duplication along the eastern margin of the Green Mountain massif where thrust motion was  $N.65^\circ W$ . The  $D_3$ -imbricated tectonic cover in Vermont is folded over the Chester-Athens domes east of the Green Mountains, as most of the  $D_3$  strain of the Berkshire massif in Massachusetts is concentrated east of the Green Mountain massif in Vermont (Stanley and Ratcliffe, 1985). In Massachusetts wholesale transport of the Berkshire massif and its composite slivers of basement moved due W to  $S.75^\circ W$ ., and in southern New York  $S.70^\circ$  to  $65^\circ W$ . where steeply southeast dipping  $D_3$  faults are developed. Resolved motion indicates a component of left-slip in Vermont and right-slip in New York. This pattern suggests that the final Taconic dynamothermal event ( $D_3$ ) produced final destruction of an eastward facing promontory in the ancient Laurentian margin which was centered in the location of the present Berkshire massif, as the compliment to the Quebec reentrant. The geology of the Berkshire massif and western Massachusetts in general is the result of the Ordovician destruction of this promontory.

### STRATIGRAPHY

Stratigraphic columns are given in figure 1.2. Middle Proterozoic gneisses in the Green Mountain and Berkshire massifs range from 1.3 b.y. to 1 b.y. based on U-Pb zircon studies (Ratcliffe et. al., 1992). In both areas hornblende-granulite grade or higher metamorphism and deformation preceded intrusion of coarse-grained post-tectonic rapakivi granites dated by U-Pb zircon studies at about 965 m.y. (Karabinos and Aleinikoff, 1990). It is certain therefore, that the core gneisses of the Berkshire massif and Green Mountains were affected by the Grenville orogeny and were originally part of Laurentia and not exotic sialic flakes. These formed the basin floor for deposition of Late Proterozoic to Lower Cambrian clastic rocks of the Dalton-Cheshire and Hoosac sequence. Lower Cambrian through Lower Ordovician shelf-carbonate rocks of the Stockbridge Formation were built up on the western part of this pre-shelf sequence. Clasts of Middle Proterozoic gneisses in the oldest units of the Taconic allochthon, suggest that these units were derived from basement like that of the Berkshire massif.

Paleozoic cover rocks are divided into two sequences, a western shelf sequence and an eastern more eugeoclinal belt (fig. 1.2). The western sequence consists of shallow-water quartz-rich clastic (Dalton and Cheshire) and carbonate rocks (Stockbridge Formation) totaling 1,000-1,600 m in thickness. A Middle Ordovician carbonaceous phyllite and limestone sequence, the Walloomsac Formation, unconformably overlies the Stockbridge. Both these units are autochthonous with respect to the Green Mountain basement. Faulted slices of green and gray aluminous phyllite, coarse graywackes, and minor basaltic volcanic rocks of the Taconic allochthon overlie the autochthon discordantly. Meager fossil evidence and regional correlation of lithofacies suggest that fault slices consists of rocks ranging in age from late Precambrian to Early Ordovician. Multiple overlapping slices are recognized (Zen, 1967), with the higher more easterly slices overlapping the trailing edges of more westerly slices. It is widely held that the oldest of these slices were emplaced on near surface thrusts but that the later higher slices were emplaced on deep synmetamorphic thrusts faults. The original site of deposition of what are now the Taconic allochthonous rocks was not on the exposed portions of the Berkshire massif, but at some point to the east of the restored position of the Berkshire massif.

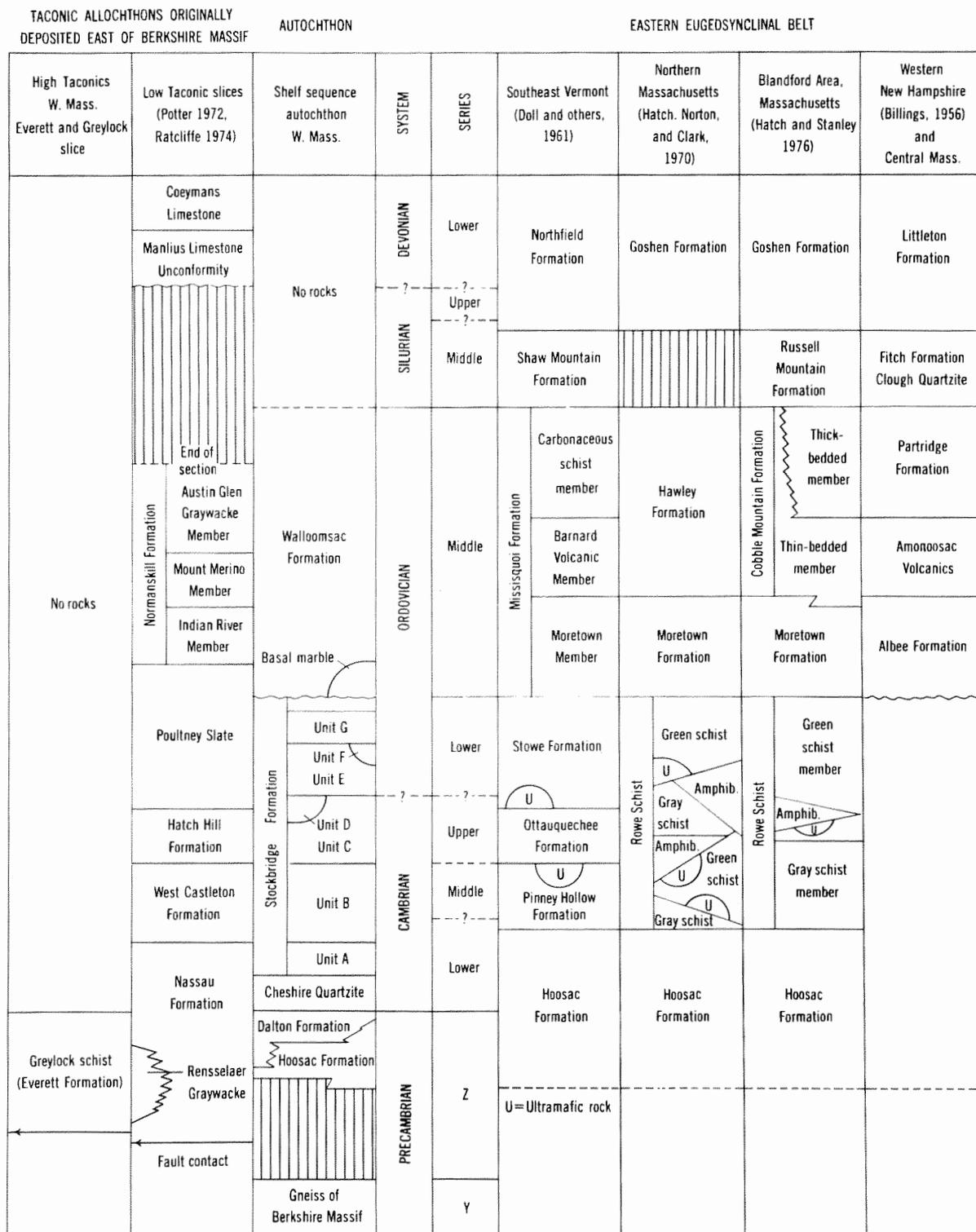


Figure 1.2. Stratigraphic sections and proposed correlations among sections in the Taconic allochthons, autochthon, and sections east of the Berkshire massif.

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East of the Berkshire massif is a succession of graywackes, shales, and volcanic rocks which have been metamorphosed to produce garnet to sillimanite-grade schists, gneisses, and amphibolites (fig. 1.2). Although many individual mapped units are lenticular and thus discontinuous along strike, and a few major facies changes have been mapped, the major lithologic units and their sequence are remarkably continuous in this belt from the Quebec border south through Vermont, Massachusetts, and Connecticut. Stanley and Hatch (1988) have interpreted these rocks to be a tectonic *mélange* and accretionary wedge produced during Taconic plate convergence. Within these rocks are abundant slivers or olistoliths of ultramafic rocks. The rocks of the Rowe Schist and Moretown Formation, east of and above the Whitcomb Summit thrust, therefore are interpreted to mark the locus of the Taconic suture (Stanley and Ratcliffe, 1985).

Allochthonous rocks present in the Taconic slices are thought to have rooted from an area situated between the Hoosac Formation of the Berkshire massif and Rowe Schist. In the area of this trip the allochthonous rocks belong to the Nassau Formation and Rensselaer Graywacke (and associated rift basalts) in the Chatham and Rensselaer Plateau slices, and the Everett Formation and Greylock Schist of the high-Taconic slices. The latter were emplaced by synmetamorphic thrust faulting. Cross sections (fig. 1.3) show the projected to cations of stop; these simplified sections are along the section lines of the State map (Zen et. al., 1983).

**ROAD LOG: DAY 1**

**Start** Black Swan Inn South end Laurel Lake, Rt. 20 Lee. Stockbridge Quadrangle (Ratcliffe, 1974b). **0.0**  
Turn right (south) on Rt. 20 and follow to High Street turn left **0.3** and go one block to Stop 1. **0.5**

**STOP 1. THE PINNACLE-ST. MARYS SCHOOL, LEE DALTON FORMATION-THRUST FAULT FABRIC IN PARAUTOCHTHON.** Large cliffs east of the school are of tectonically laminated feldspathic quartzite and schist of the Dalton Formation which are thrust over the Stockbridge Formation. These rocks are part of an arcuate thrust sheet of the cover rocks of the Berkshire massif that lie west of the train thrust slices of Middle Proterozoic rocks in the nested thrust sheet which comprise the Berkshire massif. The isoclinal reclined folds and intensely lineated rocks exhibit the style of the fold-thrust fabric of the Taconic D<sub>3</sub> event.

Turn left on High Street and in 0.2 turn left on Rt. 20, follow route 20 1.1 miles to intersection with 102 turn right. 0.2 mile turn left on 2.00 Tyringham Rd. turn left. 6.3 Center of Tyringham, turn right on Jerusalem Rd. 6.6 Park at gas powerline.

**STOP 2. COBBLE HILL-SOLE OF BEARTOWN MOUNTAIN NAPPE MIDDLE PROTEROZOIC TYRINGHAM GNEISS ON MIDDLE ORDOVICIAN WALLOOMSAC FORMATION.** Middle Proterozoic gneiss of the Beartown Mountain slice of the Berkshire massif overlies rocks of the Stockbridge and Walloomsac Formations in the east and north facing cliffs of the Cobble. These exposures are of the trailing edge of the Beartown Mountain slice, 11 km back, of the leading edge near Stop 1. Distinctive blastomylonitic gneiss reclined folds and well developed subhorizontal fold-thrust fabric of the Taconic D<sub>3</sub> event are parallel to the exposed thrust contact. This is the master sole thrust that carried the Berkshire massif and its nested internal thrust slices westward over the autochthon in the waning stages of the Taconic orogeny. <sup>39</sup>Ar/<sup>40</sup>Ar hornblende and biotite ages from blastomylonite of the thrust sheet, at staurolite-kyanite grade yield plateau ages of about 440 to 460 m.y. from exposures to the north near Lee and internal to the massif. Sutter et. al. (1985) interpret these ages to mean that the thrusting of the massif was synmetamorphic and occurred in the Taconic orogeny. After hike return to base of hill in Tyringham where bus will meet us.

**6.9** Turn left on Tyringham Rd. and retract route to Rt. 102. **11.5** Turn right on 102 go 0.2 mi and enter Mass Turnpike headed west (Albany). Follow turnpike to Rt. 22 at N.Y. State Line. **24.0** Exit to Rt. 22 to right. Follow 22 north to Tunnel Rd. **25.2** Turn left on Tunnel Hill Rd. continue west to west opening of tunnel and descend to RR tracks. **26.3**

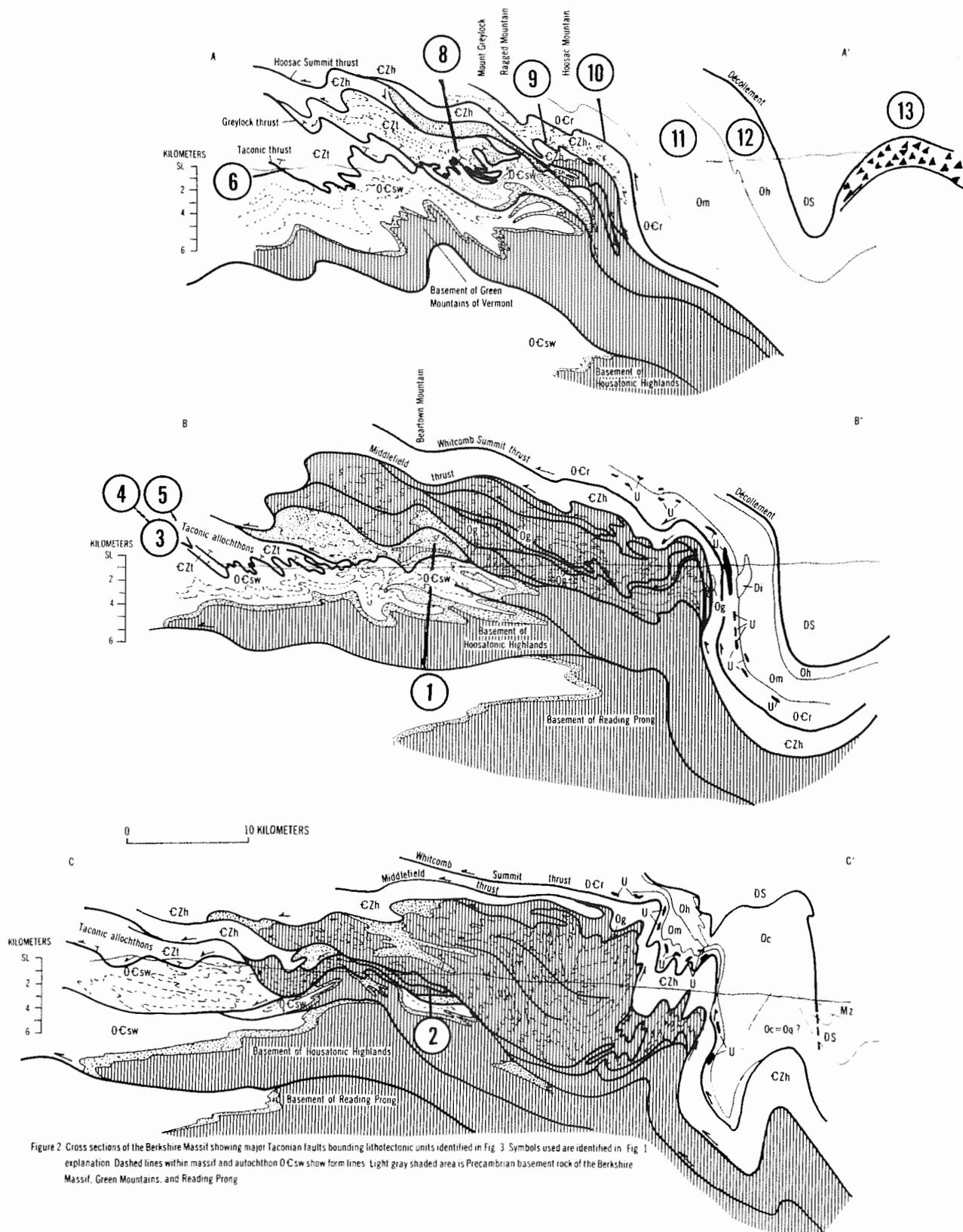


Figure 2. Cross sections of the Berkshire Massif showing major Taconian faults bounding lithotectonic units identified in Fig. 3. Symbols used are identified in Fig. 1 explanation. Dashed lines within massif and autochthon OCsw show form lines. Light gray shaded area is Precambrian basement rock of the Berkshire Massif, Green Mountains, and Reading Prong.

Figure 1.3. Cross sections along lines A-A', B-B', and C-C' showing stop locations.

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**STOP 3. CARBONATE ROCKS OF THE STOCKBRIDGE FORMATION AND THE SOLE OF THE CHATHAM SLICE OF THE TACONIC ALLOCHTHON.** From stop 2 to 3 we have crossed westward from the staurolite-kyanite grade rocks to chlorite grade rocks. In the south facing rock cut just east of the railroad tunnel the folded and cross foliated contact of green phyllite of the Nassau Formation (Late Proterozoic and Cambrian) is resting on black phyllite of the Walloomsac and limestone of the uppermost part of the Stockbridge Formation (Middle Ordovician). Examination of the geologic maps of this area show that the allochthonous rocks locally rest directly on the carbonate rocks of the autochthon; but that they are complexly folded together. Nonetheless, the regional foliation crosses the contact at a high angle, illustrating a regionally important point, namely that the first recognized regional metamorphism and foliation postdates emplacement of the Chatham slice. This relationship is universally true for the lower slices of the allochthon.

Continue west on Tunnel Hill Rd. to 0.6 mi to Columbia Co. Rt. 5, turn left follow five over thruway and up hill to exposures of purple and green phyllite of the Chatham slice. **28.8**

**STOP 4. NASSAU FORMATION AND REGIONAL SLATY CLEAVAGE. AGES OF CLEAVAGES. FOLIATION IN NASSAU FORMATION OF THE CHATHAM SLICE.** Road cuts of purple and green phyllites contain muscovite oriented in the regional cleavage that has given a K-Ar age of  $443 \pm 16$  m.y. (table 1 of Sutter et. al., 1985). This age agrees well with the recent  $^{39}\text{Ar}/^{40}\text{Ar}$  hornblende plateau age obtained by Hames of  $443 \pm 3$  Ma from the Walloomsac Formation at staurolite grade about 30 km to the south (Hames et. al., 1991). Hames' demonstrated that in broad areas of western Massachusetts, at grades lower than Acadian fibrolite, Taconic mineral fabric schistosity and chemical compositions are preserved. This is an important observation because it illustrates that even in severely overprinted polymetamorphic areas, disequilibrium assemblages and schistosity of older events very stubbornly persist into zones of remetamorphism and this in turn means that old structural fabrics are likewise preserved.

Continue south on Rt. 5 3.3 miles to intersection with Rt. 24, continue go to intersection with Rt. 22. **32.5** turn right on Rt. 22. **32.8** turn right on Rt. 203. **35.8** turn right on Weed Rd. and proceed 1 mile to drive on right. **36.8** turn right follow private drive to end of road and park.

**STOP 5. BOULDER CONGLOMERATE OF THE RENSSELAER GRAYWACKE IN THE CHATHAM SLICE AND DISCUSSION OF RIFT CLASTICS OF THE TACONIC ALLOCHTHON.** Some of the coarsest conglomerates in the allochthon occur in the Chatham slice. These excellent cliff exposures contain boulders of various kinds of granitic and retrograde paragneiss resembling Middle Proterozoic basement rocks exposed in the Adirondack Mountains. Autoclastic debris consisting of graywacke, quartzite and locally of basalt can also be found in the conglomerates. In addition basaltic pillow lava, volcanoclastic layers and diabasic basaltic flows are relatively abundant in the Rensselaer Plateau slice and Chatham slice. These rocks are interpreted as a rift clastics and basaltic submarine deposits formed during early continental extension. The alkalic chemistry of the basalts is consistent with low degrees of partial melting and are comparable to other rift basalts of the northern Appalachians such as the Tibbit Hill Volcanic member of the Oak Hill Group in Vermont (Ratcliffe, 1987; Coish et. al., 1985) and metabasalts in the basal Hoosac Formation.

Leave and retrace steps west on 203 to 22 and then Rt. 22 7 miles north to the thruway at State Line. Follow Rt. 22 north 7.7 miles to intersection with Rt. 20. **53.7** Turn right on 22 and 20 proceed 0.6 to branch follow Rt. 22 to the left to Lenanon springs follow Rt. 22, 6.6 miles north to intersection with Rt. 43. **60.3** turn right on 43. **61.5** turn left on East Rd. 0.2 mile park at powerline crossing.

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**STOP 6. METABASALT AND RENSSELAER GRAYWACKE AT SOLE OF RENSSELAER PLATEAU SLICE.** These outcrops are typical of the highly deformed metabasalts and metavolcaniclastic rocks of the eastern part of the Rensselaer Plateau slice described in Ratcliffe (1987, loc. 13). These thin epiclastic(?) laminated greenstones are interpreted as volcanoclastic deposits, although some epidote rich pods may be transported and broken pillows.

61.7 Return to 43 turn left east. 64.0 turn right on Brodie Mountain Rd. (to Lanesboro). 64.6 Crops opposite entrance to Jiminy Peak Ski area.

**STOP 7. SMALL KLIPPE OF NASSAU AND EPIDOTE GREENSTONES OF THE CHATHAM SLICE IN CONTACT WITH STOCKBRIDGE FORMATION.**

Continue east on Brodie Mountain Rd. 66.4 just over crest of hill crop of Walloomsac Formation. Continue down hill to intersection with Rt. 7. 67.4 Intersection Rt. 7, turn right past crop of Walloomsac Formation and Stockbridge Formation. 70.4 Turn left on road to Greylock Reservation, follow around sharp right bend. 72.2 Stop at Visitors Center and pile into school bus for trip to the top. The tour bus will meet us north of the mountain on Rt. 2 in the town of Greylock. 80.6 Top of Greylock

**STOP 8. GREYLOCK SLICE OF THE TACONIC ALLOCHTHON AND RELATIONSHIP OF THE TACONICS TO THE EUGEOCLINAL SEQUENCE ON HOOSAC MOUNTAIN.** The Greylock slice consists of green albitic and chloritoid-bearing phyllite and minor quartzite and beds of conglomerate. Overall, these rocks resemble closely certain rocks of the Hoosac Formation and Rowe Schist of the eastern eugeoclinal sequence. This was noted by Prindle and Knopf (1932), who concluded that the schists on Mount Greylock were thrust over the Stockbridge carbonates in a complex recumbently folded fault slice. They proposed that the sole of the fault was recumbently folded in transport as a result of a prolonged movement history. This paper was one of the earliest papers to document alpine-style deformation in the eastern U.S. After extensive remapping the present coauthors (Ratcliffe et. al., 1993) agree closely with the interpretations of Prindle and Knopf as regards the recumbent folds of the allochthon and the structure generally. Recumbent folds within the Greylock sequence predate the detachment, and the sole is also recumbently folded; a metamorphic fabric is developed at the thrust contact. Thus, emplacement of the Greylock slice was a synmetamorphic event, in contrast to most of the other Taconic allochthons which predate the regional foliation of probable Taconic age. Weather permitting, excellent views of the major regional structures can be seen. To the south is the sinuous front of the Berkshire massif and the lobate bulge of Beartown Mountain. Local peaks in the carbonate lowland are the Pinnacle, Stop 1, and Monument-East Mountain. Large ridges to the south and southwest are Canaan Mountain and Mount Everett of the Taconic allochthon. To the west, the main ridge of the Taconic Range in New York can be seen with the higher slope underlain by the Rensselaer Plateau slice. To the north, the south-plunging core of the Green Mountains can be seen. Hoosac Mountain with its core of Precambrian (Proterozoic) gneiss and unconformable sedimentary cover constitutes the long ridge to the east.

89.7 Exit Mount Greylock at Rt. 2, rejoin bus at Friendly's Parking lot north side of Rt. 2 a short distance west on Rt. 2. 90.2 Friendly's Restaurant, head east on Rt. 2 through North Adams on Rt. 2. 95.5 hair pin turn headed up Hoosac Mountain. 96.3 Stop at crest

**STOP 9. HOOSAC SUMMIT THRUST AND DISCUSSION OF THE HOOSAC FORMATION AND THE ROOT ZONE OF THE TACONIC ALLOCHTHONS.** Green chlorite-chloritoid-muscovite  $\pm$  quartz  $\pm$  garnet schists, albitic granofels and dark-rusty albitic quartz schist characterize the Hoosac Formation here at its type locality. To the north in Vermont and to a limited extent to the south the Hoosac contain layers of metabasalt. The metabasalts low in the Hoosac are identical to the alkalic types found in the Taconic allochthons, but basalts near the top of the Hoosac become increasing MORB-like and resemble closely those found in the overlying Rowe Schist, which are interpreted (Coish et. al., 1985) as oceanic basalt. It is thought

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the root zone for the Taconic allochthons was east of the Hoosac, which rest unconformably on the highest slice of the Berkshire massif, but west of the Rowe Schist. View to the north is of the North Adams Gap, the south plunging end of the Green Mountain massif, and the domal uplift of the Chester and Athens dome.

**99.0**

**STOP 10. WHITCOMB SUMMIT THRUST-ROWE SCHIST.** This contact between albitic schists, to the west, belonging to the Hoosac Formation and highly foliated, strongly rodded schists of the Rowe Schist is a regionally extensive and major tectonic boundary in western New England. East of this contact abundant slivers, pods and large masses of ultramafic rocks occur intermixed with a great variety of highly disarticulated metasedimentary and metabasaltic rocks. Hatch and Stanley (1988) interpreted the Rowe Schist as a *mélange* zone in the general position of the Taconic suture zone. The intense rodding and down dip lineation is produced by the hingelines of abundant isoclinal reclined folds. Throughout western Massachusetts and southern and central Vermont this deformation style characterizes the rocks near the Whitcomb Summit thrust. It is similar in spatial distribution to and orientation that of the major dated D<sub>3</sub> mylonite zones of the Berkshire massif seen earlier at Stops 1 and 2. Unfortunately we are now well above Acadian garnet grade and <sup>39</sup>Ar/<sup>40</sup>Ar ages of biotites and hornblende yield Acadian cooling ages (Sutter et. al., 1985).

Continue east on Rt. 2. **99.8** Whitcomb Hill Rd. continue east on 2 crops of Rowe at intersection. **102.2** Mobil Station at Drury, ultramafic serpentinites in Rowe near here. **104** Bridge over Cold River and exposures of the Moretown Formation. **108.9** Rt. 2 crosses Deerfield River immediately turn left on River Rd. Crops of Hawley Formation then schist and amphibolite of the Moretown proceed. **100** Large outcrops to right of Moretown Formation

**STOP 11. BIOTITE QUARTZ PLAGIOCLASE GRANOFELS (PINSTRIPE) OF THE MORETOWN FORMATION.** Large exposures of typical feldspathic granofels of the Moretown, here contain distinctive biotite-diorite dikes or sills. The Moretown is a regionally important unit recognized throughout Massachusetts and Vermont. Stanley and Ratcliffe (1985) and Stanley and Hatch (1988) interpret it a fore arc deposit developed between the Bronson Hill arc and the growing Taconic accretionary wedge. Therefore detritus was supplied from recycled Laurentia-derived sediments of the wedge and potentially also from arc volcanics in the Bronson Hill arc. The age of the Moretown is uncertain but is thought to be Ordovician. A recent total evaporation <sup>207</sup>Pb/<sup>206</sup>Pb zircon age of 484 ± 7 Ma comes from a trondhjemitic gneiss that intrudes the Moretown at Hallockville Pond (Williamson and Karabinos, 1993) in support of this age. Many of the large ultramafic bodies in western New England occur within or near the contact of the Moretown with rocks beneath. In Vermont the base of the Moretown is recognized as a major thrust fault (Ratcliffe et. al., 1992).

Return on River Rd. to route 2. **101.2** Turn left on Rt. 2 to Charlemont. **108.9** Turn right on 8A, cross RR tracks and turn right still on Rt. 8A south. Follow winding 8A 3 miles south to sharp right bend in 8A. **111.9** Turn left at bend onto Pudding Hill Road go 0.1, immediate turn left onto Middle Rd. and park by small bridge--crops are down the brook.

**STOP 12. PILLOW BASALTS AND VOLCANICLASTIC MAFIC ROCKS IN THE HAWLEY FORMATION.** The Hawley Formation is named for a collection of black sulfidic schists, interlayered metabasalts, and felsic gneisses interpreted by Stanley and Hatch (1988) as Ordovician arc volcanics and metasediments and part of the edifice of the Taconic arc. The age of the Hawley and its significance are currently being reexamined in Massachusetts by John Kim and Robert Jacobi (1993) and by Thomas Armstrong in Vermont (Armstrong, 1993), who find the bulk of the mafic and felsic rocks to be intrusive, diabase and trondhjemite rather than entirely volcanic igneous rocks. The age of the Hawley is not well constrained. Similar rocks in Vermont belonging to the Barnard Formation of Doll et. al. (1961) may range from

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Cambrian to Silurian (Aleinikoff and Karabinos, 1990) based on U-Pb ages of zircon. It may well be that this belt of rocks contains Taconic arc-related intrusives and volcanics as well as post-Taconic pre-Acadian intrusives. The pillows seen here constitute the best evidence known for volcanic rocks in the Hawley.

Return to **8A**, turn right (north) and return to Rt. 2 at Charlemont. **119.0** Turn right (east) and proceed on Rt. 2 toward Shelburn Falls. **121** Rt. 2 at Shelburn Falls. Park and descend into the bed of the Deerfield River.

**STOP 12. CORE GNEISSES OF THE SHELBURN FALLS DOME--LEADING EDGE OF THE BRONSON HILL VOLCANIC ARC.** A heterogeneous assemblage of mafic, felsic and intermediate gneiss assigned to the Collinsville Formation (Zen et al., 1983) appear in core of the Shelburn Falls dome, rimmed by Silurian and Devonian rocks of the Goshen Formation. These gneisses are generally regarded (Stanley and Hatch, 1988) as broadly correlative with the Ammonoosuc Volcanics and possibly other gneisses of the Bronson Hill arc, which here form the leading edge of the allochthonous arc that collided with the Laurentian margin and Taconic accretionary prism (Stanley and Ratcliffe, 1988). At present the age of the complex assemblage of the volcanic and intrusive rocks in the dome is poorly understood; however a recent discordant U-Pb zircon age suggests formation of the trondhjemitic gneiss of the dome might be as old as 488 Ma (Karabinos and Tucker, 1992). This age is compatible with some  $^{39}\text{Ar}/^{40}\text{Ar}$  data from north-central Vermont for early high pressure metamorphism at 470 to 500 Ma (Laird et al., 1974), which suggest subduction zone metamorphism and therefore potential arc formation by that time. On the other hand recent work by Tucker and Robinson (1990) has shown that the Ammonoosuc Volcanics and intrusive gneisses of the Bronson Hill are Late Ordovician and probably late arc-collisional (see discussion Day 2). The age and duration of Taconic volcanic arc or arcs(?) is really unknown at present; much complex geology undoubtedly remains to be discovered in these rocks.

## DAY 2: THE BRONSON HILL ANTICLINORIUM AND WESTERN PART OF THE MERRIMACK SYNCLINORIUM, CENTRAL MASSACHUSETTS

by Peter Robinson

### INTRODUCTION

Before 1965, and the application of plate tectonics, the Bronson Hill anticlinorium was known for fossil-based stratigraphy of medium- to high-grade metamorphic rocks and for gneiss domes. As early as 1890 B. K. Emerson (1898) carried out mapping with a keen sense of stratigraphy, but it was Marland P. Billings (1937), beginning in the Littleton-Moosilauke area in New Hampshire in 1933, that began the modern wave of detailed stratigraphic mapping. Billings and co-workers also developed the concept of gneiss domes. Initially the domes were considered to be Devonian laccolithic intrusions, but Jarvis Hadley (1942) showed that they contain a tectonic fabric post-dating their igneous history. In 1969 Naylor proved that the granitoid rocks in the domes are Ordovician, but the exact nature of their contact with the stratified Ordovician cover still remains controversial. An intrusive contact (Leo, 1991), unconformity (Robinson, 1981), detachment fault (Tucker and Robinson, 1990) and thrust fault (M. J. Kohn and F. S. Spear, personal comm., 1991) have all been proposed. More recent important discoveries include the recognition of major west-directed Acadian fold nappes (Thompson et al., 1968); the discovery of new fossil localities, some in high-grade metamorphic rocks (Boucot et al., 1958; Boucot and Thompson, 1963; Elbert et al., 1988); the recognition of metamorphic overhangs in the Connecticut Valley region (Thompson et al., 1968; Spear and Chamberlain, 1986) and of Devonian metamorphism as high as the pyroxene-granulite facies (Hess, 1969, 1971; Tracy et al., 1976; Hollocher, 1985; Robinson et al., 1989); the demonstration that the early fold nappes are followed by major synmetamorphic thrust nappes (P. J. Thompson, 1985; Robinson et al., 1991; Berry, 1989); and the discovery that the last ductile deformation and metamorphism in part of the anticlinorium was Pennsylvanian (Tucker et al., 1988; Gromet and Robinson, 1990; Robinson et al., 1992).

Within the context of this field trip, emphasizing plate-tectonics, and within the imposed space and time restrictions, a complete summary and field examination of regional stratigraphy and tectonics is impossible. I can only provide a broad outline, within which the significance of individual stops can be evaluated, and in which the conditions of field, petrologic, and geochronologic studies can be appreciated.

### STRATIGRAPHY AND INTRUSIVE ROCKS

For this trip the rock layers and igneous intrusions has been subdivided into seven units, each with its own problems of protoliths, plate-tectonic setting, and tectonic and metamorphic history that will be included here,

although outside the narrow realm of stratigraphy. Next to each heading the stops are listed where the unit will be seen. Naturally the deepest, oldest rocks have the most complex history and are the most difficult to put in context.

**Late Proterozoic Basement (Stops 2, 7).** The Dry Hill microcline Gneiss with its Pelham Quartzite Mbr. and the Poplar Mountain quartz-biotite Gneiss with its basal Poplar Mountain Quartzite Mbr. in the core of the Pelham gneiss dome (Ashenden, 1973; Robinson et al., 1992) are the deepest, oldest rocks exposed in the Bronson Hill anticlinorium. The stratigraphic relations suggest that the Dry Hill Gneiss forms the core of an early, probably middle Acadian, E-directed recumbent fold nappe that is both overlain and underlain successively by Poplar Mountain Quartzite and Poplar Mountain Gneiss. A suspected late Proterozoic igneous age for the Dry Hill was shown by Naylor et al. (1973) and confirmed by Tucker and Robinson (1990) based on euhedral zircons with a U-Pb age of  $613 \pm 3$  Ma. The Dry Hill Gneiss is considered to be a series of alkali rhyolite lavas or pyroclastics (Hodgkins, 1985) with minor interbedded quartzites and the major Pelham Quartzite Member. The Poplar Mountain Gneiss is a geochemically related sequence of sedimentary derivation, with the Poplar Mountain Quartzite Member and with local felsic interbeds identical to Dry Hill Gneiss. An intrusive origin for the Dry Hill cannot be ruled out; associated quartzites yield detrital zircons only in the age range 2800-930 Ma and none of Dry Hill age. This Proterozoic basement most closely compares with Potter Hill Gneiss and associated quartzites from the Hope Valley terrane of southeastern Connecticut, considered by most to be part of Avalon. Furthermore, Robert Ayuso (pers. comm., 1993) believes that the radiogenic Pb isotopic signature of the Dry Hill overlaps the Pb isotopic range in Proterozoic rocks in Avalon. Detrital single zircons from the Pelham Quartzite Member are in three age groups that correspond to inheritance ages in one sample of Dry Hill Gneiss. These age groups are consistent with either South American or Laurentian sources.

The Dry Hill Gneiss has yielded U-Pb ages on sphene (Tucker et al., 1988) and Sm-Nd and Rb-Sr mineral isochrons (Gromet and Robinson, 1990) showing that the last significant ductile deformation and metamorphism was late Pennsylvanian (295-290 Ma). Gromet has found no convincing isotopic evidence that the Dry Hill Gneiss was ever subjected to Acadian metamorphism, and he has postulated that the Dry Hill Gneiss was a part of Avalon that was underthrust into its present position during the Late Pennsylvanian. In a single-zircon study of the Poplar Mountain Gneiss, Tucker (Robinson et al., 1992) has found a variety of Proterozoic discordant grains and three concordant grains at 585, 370, and 302 Ma. These might be interpreted, respectively, as Proterozoic igneous zircon from the Dry Hill Gneiss or a related rock, Acadian zircon from a dismembered pegmatite, and late Pennsylvanian metamorphic zircon. These results are consistent with structural arguments that the Dry Hill Gneiss and related rocks were in roughly their present position in time to be involved in middle Acadian E-directed recumbent folding.

If the Dry Hill Gneiss and related rocks do represent the basement of the Ord. Bronson Hill magmatic arc, then their provenance is crucial to plate-tectonic interpretations. The recent proposal of Dalla Salda et al. (1992) that the Taconian orogeny was a continental collision between Laurentia and western South America could be consistent with the detrital zircons from the Pelham Quartzite as well as paleomagnetic reconstructions (Trond Torsvik, pers. comm., 1993) showing western South America at roughly equatorial latitudes in the Ordovician when the Avalon part of Gondwana was still at high southern latitudes. The Pb isotopic data of Ayuso would then have to be explained in terms of some broader effect within Gondwana. If the Gromet interpretation of Penn. underthrusting of Avalon crust is correct, then the basement of the Bronson Hill magmatic arc will have to be sought elsewhere.

**Paleozoic Undercover (Stop 6).** In the southern part of the Pelham dome, near the upper contact of the Late Proterozoic rocks described above, there is a sequence of sillimanite- and kyanite-bearing mica schists, amphibolites, quartzites, and gneisses. These were mapped together as the Mount Mineral Formation and were interpreted as a more aluminous facies of the upper layer of Poplar Mountain Gneiss in the northern part of the dome. These proved to have relics of a near granulite-facies metamorphism overprinted with kyanite-staurolite-muscovite grade assemblages. These were interpreted (Roll, 1987) as relics of a Proterozoic metamorphism overprinted by the "ambient" Acadian facies. The U-Pb isotopic research of Tucker (Robinson et al., 1992) has caused a drastic reinterpretation. Monazite U-Pb ages in the rocks with granulite-facies relics are  $367 \pm 2$ , whereas monazites in schists and quartzites with kyanite-muscovite overprints are  $298 \pm 2$  Ma. Thus, within a hundred meters of the Dry Hill and related rocks we have evidence of an Acadian metamorphism of the same age and temperature, and of slightly higher pressure than the central Massachusetts granulite-facies high! Studies of detrital zircon within the upper muscovite-bearing quartzite of the Mount Mineral Fm. have yielded three grains with concordant ages of 459, 441 and 439; thus the quartzite can be no older than the early Sil. and is probably the Clough Quartzite (see below). Correlation of other parts of the thin Mount Mineral Fm. with various Ord., Sil. and Lower Dev. units is now in progress, and its outcrop must now be considered as part of a tectonic window beneath Ordovician intrusive basement, bounded upward either by a thrust or, less probably, by a fold nappe of the Ord. intrusive rocks.

**Late Ordovician Intrusive Basement (Stops 3, 5, 8, 13).** The predominant rocks exposed in the cores of the domes are quartz-feldspar gneisses and amphibolites, interpreted as a highly deformed and metamorphosed intrusive igneous complex. In some parts the complex is homogeneous, representing large intrusive masses of calc-alkaline granite or tonalite. Elsewhere it is strongly layered with alternating tonalite and monotonous hornblende-biotite amphibolite (see Stop 8), but lacking amphibolites with Fe-Mg amphiboles and felsic rocks with garnet, muscovite and sillimanite characteristic of the overlying Ammonoosuc Volcanics. The complex resembles rocks in other orogens inferred to have come from the roots of an island arc. Extensive U-Pb zircon dating based on large well studied exposures (Tucker and Robinson, 1990) has yielded a total age range of 455-443 Ma. The upper age limit corresponds almost exactly to the recognized time of emplacement of the Giddings Brook slice of the Taconic allochthons. Thus, the magmatic rocks exposed here cannot represent the magmatic arc as it existed prior to the Taconian collision, but only rocks produced during the collision and subsequent metamorphism of the suture zone, for which Hames (Hames et al., 1991) has determined a hornblende Ar-Ar cooling age of about 440 Ma.

**Late Ordovician Stratified Cover (Stops 4, 5).** The Ordovician intrusive complex is separated from the Ordovician stratified cover by a sharp planar contact. The strata consist of the Ammonoosuc Volcanics and the Partridge Formation. The belt can be traced discontinuously into Maine where equivalents of the Partridge contain Caradocian graptolites (Harwood and Berry, 1967).

The Ammonoosuc Volcanics consists of three mappable members (Robinson, 1963; Schumacher, 1988) and ranges in present mappable thickness from 30 to 1300 meters. The Mafic Lower Mbr. consists predominantly of metamorphosed basaltic and andesitic lavas and tuffs of tholeiitic arc affinities, or their hydrothermally altered equivalents, with relatively minor felsic interbeds. Locally pillows, graded tuffs, agglomerates, and other features are preserved even at sillimanite-grade. Hornblende is typical of many of these rocks, but great chemical variety leads to metamorphic assemblages with garnet, epidote, diopside, and particularly anthophyllite, gedrite and cummingtonite. The hydrothermally altered basalts are metamorphosed to cordierite-gedrite gneisses. Locally the top of the member is marked by marble, marble-matrix volcanic conglomerate, or diopside-rich amphibolite. The Middle Garnet-Amphibole Quartzite Mbr. (Stop 6), commonly with cummingtonite and/or gedrite, hornblende, and magnetite is a widespread very thin unit considered to be a deposit from volcanic exhalations; it marks an abrupt change in volcanism from predominantly mafic to predominantly felsic. The Felsic Upper Mbr. consists of metamorphosed rhyolites and dacites or their altered equivalents. These are characteristically peraluminous rocks with abundant metamorphic garnet and muscovite, believed on the basis of major and trace elements to have melted from a basaltic source. Hydrothermally altered equivalents are now various muscovite, kyanite, and sillimanite-rich rocks, including the sillimanite nodular gneiss at Stop 6. A metamorphosed qtz.-phyric rhyolite from the Felsic Upper Mbr., about 30 m below the base of the Partridge Fm. (see road log after Stop 1), has yielded a U-Pb zircon age of  $453 \pm 2$  Ma.

The overlying Partridge Fm. is dominated by metamorphosed sulfidic black shales, now garnet-mica-kyanite or sillimanite-ilmenite-graphite-pyrrhotite schists. The lower part contains abundant volcanic interbeds like those of Lower and Upper Members of the Ammonoosuc. That mafic volcanism characteristic of the Lower Ammonoosuc continues into the Partridge suggests the Felsic Upper Ammonoosuc may not represent a termination of mafic magmatism so much as a prodigious, possibly very brief explosive punctuation by felsic magmatism. A 1 m-thick pyroclastic rhyolite bed about 10 m above the base of the Partridge has yielded a U-Pb zircon age of  $449 \pm 3-2$  Ma.

The basal contact of the Ammonoosuc is a major regional problem. It has been described as an intrusive contact (Billings, 1937; Leo 1991), but none of the well mapped contacts within the Ammonoosuc or between the Ammonoosuc and Partridge are ever seen truncated by the contact, nor do any of the mafic xenoliths in the gneisses correspond to any of the distinctive and unusual amphibolites of the Ammonoosuc. It has also been described as an unconformity (Robinson, 1981) on the basis of three occurrences of quartzite and quartz-pebble conglomerate at the base of the Ammonoosuc, and the petrographic and geochemical dissimilarity of the two sequences (Robinson et al., 1989). Present geochronology indicating the two sequences overlap in age, seems to preclude both these possibilities, and to suggest instead a fault, possibly a detachment fault in an arc setting, that juxtaposed a tholeiitic cover sequence on top of the roots of an adjacent calc-alkaline arc.

**Silurian-Devonian Stratified Cover (Stops 1, 3, 5, 10, 11, 12, 14, 15).** The Sil.-Dev. cover rests unconformably on the Partridge Formation, the Ammonoosuc Volcanics or the tonalitic gneisses and amphibolites of the Ordovician intrusive basement. The Sil. and earliest Dev. is divided into two facies, originally separated by a tectonic hinge near the present eastern edge of the Bronson Hill anticlinorium. A western sediment source is inferred from studies of less metamorphosed strata in Maine. The western sequence (Connecticut Valley belt) is a relatively thin, discontinuous section deposited across the ruins of the Taconian orogeny. The eastern

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sequence (Merrimack belt) is a much thicker, continuous sequence of coarse and more impure clastic sedimentary rocks deposited east of the tectonic hinge, either in an extensional trough or as part of the continental slope-rise deposits of post-Taconian North America. This latter interpretation could be consistent with the separation of South America following the Taconian collision (Dalla Salda et al., 1992).

The thin western Silurian consists of two units: Late Llandoverly Clough Quartzite and the Pridoli to Lochkovian Fitch Formation. The basal Clough is typically conglomerate with over 95% vein quartz pebbles and garnet, muscovite  $\pm$  sillimanite in the matrix. This is succeeded upward or in facies relation by well-bedded quartzite. The uppermost Clough is a calcareous metasandstone, now diopside-grossular calc-silicate, with rare fossil fragments that include crinoid columns and brachiopods (Boucot et al., 1958). The Fitch Formation is predominantly metamorphosed calcareous shale, now commonly zoisite-diopside-biotite calc-silicate, with interbeds of sulfidic shale, now mica schist, and rare interbeds of marble. At Littleton, NH, where corals and brachiopods had long been known, conodonts (Harris et al., 1983) indicates a Pridoli (uppermost Silurian) age. At Bernardston, MA, a crinoid-bearing marble (Stop 1) has yielded Lochkovian (lowermost Devonian) conodonts (Elbert et al., 1988).

The thick eastern Silurian is dated by lithic correlation to fossiliferous strata in central and eastern Maine (Robinson et al., 1991). It begins with the thick Lower Sil. Rangeley Formation (Stops 11, 12), mixed graphitic and sulfidic feldspathic turbidites with small lenses of calc-silicate rock and rare lenses of polymict conglomerate. This grades upward into the relatively thin Perry Mountain Fm., a more quartzose, better and thinner bedded turbidite, locally marked at its top by apatite-bearing garnet quartzite or garnet-magnetite iron formation (Robinson and Elbert, 1992). The Rangeley-Perry Mountain is overlain by a distinctive sulfidic zone, possibly related to a widespread anoxic event in the Wenlock. In the west this is a calcareous, now calc-silicate section, the Francestown Fm. In the east this is interbedded sulfidic pelite and quartzite, the Smalls Falls Fm. (Stop 10). Above the sulfidic zone are thick calcareous feldspathic turbidites and interbedded shales, now biotite granulites, calc-silicates and schists, of Ludlow to Lowest Dev. age (Warner Fm. in NH or Madrid Fm. in ME). In this package in Maine evidence has been found for the earliest reversal of source direction from North America to tectonic lands in the east (presumably the approaching margin of Avalon). In Maine, and presumably also in Massachusetts, the better-defined Merrimack-belt sequence appears to grade eastward into a section from lowest to highest Silurian dominated by calcareous turbidites like the Warner-Madrid. Such rocks, of uncertain Silurian age, have been assigned to the Paxton Fm. (Stops 14, 15) in east central Massachusetts.

The major Lower Devonian unit is the Littleton Fm. (Stop 3), metamorphosed carbonaceous non-sulfidic shale and quartzose sandstone dated by brachiopods as Emsian (Boucot and Arndt, 1960) in western New Hampshire. It is interpreted as a west-spreading deltaic complex or Acadian flysch with a source in early Acadian tectonic lands to the east (Hall et al., 1976). In the western part of the Bronson Hill anticlinorium Littleton is overlain with apparent unconformity by Erving Fm., a controversial unit of metamorphosed basaltic tuff and calcareous siltstone and shale. Robinson et al. (1988) have correlated Erving with a section in the Connecticut Valley synclinorium consisting of Goshen Fm., Waits River Fm., Standing Pond Volcanics, and Gile Mountain Fm., thus implying that these units are also Devonian. These correlations are strengthened by the major and trace element similarities (J. C. Hepburn, pers. comm., 1990) of Erving and Standing Pond Volcanics and by Emsian plant fossils recently described from Gile Mountain correlatives in southern Quebec (Hueber et al., 1990). However, Trzcieski et al. (1993) on the basis of interpretations in Vermont, have suggested that the Erving is Ordovician.

**Acadian Intrusions (Stop 9).** Acadian intrusions are abundant and beyond coverage of this guidebook. They range from early tectonic to post-tectonic, gabbro to granite. Several have been studied in detail. The largest, all early to late syntectonic, include the Hardwick Tonalite (Shearer, 1983; Shearer and Robinson, 1988), Nichewaug sill of augite diorite (Shearer, 1983), Fitchburg Plutonic Complex (Maczuga, 1981), Belchertown Quartz Monzodiorite (Ashwal et al., 1979), Coys Hill Granite which is the equivalent of the Kinsman Granite in New Hampshire (Clark and Lyons, 1986), Prescott Complex gabbro and tonalite (Makower, 1964), Mount Hermon Gabbro (Elbert, 1988), and the diorite and gabbro complex at West Warren (Pomeroy, 1977). The Belchertown intrusion, Nichewaug sill, Hardwick Tonalite, and Fitchburg complex have locally preserved contact metamorphic effects overprinted by regional metamorphism. Although small in area, the tonalite intrusions mapped by Berry (1989) in the Brimfield-Sturbridge area are important because they cut across nappe folds and thrusts, but are overprinted by peak Acadian granulite-facies metamorphism and cut by backfold-stage mylonites (Finkelstein, 1987; Berry, 1989). The Coys Hill Granite appears to be a partial melt from highly aluminous crust, thought not from the surrounding Silurian-Devonian pelites (Clark and Lyons, 1986). The Belchertown quartz-monzodiorite, dated at  $380 \pm 5$  Ma, has many of the peculiarities of sanukitoids, granitoid melts derived from small fraction high temperature melts from mantle (G. N. Hansen, pers. comm., 1992) and in addition has somehow acquired an extremely high

oxidation state (Ashwal et al., 1979). The remaining tonalites and granites fall near or on the boundary between I- and S-type granitoids and have characteristics of subduction-type magmatism. However, high K, P and Mn suggest that a connection to basaltic magmas in an extensional setting should be considered (Shearer and Robinson, 1988).

**Triassic-Jurassic Continental Sedimentary Rocks; Jurassic Basalts; and Jurassic, Cretaceous and Diabase Dikes.** The Connecticut Valley of Massachusetts and Connecticut is extensively underlain by Late Triassic and Early Jurassic continental sedimentary rocks and Jurassic basaltic lavas in the Northfield, Deerfield, and Hartford basins, best dated by fossil pollen (Cornet, 1977). They lie in a complex half graben structure with the major fault along the east, having a vertical displacement of about 5 km in northern Massachusetts, increasing to about 8 km at the Connecticut line. They conceal important Appalachian geology, but fault displacements give us three-dimensional information not otherwise available. The crystalline highlands to the east and west are cut by four sets of Jr. tholeiite dikes, two correlative with the 1st and 3rd lava extrusions in the valley (McEnroe and Brown, 1992), and two sets of Cret. tholeiite dikes that are unlike the alkali diabases and lamprophyres typical of Cretaceous intrusions elsewhere in New England (McEnroe et al., 1987; McEnroe, 1989).

## STRUCTURAL DEVELOPMENT

### Taconian Deformation

Notable in the Bronson Hill anticlinorium and the eastern flank of the Berkshire anticlinorium is the lack of evidence for major pre-Silurian deformation just beneath the Silurian-Devonian rocks. The base of the Clough Quartzite does cut down from the Partridge through the Ammonoosuc and locally into the underlying Ordovician intrusives, but the author has found only one place where there is evidence for a pre-Silurian fold, and all evidence for pre-Silurian faults (detachments, etc.) is circumstantial. This is surprising in view of the closeness of known major Taconian folds and thrusts. The best explanation is that Silurian-Devonian strata were deposited mainly on rocks of the Bronson Hill arc above the Taconian subduction and deformation zone.

### Acadian Deformation

Acadian deformation is divided into the nappe stage, the backfold stage, and the dome stage. Some features previously assigned to the dome stage have been proved to be part of a late Pennsylvanian stage of doming, ductile deformation and kyanite-grade regional metamorphism (Robinson et al., 1992).

**Nappe Stage.** The early nappe stage produced west-directed fold nappes with tens of kilometers of transport (Thompson, 1954; Thompson et al., 1968; Robinson et al., 1991) during regional metamorphism, soon after the end of Early Devonian sedimentation, and very soon after the intrusion of the Kinsman Granite, dated by Sm-Nd and U-Pb garnet at  $413 \pm 5$  Ma (Barreiro and Aleinikoff, 1985). The best characterized of these are the Bernardston nappe (Stop 1) and the overlying Skitchewaug nappe near Springfield, Vermont, first described by J. B. Thompson. Curiously many of the important fossil localities are localized close to the anticlinal hinges of these two nappes.

More subtle and more difficult to characterize are the later thrust nappes (P. J. Thompson, 1985; Robinson et al., 1991), also formed fairly early in the regional metamorphism, which cut through the pile of earlier fold nappes, leading to very confusing contact relations. Most important of these is the Brennan Hill thrust, which brings previously folded strata of the western part of the Merrimack belt westward over folded strata of the Connecticut Valley belt. A second and higher major thrust, the Chesham Pond thrust, just beneath the Cardigan pluton of Kinsman Granite and inferred to run along the west margin of the Coys Hill Granite (Stop 9), brings a more eastward facies of the Merrimack belt, including the Smalls Falls Fm., over a more westward facies that includes the Francetown Fm. Farther east Berry (1989, 1992) has recognized that early W-directed folds are truncated by a series of W-directed thrust slices, each involving pre-Silurian granitoid and stratified basement (Stop 13) and Silurian strata of the Merrimack belt, all formed before peak granulite facies metamorphism at 369-362 Ma (Thomson et al., 1992). Field trip stops 1, 2, 3, 4, 6, 7, and 8 are below the Brennan Hill thrust; stop 5 may be on it; overnight in Amherst October 22 and Stop 11 are above it; and Stops 9, 10, 12, 13, 14, and 15 are above the Chesham Pond thrust.

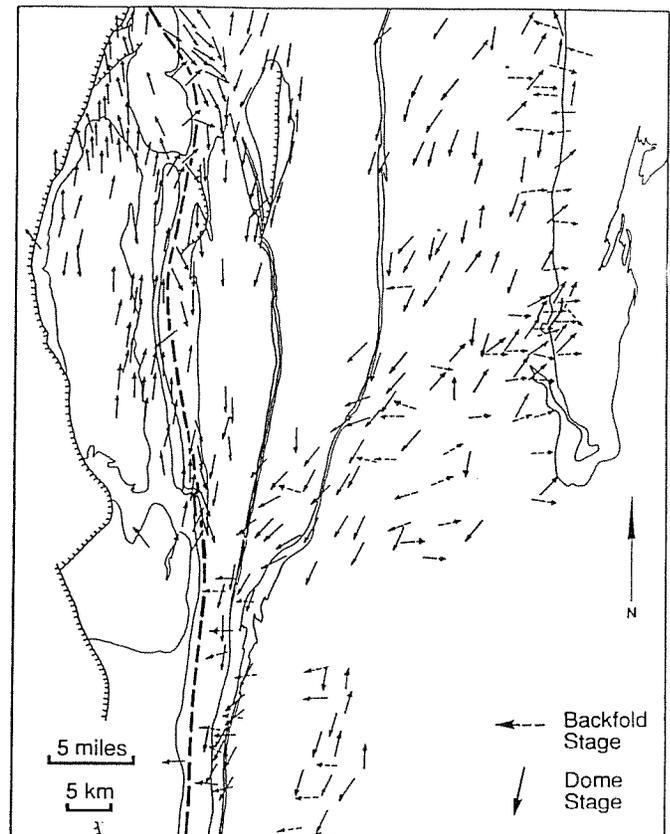
**Backfold Stage.** Very few minor structural features relate to the nappe stage in central Massachusetts. In the backfold stage, well preserved fabrics were formed that overprint the peak metamorphic fabric in the granulite facies region. This was when the east side of the Bronson Hill anticlinorium, with its W-directed fold and thrust nappes was overturned to the east, and also when the main and Tully bodies of Monson Gneiss were transported northward and overturned, so they now lie above surrounding younger strata. It was also probably the time of E-directed recumbent folding and thrusting of Proterozoic to Lower Dev. strata near the Pelham dome. The important

fabric is an E-W trending lineation, rarely accompanied by parallel folds and commonly by a consistent west-over-east shear sense along the lineation. This fabric is observed in stratified and intrusive rocks east of the Bronson Hill anticlinorium (Peterson and Robinson, 1993,) and is progressively overprinted by a N-S to NE-trending dome-stage fabric and folds approaching the Bronson Hill anticlinorium (Figure 2.1). U-Pb dates on monazite in the granulite facies region suggests setting at about 369-362 Ma, probably close to the onset of post-peak shearing. Stop 11 will show this fabric in the Conant Brook shear zone where it is progressively overprinted by the dome-stage shearing.

**Dome Stage.** The Acadian dome stage is characterized by N-S to NE-trending, mostly S-plunging folds and parallel lineations. The folds outline the major domes in the eastern part of the anticlinorium. Mineral grain shapes and elongate conglomerate pebbles indicate this is a stretching lineation quasi-parallel to fold axes. This was poorly understood until Peterson (1992; Peterson and Robinson, 1993) discovered a right-lateral shear fabric in sillimanite-orthoclase grade rocks parallel to this lineation near the east margin of the Monson Gneiss and inferred its origin by orogen-parallel elongation. A wider study awaits for shear fabrics parallel to lineation, using Peterson's techniques.

**Pennsylvanian Dome Stage.** Elongation lineations quasi-parallel to fold axes occupy an enormous region in central Massachusetts. In the northern part of the Pelham dome, where lineation plunges gently north, a series of late asymmetric folds including sheath and tubular folds formed during lineation-parallel shear, indicate north-over-south shear parallel to the axis of the dome (Ashenden, 1973; Onasch, 1973). This sense was confirmed in fabric analyses by Reed (1993, Reed and Williams, 1989) and it can be shown that upper levels were sliding south relative to deeper levels throughout the length of the Pelham dome. In going from the region of north-plunging lineations in the northern part of the Pelham dome eastward to the region of south-plunging lineations in the southern part of the Keene dome (Figure 2.1), one might expect the lineations to pass through the horizontal. In fact they are connected by a swirl within which lineations plunge due east down the dip of the foliation. This "great swirl" (Robinson, 1963), has long been a puzzle and was thought to be a product of strain reorientation during the Acadian. A solution appeared from the U-Pb dating of metamorphic sphene and monazite by Tucker (Robinson et al., 1992) showing that the axis of the swirl is essentially the locus along which Acadian metamorphic minerals were recrystallized in the late Pennsylvanian. Our new model for the swirl (Peterson, et al., 1993) is that it is the pattern of Acadian lineation as it passes progressively westward into a Late Pennsylvanian strain field.

Figure 2.1. Regional map showing distribution and representative orientations of backfold- and dome-stage lineations across central Massachusetts (from Peterson, 1992). Heavy dashed line shows the eastern limit of strong late Pennsylvanian ductile deformation and recrystallization.



## METAMORPHISM

At face value the metamorphic zones in west central Massachusetts are of a classic Barrovian pattern, from the chlorite zone of the Connecticut Valley metamorphic low to the granulite facies in the central Massachusetts high (Tracy et al., 1976; Robinson et al., 1989). The abundance of normal shale compositions as contrasted to the Moine-Dalradian of Scotland makes the characterization of metamorphic facies relatively easy. The abundance of mafic rocks, especially Fe-Mg amphibole rocks of the Ammonoosuc and Partridge, has allowed successful parallel studies (Robinson and Jaffe, 1969a,b; J.C. Schumacher and Robinson, 1987; J.C. Schumacher, 1988; Robinson et al., 1982a; Hollocher, 1985, 1991; Schumacher et al., 1989, 1990; R. Schumacher, 1986). Closer scrutiny, however, leads to many unsolved problems.

West of the Mesozoic Connecticut Valley fault near the north border of Massachusetts, metamorphism is progressive and apparently Acadian from chlorite zone through biotite, garnet (Stop 1), and staurolite zones until sillimanite-muscovite assemblages are attained just beneath the Ashuelot Pluton of Kinsman Granite. This is a region of inverted metamorphism in which structurally higher rocks in fold and thrust nappes are at higher metamorphic grade, and has been studied in greater detail in western New Hampshire (Spear and Chamberlain, 1986; Elbert, 1988, Spear et al., 1990). This metamorphism reaches sillimanite-muscovite-orthoclase grade in the Amherst block of Rangeley Formation exposed between the Deerfield and Hartford Mesozoic basins, where early Acadian U-Pb monazite ages are obtained at around 400 Ma (R. D. Tucker, personal communication, 1992), similar to monazite ages in the sillimanite-orthoclase Acadian high in central New Hampshire (Eusden and Barreiro, 1988).

Directly east of the Mesozoic fault pelites are much coarser grained and contain the assemblage muscovite-biotite-garnet-staurolite with or without kyanite. Garnets show prograde growth zoning (Tracy et al., 1976) and the rocks yield estimated temperatures of about 580°C and estimated minimum pressures of 6 kbar. Based on new geochronology we know this metamorphism was Pennsylvanian and overprinted a largely unknown Acadian pattern. We know that part of the Mount Mineral Formation was subjected to near granulite-facies Acadian metamorphism with assemblages indicating 700°C and 6.8 kbar, similar to but at higher pressure than the central Mass. high.

Beyond the eastern limit of Pennsylvanian overprinting, a rational pattern of zones appears in pelites beginning with sillimanite-muscovite-staurolite (II), then sillimanite-muscovite (III), sillimanite-muscovite-orthoclase (IV) (Tracy, 1978), sillimanite-orthoclase (V), and finally sillimanite-orthoclase-garnet-cordierite (VI). However, the pattern bears no rational relation to obvious structural features or intrusive rock distributions, although it does seem to follow the shape of a positive gravity anomaly (Kick, 1975; Fitzpatrick, 1978). Nearly everywhere east of the Bronson Hill anticlinorium pelites can be found containing sillimanite pseudomorphs after andalusite, indicative of early high-temperature/low-pressure before the onset of granulite-facies conditions. Pelites imply peak metamorphic conditions at 700-800°C and 6-6.5 kbar (Thomson, 1992; Thomson et al., 1992), and very dry as shown by the rarity of retrograde reactions on mineral rims except simple local ion exchange, even within such post-peak features as the Conant Brook shear zone (Peterson, 1992). There is evidence of cooling at the same or slightly increasing pressure in garnet-sillimanite-quartz symplectites inside cordierites (Thomson, 1992; Thomson et al., 1992).

In the sillimanite-muscovite zone, cordierite-gedrite gneisses show the development of aluminous enclaves indicative of pressure release during uplift of the Keene gneiss dome (Robinson and Jaffe, 1969a; Schumacher and Robinson, 1987). In higher grade mafic rocks Hollocher has charted the progressive breakdown of Fe-Mg amphiboles and then of hornblende to produce augite-orthopyroxene-plagioclase assemblages, and then the beginnings of the breakdown of biotite to produce orthopyroxene + orthoclase. New evaluations of conditions in these pyroxene granulites suggest peak conditions as high as 850°C (Schumacher, Hollocher and Frost, in prep.). Berry has described several occurrences of marble containing wollastonite in quartz-calcite-anorthite rocks and has shown that grossular-rich garnet only grows at calcite-anorthite or wollastonite-anorthite contacts when supplied with requisite FeO and MgO from adjacent diopside (Berry, 1989; Robinson, 1991). He has also described a gneiss (Thomson et al., 1992) with the assemblage biotite-plagioclase-orthoclase-garnet-corundum-sillimanite-spinel close to experimental calibrations and indicating a temperature of 800-825°C at 6-6.5 kbar.

It is speculated that the early low-pressure heating, indicated by the widespread occurrence of andalusite pseudomorphs, may have accomplished much of the necessary dehydration required to produce the later granulites. This could have occurred under extensional conditions that thinned the crust and brought mantle close to the surface (Berry and Robinson, 1989), or by delamination of continental lithosphere in front of an east-dipping subduction zone, allowing asthenosphere to come close to the base of the crust and provide both heat and mafic magmas that

gave the impetus for most of the Acadian magmatism. This vision is in stark contrast to the model of Chamberlain and Sonder (1990), who propose that an abundance of heat-producing elements in the Silurian-Devonian section under deep burial could have provided all of the thermal energy for Acadian metamorphism and magmatism. The very importance of mafic and intermediate magmas in the region belies this interpretation.

### ACKNOWLEDGEMENTS

The research results covered in this article involved a large number of persons over a period of more than 30 years, and most have been acknowledged previously. Several of the stop descriptions have been adapted from guidebook material published previously; this is cited for each stop. The research covered in this article was supported by the U. S. Geol. Survey during preparation of the Massachusetts Bedrock Map and by Natl. Science Foundation Grants to Robinson both before and after. The manuscript was prepared while the author was a guest researcher at the Depart. of Mineralogy and Petrology, Lund Univ., Sweden, and in part during time when he should have been taking care of his daughter Sandra, for which his wife Suzanne A. McEnroe is gratefully acknowledged.

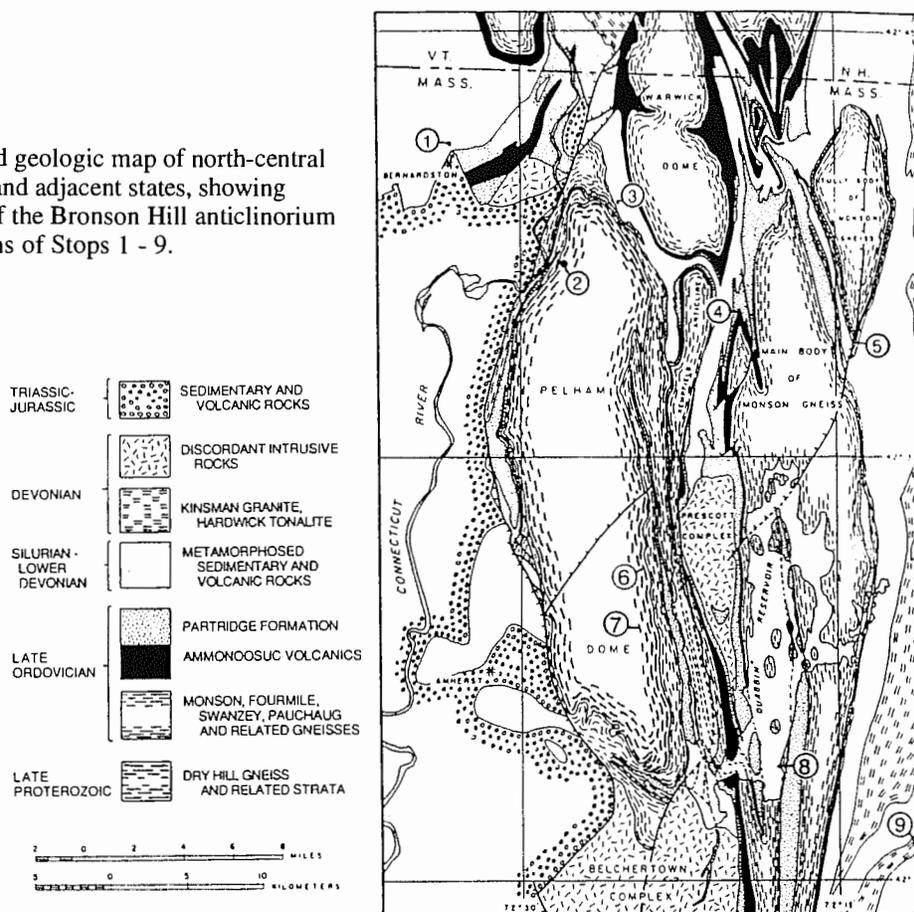
### ROAD LOG FOR DAY 2

From the University of Massachusetts Campus Center proceed via Route 116 and Interstate 91 to junction of Interstate Route 91 and Massachusetts Route 2, west of Greenfield, where the mileage log begins.

**0.0** Proceed north on Interstate Route 91. **2.5** Route 2 bears right. Stay on I 91 North. Cuts in Sugarloaf Arkose. **7.6** Take Exit 28B for Route 10 South. **7.8** Bear right (west) on Route 10 South. Cross over I-91, enter village of Bernardston. **8.5** T junction with U. S. Route 5. Turn right (north). **9.2** Pull onto grassy shoulder on right (east) side and park. Cross road. Walk up short dirt track, through gate into pasture. **CLOSE GATE BEHIND YOU!!** Most stops are on private property. Please respect outcrops and the rights of landowners.

**STOP 1. BERNARDSTON FOSSIL LOCALITY** (55 minutes) See Figure 2.2; for complete stop description and maps see Robinson and Elbert (1992), Elbert (1988). Outcrops in lowest part of pasture are in the

Fig. 2.2 Generalized geologic map of north-central Massachusetts and adjacent states, showing gneiss domes of the Bronson Hill anticlinorium and the locations of Stops 1 - 9.



lower member of the Clough Quartzite, gray weathering phyllite containing quartz-muscovite-chlorite-biotite-garnet-graphite-leucosene typical of the lower- to middle-garnet zone. Outcrops approximately halfway up the pasture are metamorphosed conglomerates of the middle member of the Clough with a few schistose clasts as well as an impure schistose matrix. Cross barbed-wire fence at edge of woods at northwest corner of pasture and continue north on logging road. Outcrops on west side of road, are clean, quartz-pebble conglomerates with quartzite matrix, typical of Clough as mapped by most workers. Pebbles are uniformly vein quartz. Proceed to point where road takes nearly right-angle bend to the west, the location of the no-longer-exposed, stratigraphically highest part of the Clough with its fossiliferous section interpreted to be a metamorphosed marine deposit with stenohaline, marine invertebrates (Boucot et al., 1958) of Llandoverian to Wenlockian age. Marbles of the Fitch Formation are exposed in pits dug during mining of a magnetite bed in the 18th and 19th centuries. Here it is chiefly white to light-gray calcite marble containing scattered crinoids and coral fragments. Over 1000 recognizable conodonts were recovered from 129 kg of the marble, the world's largest and highest grade (CI=8) collection of regionally metamorphosed conodonts (Elbert et al., 1988), indicative of the Lochkovian (earliest Devonian) *woschmidti* to *eurekaensis* Zones. Lensoid bedforms, relict crossbedding(?), abrupt grain size variations, wide size range of bioclasts, and the fossils indicate deposition in a nearshore, shallow water, generally high-energy, marine environment and suggest that the Fitch near Bernardston was a carbonate shoestring sand formed as a channel filling, bar, or beach. A bed of magnetite-chlorite-quartz-garnet granulite, to 0.8 m thick, is present in several pits. One small outcrop of gray schist and quartzite of the Littleton Fm is at the north edge of the pits. The map pattern of the Fitch suggests that it unconformably overlies the Clough and is unconformably(?) overlain by Littleton. The south dip of units, clear-cut stratigraphic distinctions and paleontological ages show the rocks are inverted. This inverted belt continues tens of kilometers north along strike as the inverted limb of the Bernardston nappe, lowest of the Acadian fold nappes at this latitude.

Retrace route to bus. Leave fences, gates as found. Continue north, Route 5. **9.7** Right onto Burke Flat Road. **9.9** Bridge over Interstate 91. **10.0** Junction, Bald Mountain Road, bear right. **11.3** Junction Route 10, turn left (Route 10 north). **14.3** Outcrops, sulfidic schists and amphibolites, Partridge Formation. **15.0** Bennett Meadow Bridge, Connecticut River. **15.6** Cross RR bridge. Lower Jurassic conglomerate, Turners Falls Sandstone. **15.7** Flashing light, stop sign, turn right (south) on Route 63. **20.4** Bridge, Fourmile Brook. Type locality of Fourmile Gneiss (Ashenden, 1973). **21.0** Turn left (east) into entrance to Northfield Mountain Pumped Storage Hydroelectric Project. **21.2** Junction for Visitor Center on right. Stay straight. **21.4** Fork. Bear left toward warehouse. **21.5** Fork. Bear right (south) to warehouse. **21.7** Turn around in lot and park.

**STOP 2. DRY HILL GNEISS HORNBLENDE MEMBER OVERLAIN BY POPLAR MOUNTAIN QUARTZITE AND GNEISS** (25 minutes) See Figures 2.2 and 2.3; for more complete description see Robinson et al. (1992) stops 1 and 2. The cuts extend from the parking lot east and then south around the warehouse toward the mouth of the Access Tunnel to the underground powerhouse. Here we are at the top of a W-dipping layer of Dry Hill Gneiss 750 ft. thick (Figure 2.4) that is interpreted as the core of an east-directed recumbent anticline. At Stop 2 the Dry Hill (Hornblende Mbr.) is in direct contact with the overlying Poplar Mountain Quartzite, exposed in the drainage area at the bend in the fence. The contact is near the warehouse parking lot, and the sequence is locally repeated by a NW-plunging late asymmetric fold. The Dry Hill Gneiss here is typical of the Hornblende Member (Ashenden, 1973; Hodgkins, 1985), with interlayered gray-pink biotite-feldspar gneiss and pink hastingsite leucogneiss, containing quartz, microcline, oligoclase, green biotite, hastingsite, and accessory allanite, sphene, epidote, calcite, garnet, apatite, zircon, and pyrite. Widely spaced megacrysts of pink maximum microcline (Laird, 1974) as large as 15 cm across with local graphic quartz intergrowths may be fragments from tectonically dismembered pegmatites. Tiny garnets appear to show metamorphic growth zoning. The bulk composition is typical of an A-type alkalic granite, including an extremely high FeO/MgO ratio, or an equivalent rhyolite, and is very close to the Potter Hill Gneiss in the Hope Valley terrane, SE Connecticut (Hodgkins, 1985). The Dry Hill is considered to be a volcanic sequence produced in an extensional environment.

The prominent mineral lineation trends N-S parallel to the dome axis and to the lineation in mantling strata of the dome. In many outcrops it has suffered great circle rotation by late asymmetric folds (Ashenden, 1973; Onasch (1973). Lineation development preceded some asymmetric folds, followed others and was part of the same process. The asymmetric folds, variably overturned to SE, S, or SW, show a "separation angle" of 20° that is bisected by the mineral lineation; thus the lineation is parallel to their transport direction. The folds were in process of rotation into parallelism with the lineation, but did not achieve it so completely here in the dome as they apparently did in the mantling strata. The consistent asymmetry of these folds throughout the dome, regardless of structural or stratigraphic position, implies relative southward sliding of the cover relative to the core. Such a sense is consistently demonstrated by microfabrics (Reed and Williams; 1989; Reed, 1993). These studies and the isotopic dating imply that this major strain field is Pennsylvanian.

After stop return north. 21.8 Bear left at three-way junction. Return to Route 63. 22.3 Junction with Route 63. Turn right (north). 23.0 Cross north over Fourmile Brook. 25.7 Sharp right turn, South Mountain Road. Sign for Linden Hill School. 26.9 Road ascends dip slope, west limb of Pelham dome. 28.5 Four-way junction. Straight up hill on Sky Farm Road. 28.9 Crest of hill. Park to right and walk up driveway on left (north) side for outcrops of Stop 3.

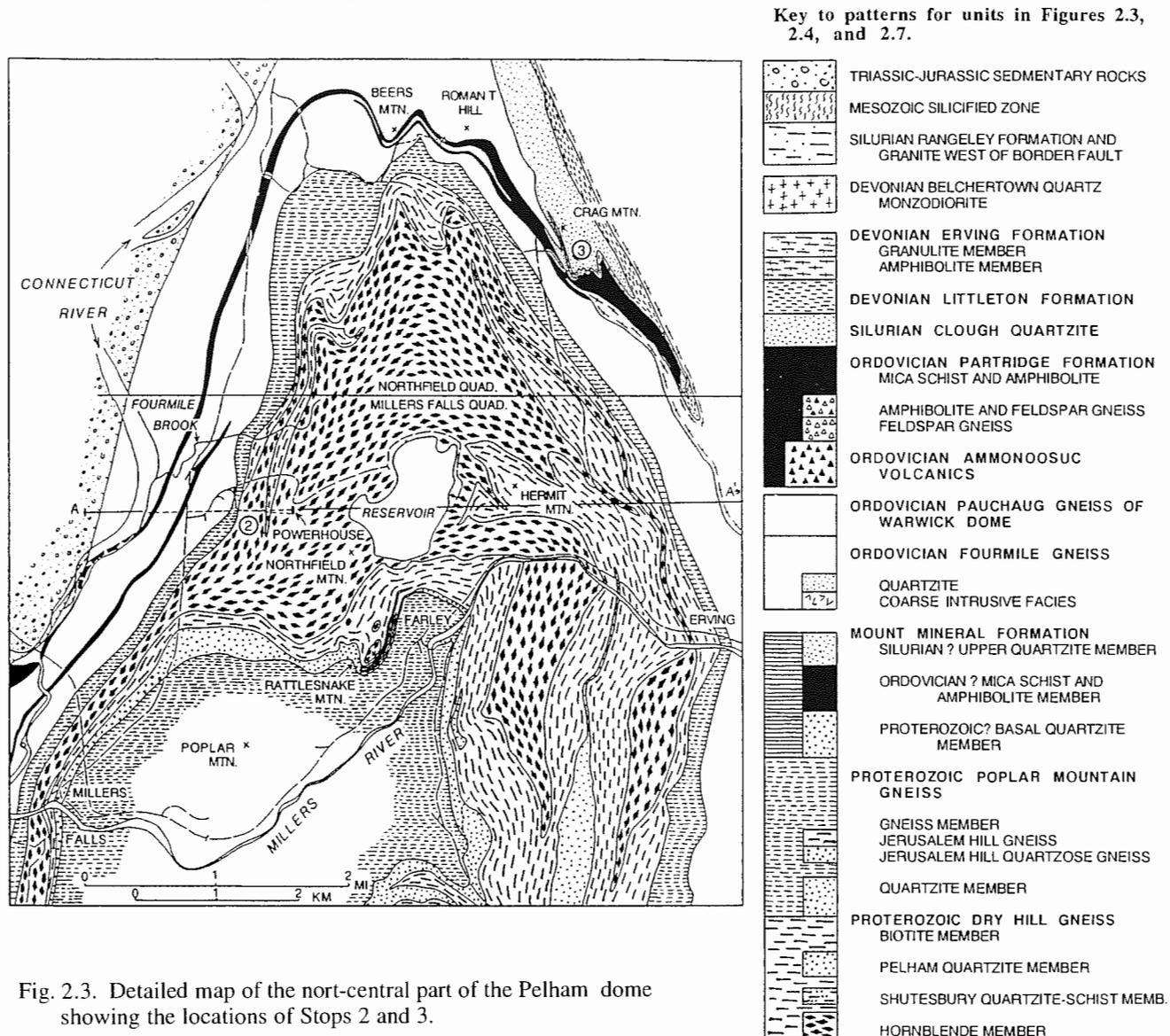


Fig. 2.3. Detailed map of the north-central part of the Pelham dome showing the locations of Stops 2 and 3.

**STOP 3. CLOUGH QUARTZITE AND DEFORMED CONGLOMERATE, LITTLETON SCHIST AND FOURMILE GNEISS IN RECUMBENT FOLDS AT CRAG MOUNTAIN (55 minutes)** See Figures 2.2, 2.3, and 2.4; for more complete description see Robinson et al. (1992) Stop 4. Crag Mountain and Brush Mountain are held up by massive Clough Quartzite and conglomerate involved in a series of east-directed recumbent folds, probably of the back-fold stage, that involve all stratigraphic units from the Fourmile Gneiss to the Lower Devonian Erving Formation. On the basis of widely spaced hinge locations, the folds appear to trend north-northeast at an angle to the dome axis (see Laird, 1974, Fig. 20).

Follow driveway (the Metacomet Trail; white blazes) north to first outcrop on left. **a.** Coarse gray musc-biot-garn-staur-graph-ilmenite schist of Littleton Fm; Crag Mountain syncline. North-plunging crenulation and mica

lineation. This unit contains typical growth-zoned garnets (Tracy et al., 1976; Robinson, 1991, Fig. 6). Reed and Williams (1989) have shown that garnets and staurolites were growing in these rocks during north-over-south shearing. Recently Tucker has obtained a concordant monazite age from the same rock of  $295 \pm 3$  Ma. **b.** First outcrop on right. Foliated and lineated pegmatite with broken feldspars in Clough on east limb of Crag Mountain syncline. Slightly to north and east is deformed conglomerate with sword-like stretched pebbles parallel to fold axes. **c.** Proceed north on driveway. Where it turns left and Metacomet Trail turns right, go straight on older trail. Where older trail turns right (east) up steeper slope, turn sharply west off trail to outcrops on brink of hill 75 feet away. Clough Quartzite containing a foliated pegmatite sill is here in direct contact with Fourmile Gneiss of the Pelham dome, in the core of the Jacks Brook recumbent anticline. The contact, well exposed for 3 mi. north of here, is interpreted as an unconformity (Emerson, 1898; Balk, 1956; Robinson, 1963). The Fourmile is a dark type, with strong N-S lineation consisting of quartz, andesine, microcline, biotite, and hornblende with accessory epidote, garnet and apatite. **d.** Return to trail and proceed up hill to schist in trail, a tectonic inclusion of Littleton in Clough. **e.** Continue up steep trail to "The Crag" of Clough conglomerate. View to E, SE, and S, includes Mount Grace (1617'), formed of Ammonoosuc Volcanics up to 4000' thick in a cross fold on the east flank of the Warwick dome. Mt. Monadnock (3165') with its satellites Little Monadnock (1883') and Gap Mountain (1862') lie to the right of Mt. Grace. Farther on the eastern horizon are the Pack Monadnock Group, Watatic (1832') and Wachusett (2006') on the west contact of the Fitchburg plutons. **f.** Summit of Crag Mountain (1503'). Best stretched pebbles and cobbles, oriented as a result of Pennsylvanian shearing. Type area for "northfieldite" a siliceous border of the "Pelham granite" (Emerson, 1915, 1917) here in "pseudoconglomeratic" facies. Pebbles are >90% vein quartz with minor quartzite and black quartz-tourmaline vein material suggesting deep chemical weathering of the nearby Ordovician source. Pebbles and cobbles are sword-blade-like ellipsoids or cigars with extreme elongation parallel to fold axes. On one outcrop, five miles north, average axial ratios of 1:5:27.5 were measured, giving axial ratios of 0.19:1.01:5.24 to a unit sphere of equal volume, and an elongation of five parallel to fold axes. This lineation is the transport direction of a group of asymmetric folds inside the Pelham dome that are rotated toward the lineation. At Crag Mountain all folds seem to be coaxial, suggesting nearly total rotation into the transport direction. All these features indicate Penn. shearing and elongation parallel to the orogen axis. Extensive view west and southwest. To SW, N-plunging foliation arch of Pelham dome is expressed in topography around Northfield Mtn. Reservoir

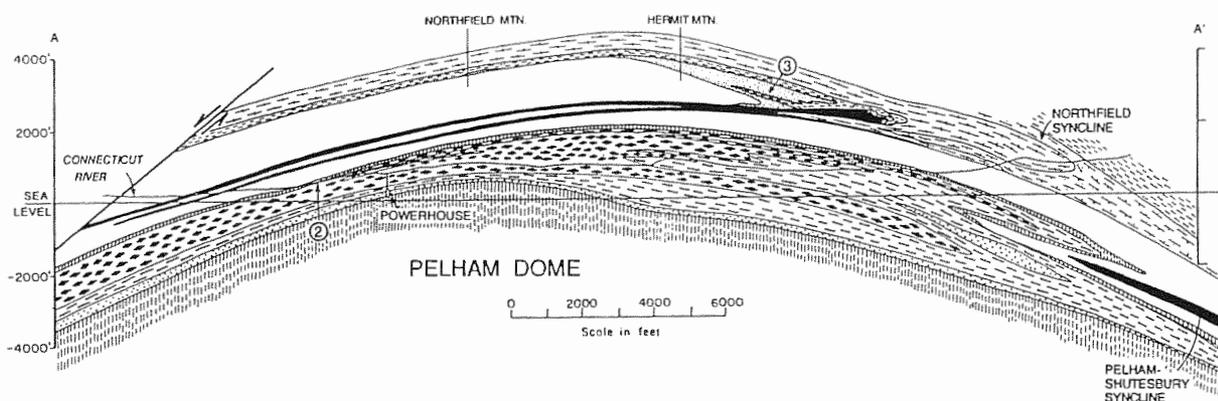


Figure 2.4 Cross section across the northern part of the Pelham dome along line A-A' modified from Ashenden (1973), showing in projection the structural positions of Stops 2 and 3.

Drive east down steep slope. **29.5** Sharp right (south) on Gulf Road. **32.1** Fork. Bear left. **32.3** Turn left (east) onto Route 2, center of Erving. **33.9** Bear left on Rt 2. **34.2** Monson Gneiss and pink Pennsylvanian pegmatites, Kempfield anticline. **34.4** Bridge over Millers River. **34.8** Orange town Line. **35.3** Exit 14; stay on Route 2. Outcrops in ramp on right, Littleton Fm. **36.6** Pull off on right beyond road cut and just beyond passing lane. Cross to east side of highway at first opportunity.

**STOP 4. PARTRIDGE FORMATION AND AMMONOOSUC VOLCANICS IN THE LAKE MATTAWA SYNCLINE** (30 minutes) See Figure 2.5. This stop is to examine typical exposures of these formations at a logistically favorable location. They are in the lower sillimanite zone, on the western limb and along the axis of the Lake Mattawa syncline, a south-plunging Acadian dome-stage syncline in structurally inverted rocks on the overturned northwest flank of the main body of Monson Gneiss. The road cut is in sulfidic schist of the Partridge with quartz-muscovite-biotite-garnet-sillimanite-plagioclase-graphite-ilmenite-pyrrhotite; staurolite is

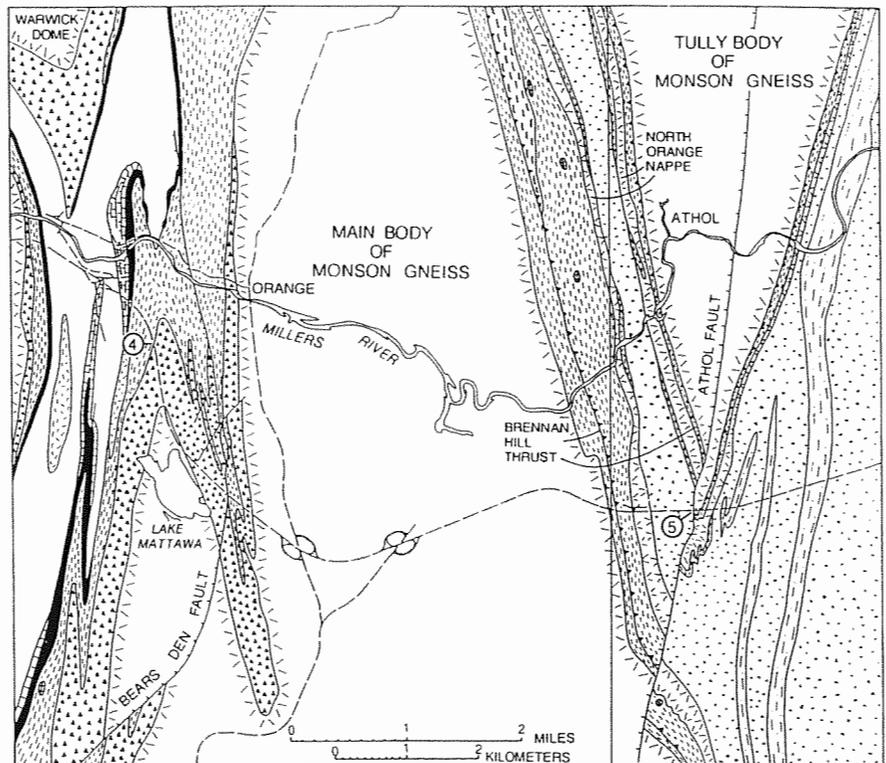
## ROBINSON, RATCLIFFE AND HEPBURN

present nearby. (Also not represented are amphibolite and felsic gneiss, the typical volcanic interbeds that make up substantial parts of the Partridge.) A mica and sillimanite lineation plunges 10-30° S parallel to minor and major folds of the Acadian dome stage. Monazite in the Clough Quartzite 1 mi. north has given a U-Pb age of 360 Ma.

The rest of the stop is a traverse up a gentle hillside, stratigraphically downward in the Ammonoosuc but structurally upward to the axis of the Lake Mattawa syncline, and then to exposures on the east limb. Four rocks on this traverse were part of Schumacher's (1988) study of the geochemistry of the Ammonoosuc. None were studied in detail mineralogically. First there are exposures of quartz-feldspar gneiss of the Felsic Upper Member including one with protruding quartz-sillimanite nodules believed to be a hydrothermally altered felsic volcanic. After a narrow covered interval, where the Middle Member should be, bear slightly SE to exposures of typical hornblende and hornblende-anthophyllite amphibolites of the Mafic Lower Member, inferred to be metamorphosed low/K tholeiites. These include one large exposure of a south-plunging fold hinge with south-plunging hornblende lineation, probably the Lake Mattawa syncline itself. Then move northeast to an exposure on the east limb, of the Middle Garnet-Amphibole Quartzite Member, that is probably of volcanic exhalative origin and forms a key marker horizon between older predominantly mafic and younger predominantly felsic volcanism. Then move northwest to coarse gedrite-garnet gneisses typical of parts of the Lower Mafic Mbr. close to the synclinal axial plane. Some gedrite-garnet and gedrite-cordierite gneisses of the Lower Ammonoosuc are basaltic rocks that underwent extensive hydrothermal sea water alteration before regional metamorphism (Schumacher, 1988). Descend to bus.

Continue east on Route 2. 37.4 Lake Mattawa on right. 38.2 Road cuts, both sides. Partridge schist and amphibolite. 38.7 Exit 15; stay on Rt. 2. 40.1 Rt. 202 enters from south; continue on Rt. 2. 41.7 Bushy cut on right, east part of main body Monson Gneiss. 42.5 Large, long trench cut, both sides. If traffic permits, and you intend to proceed west after this stop, **bear left across highway to broad and firm grass strip, north side.** Much safer and better for viewing than the narrow area on south side. Since Fig. 2.6 was drafted (Robinson, 1979) south side of Athol cut was blasted back about 10 ft., changing some details but not essentials.

Figure 2.5 Generalized geologic map of the central part of the Orange quadrangle and the west edge of the Athol quad. showing locations of Stops 4 and 5. All obvious fold hinges plunge 10-30°S and formed during the Acadian dome stage. In west part of map these folds are in previously inverted strata, so younger rocks appear in anticlines; older in synclines. The Monson Gneiss main body lies along the Williams Pond syncline. In east part of map the North Orange nappe and Brennan Hill thrust are repeated across the Temple Hill fold, believed formed in the backfold stage. South end of the Tully body is a S-plunging anticline. Key to patterns: Monson Gneiss - ticked marked border; Ammonoosuc Volcs. - triangles; Partridge Fm.- fine dashes; Partridge Fm. ultramafic pods - cross pattern; Clough Quartzite - black; Fitch Fm - brick pattern; Rangeley Fm. gray schists - heavy dots; Rangeley Fm. sulfidic schists - dash-dot; Littleton Fm. - no pattern; Acadian granite - uniform ticks.



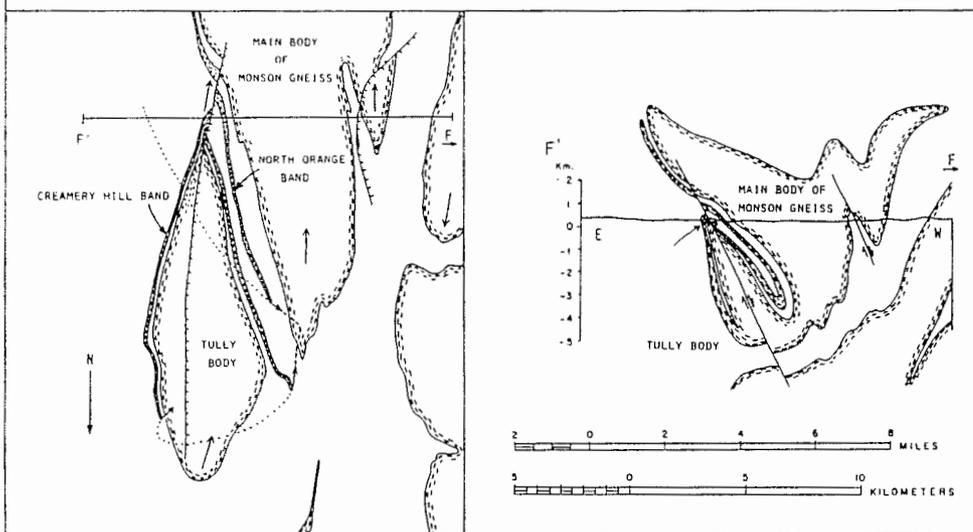
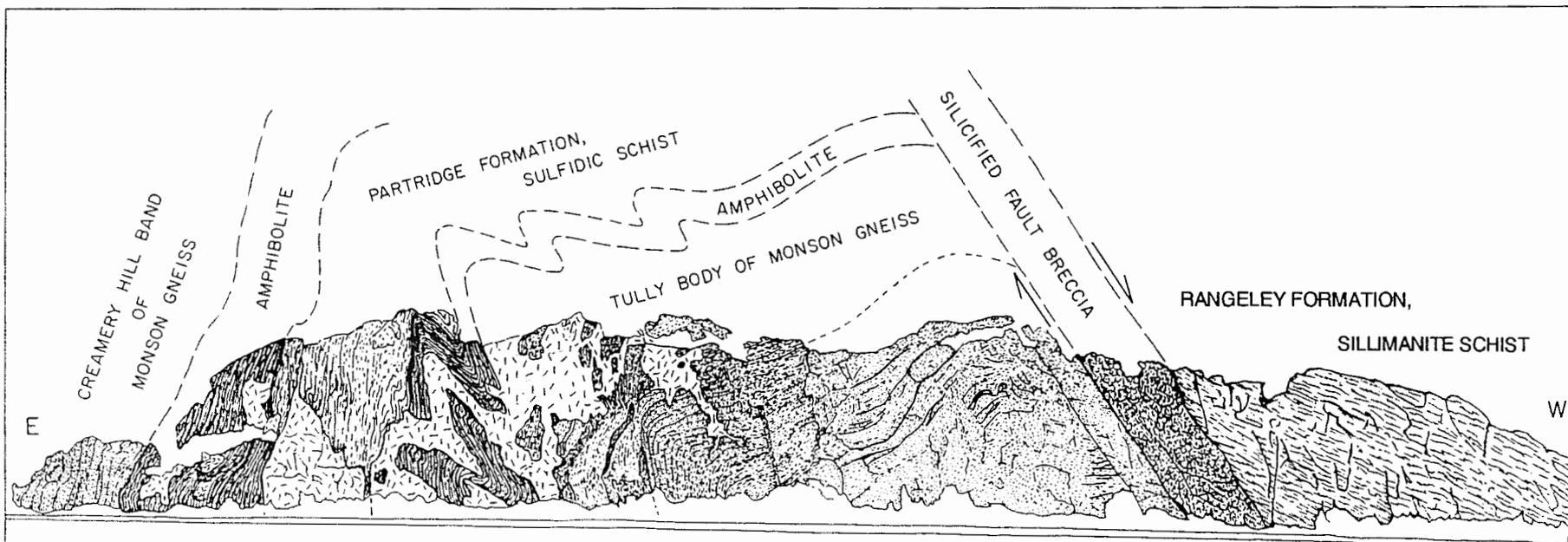


Figure 2.6 Sketch of south wall of rock cut at Stop 5. Mesozoic Athol fault cutting crest of Tully body of Monson Gneiss. Approximate location of the cut is shown by nearly invisible stippled rectangle indicated by arrow in inset cross section. The major anticlinal or dome axis and satellite folds are parallel to a strong Beta maximum (10% per 1% area) with trend N22E and plunge 18°SW (Robinson, 1963). The Creamery Hill band of Monson Gneiss is interpreted (see insets) as extremely attenuated basement nappe, the North Orange nappe, separated from the main and Tully bodies by an extremely attenuated isoclinal syncline and by the Brennan Hill thrust. The North Orange nappe is interpreted as a fold of the same generation as the Bernardston Nappe but lying tectonically higher and more easterly (Robinson et al., 1991, Fig. 30). Drawing does not have constant scale. Outcrop is approximately 50 feet (15 meters) at highest point.

**STOP 5. SOUTH END OF TULLY BODY OF MONSON GNEISS, BRENNAN HILL THRUST, AND NORTH ORANGE FOLD NAPPE, CUT BY MESOZOIC NORMAL FAULT** (30 minutes) See Figures 2.5 and 2.6; for details see Robinson and Elbert (1992) Stop 7. The south end of the Tully body exposed here is a simple anticline overturned to the east and plunging gently SSW. Dome-stage asymmetric folds occur on both limbs. It is cut just west of its crest by the W-dipping Athol normal fault, bringing gray schists of the Rangeley Fm on the west limb down about 1300 ft. into contact with Monson Gneiss of the core. The fault zone is about 10' wide and contains gouge zones, breccia, intense silicification, and vuggy quartz veins, all characteristic of known Mesozoic faults. Here the fault strikes N4W, but regionally it strikes N15E.

A thin layer of rusty Partridge schist on the east limb of the anticline has been traced for many miles. Bedded amphibolites at both contacts probably belong to the Partridge, although it is tempting to consider that they might represent Ammonoosuc Volcanics. Next east of the schist and amphibolite is a layer of Monson Gneiss (Creamery Hill band) that has been traced entirely around the southern part of the Tully body (see Figure 2.6 inset) and is interpreted as an early anticlinal fold nappe of gneiss that formed contemporaneously with the Bernardston nappe. A similar band of Monson Gneiss (North Orange band) east of the main body of Monson Gneiss is interpreted as the same recumbent anticline repeated by folding about the same syncline that separates the two bodies. The North Orange band has been traced from North Orange to the northern edge of the Palmer quadrangle in southern Massachusetts. Three anticlinal hinges for this early fold, now named the North Orange nappe, can be seen in Figure 2.6 inset. It has been severely involuted, both by dome-stage folding and by the complex backfolding that included northward transport of the main and Tully bodies (see Robinson et al., 1991, Figure 9). By unwinding these deformations one can conclude that the original transport direction for this nappe was east to west. Elsewhere it can be shown that the North Orange nappe lies above the Brennan Hill thrust, hence the thrust must also lie in this outcrop, probably close to the contact between the Partridge and the gneiss in the core of the Tully body.

Return west on Route 2. **45.3** Take Exit 16 for U. S Route 202 South and stay on it to Stop 6. **49.5** Small cut, Monson Gneiss, right. **55.4** West Branch, Swift River. **56.8** Low outcrops, left, Partridge in extremely thin Pelham-Shutesbury syncline. **57.2** Cuts, Fourmile Gneiss. **57.7** Descend long straight incline and pull off into paved ditch on right just before steep gravel driveway. **LEAVE FLASHERS ON.** Walk up driveway about 30 feet, turn sharp left (west) on old road and descend to Atherton Brook. Cross brook on stepping stones and walk about 200 yards northwest to small outcrops about 100 feet southwest of stream bank (Ashwal Loc. 160).

**STOP 6. SCHIST OF MOUNT MINERAL FORMATION WITH RELICT ACADIAN GRANULITE FACIES MINERALS OVERPRINTED BY PENNSYLVANIAN KYANITE-MUSCOVITE ASSEMBLAGE** (30 minutes) See Figure 2.7; detailed descriptions in Robinson et al. (1992) The western outcrop consists of coarse mylonitic sillimanite-orthoclase-garnet-biotite schist with north-plunging Penn. dome-stage anticlinal fold superimposed on an older west-over-east shear fabric. Locally sillimanite lineations parallel the fold and the earlier fabric. The garnet with extreme resorption zoning of Roll (1987; Robinson, 1991) and orthoclase of high structure state came from this outcrop. The relict assemblage includes sillimanite, orthoclase, biotite, garnet (pyrope 35, spessartine 1.1) and rutile, indicative of about 700°C and 6.8 kbar. Garnet rims fall off to pyrope 10-11. They are commonly in a matrix of muscovite, biotite, and kyanite indicative of more hydrous re-equilibration conditions at about 580°C and 6 kbar. Orthoclase is commonly rimmed by muscovite and sodic plagioclase against the aluminous matrix. The schists at this outcrop successfully resisted Penn. recrystallization due to scarcity of water, and they contain an older E-W linear fabric involving oriented sillimanite, highly elongate quartz ribbons, tails around orthoclase and garnet, and evidence that orthoclase was undergoing strong size reduction with limited recrystallization. The granular top surface of the outcrop is sillimanite-orthoclase-garnet-rutile mylonite with subordinate biotite, in which quartz, plagioclase and orthoclase all underwent severe grain-size reduction with minimal recovery. These features are evidence of an earlier higher temperature phase of shearing with top-side-east, probably soon after the peak of Acadian granulite facies metamorphism. A U-Pb age of  $367 \pm 2$  Ma was obtained from metamorphic monazite in this rock. The eastern outcrop consists of mica schist containing kyanite, staurolite, muscovite, biotite, and euhedral garnets, that is believed to be chemically equivalent to the western outcrop but has undergone hydration and complete chemical reconstitution during Pennsylvanian metamorphism and deformation. The garnets have the characteristics of ones that have undergone prograde growth zoning. In this outcrop there is strong evidence of north-over-south shearing parallel to a subhorizontal Pennsylvanian mineral lineation.

Continue south, Route 202. **58.1** Cut on right. Early Jurassic Pelham Fe-rich tholeiite dike 15 m thick in schist, Mount Mineral Formation. **58.3** Cuts on right. Amphibolites, Mount Mineral Formation. **59.8** Cut on right, turn out, left. View of Quabbin Reservoir and Mt. Wachusett (2006').

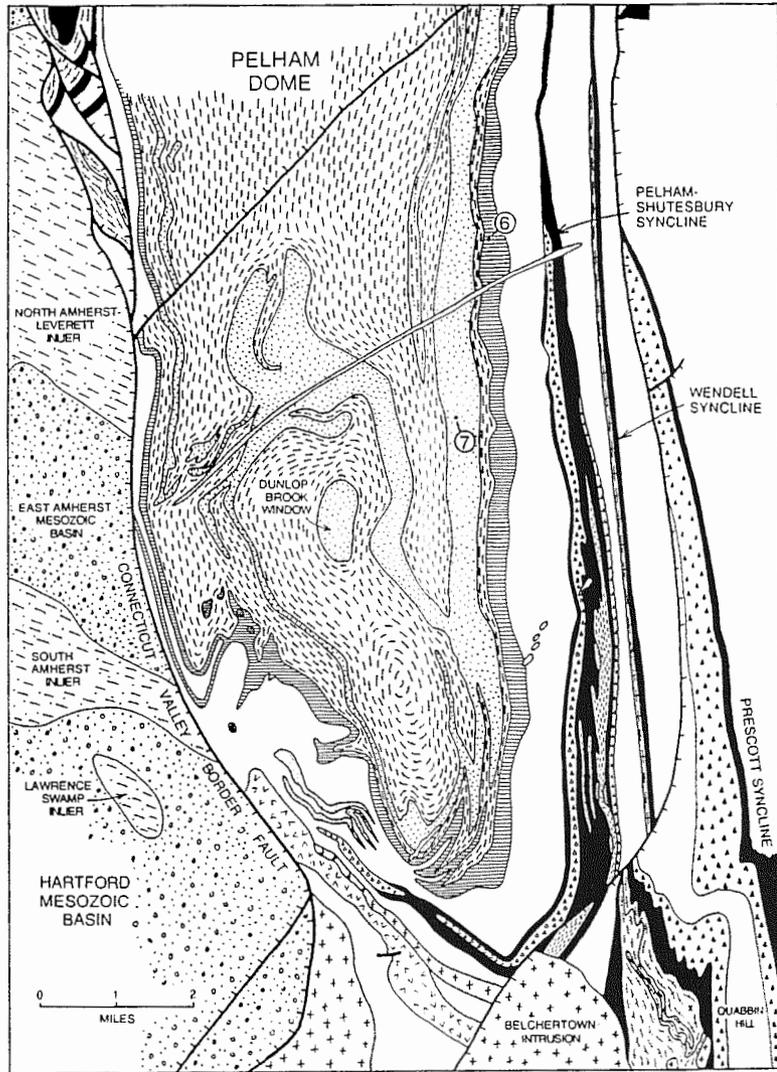


Figure 2.7 (left). Generalized geologic map of the southern part of the Pelham dome (Robinson et al., 1992) For key to pattern see Figure 2.4.

**KEY TO PATTERNS**  
**FIGURE 2.8**

- Monson & Leadmine Pond (SE) Gneisses - tick border
- Ammonoosuc Volcs. & Partridge Fm - fine dark dash
- Rangeley Fm. sulfidic schist - no pattern
- Rangeley gray schist - horizontal rule
- sulfidic calcisilicate Francestown Fm (W of Coys Hill); extremely sulfidic schist of Smalls Falls & Rangeley Fms.- black
- Madrid Fm. - fine dots
- eastern facies Paxton Fm. - coarse dots
- east facies gray schist - curvy dash
- eastern facies sulfidic schist - fine light dash
- Coys Hill Granite pluton - large dash
- Hardwick Tonalite - double dash border
- tonalites, diorites & gabbro - crosses

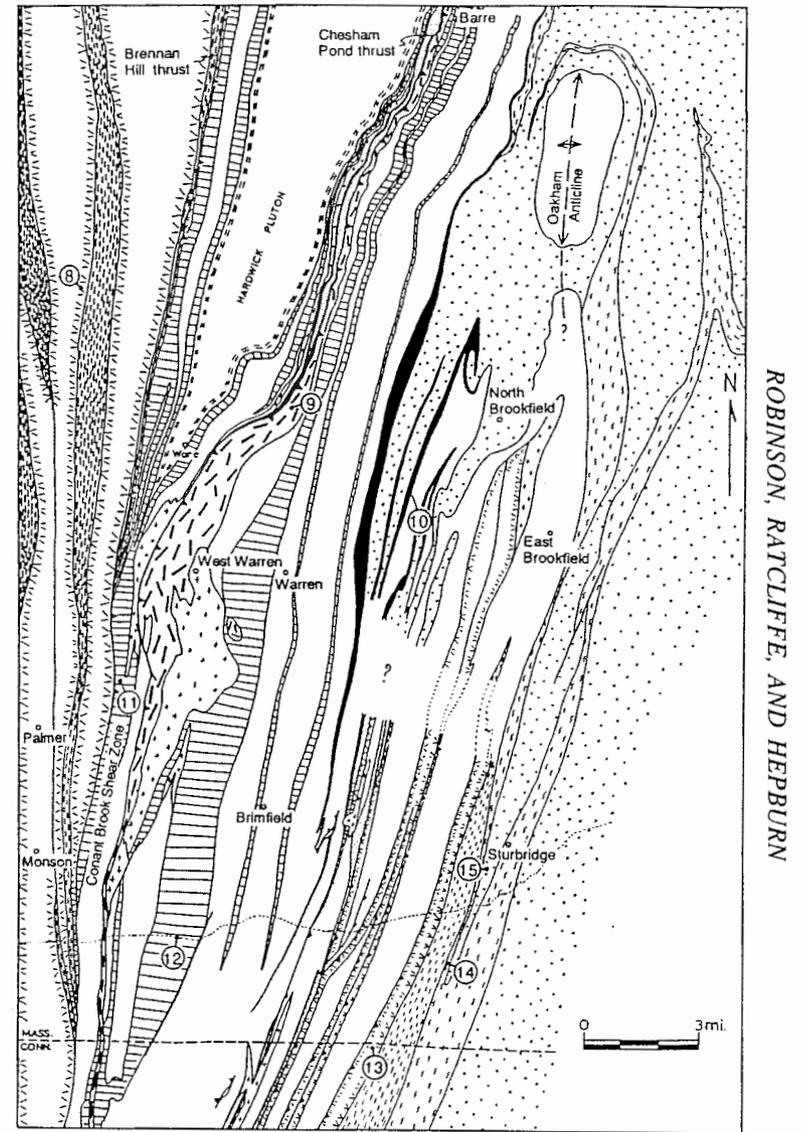


Figure 2.8 (right) Generalized geologic map of part of south-central Mass. with Stops 8-15 (modified from Robinson et al., 1989) shows proposed southward extensions of the Brennan Hill and Chesham Pond thrusts, reinterpretation in Brimfield-Sturbridge area (Berry, 1989) and Conant Brook shear zone (Peterson and Robinson, 1993).

**STOP 7. PELHAM QUARTZITE ON EAST LIMB OF PELHAM DOME** (optional; 10 minutes) See Figure 2.7; rock and enclosed zircons obtained are discussed in detail in Robinson et al. (1992). The Pelham Quartzite commonly contains actinolite indicative of a former dolomite cement. Eighteen detrital zircons from this outcrop and another 12 detrital zircons from a quartzite at Dunlop Brook, were analysed as single crystals. Those from this outcrop are < 6% discordant, hence indicate reliable source ages. Combined with somewhat more discordant results from Dunlop Brook, the detrital grains fall into the same age groups found in the Dry Hill Gneiss at Stop 2, the oldest between 2616 and 2679 Ma. The youngest detrital grain has a  $^{207}\text{Pb}/^{206}\text{Pb}$  age of 933 Ma, establishing a *maximum* depositional age of the quartzite. Late Proterozoic, ca. 700-600 Ma, detritus is apparently lacking in the Pelham, Dunlop Brook, and Poplar Mountain Quartzites, implying the Dry Hill Gneiss was not a major source of detritus nor were any other rocks of Late Proterozoic (Avalonian) affinity. Good agreement between age groups defined by the detrital-grain and composite-grain studies suggests that the source of inheritance in the Dry Hill (Stop 2) was detritus at the site of eruption, with sources in Early and Middle Proterozoic rocks.

**60.5** Continue Rt. 202 past yellow light at Amherst Road. **67.9** Lights, Junction Rt. 202 and Rt. 9. Turn left (east) Rt. 9. **71.0** Turn left (NE); enter Quabbin Reservation. **71.6** Headquarters and Visitor Center; cross Winsor Dam, view west arm of Quabbin Reservoir. **72.1** Cross spillway underlain by deformed Spillway stock of Belchertown intruding L. Dev. Erving. Turn left (north) toward Quabbin Hill. **73.5** Direct through rotary. **73.9** Enfield Overlook, left. View over Prescott Peninsula to Mt. Monadnock, NH. **76.0** Left (east) toward Goodnough Dike. **76.2** Bear left at fork. **76.5** Turn around at rotary and park, west end of Goodnough Dike. View, east arm of Reservoir, underlain by main body of Monson Gneiss. Small exposure in lieu of Stop 8 if high water. **77.0** Turn left (south) on Reservation Road toward Ware. **78.0** Reservation Rd and Rt. 9. Turn left (east) on Rt. 9. **82.9** Junction, Greenwich Plain Road. Bus parks and participants transfer to vans for one-mile drive to stopping point for Richards Ledges. Vans will be taken through locked Gate 49 by arrangement with reservoir management. Ledges can be reached on foot in 30 minutes of walking in area of public access. Follow main gravel road straight until shore of Reservoir can be seen. Ledges reached by walking south about 5 minutes.

**STOP 8: MONSON GNEISS AT RICHARDS LEDGES** (40 minutes) See Figure 2.2; characteristics of Monson Gneiss near Quabbin Reservoir discussed in greater detail by Robinson et al. (1989), stop 1. The outcrops in this area are from metamorphic Zone V, within the main body of Monson Gneiss on the west limb of the Greenwich syncline. They are composed of gray tonalitic gneisses, amphibolite, hornblendite, and gabbroic anorthosite. Recent precise radiometric dating of a variety of these rocks by R.D. Tucker has shown an age range from 454 to 443 m.y. (+3/-2 on six best samples, +9/-3 and +11/-8 on two others). Except for rare diopside calc-silicate and quartzite layers, the Monson consists of rocks of broadly igneous composition (Robinson, 1963, 1967; Hollocher, 1987, 1988) as follows: **(A)** Coarse- to fine felsic gneisses, generally massive and intrusive-looking, including dark-gray biotite-hornblende-andesine tonalite, medium-gray biotite-oligoclase tonalite, light-gray biotite-garnet-oligoclase tonalite, light-gray to pink microcline-hornblende augen gneiss, and rare white microcline-albite-quartz-biotite-garnet alaskite. **(B)** Mafic and ultramafic rocks as large and small blocks, commonly included in or interlayered with the more felsic types. Ultramafic rocks include serpentine and talc-bearing lenses, olivine hornblendite, augite hornblendite, and hornblendite (Tracy et al., 1984). Small patches of hornblendite may be residues of partial melting. Mafic rocks include coarse amphibolites that look like massive gabbros, some with relict ophitic texture; fine amphibolites commonly containing patches that are probably recrystallized plagioclase or mafic phenocrysts and commonly cutting metamorphic foliation and layering indicating they are dikes (one such dike in the Keene Gneiss dome is dated at  $381 \pm 3/-2$  (Tucker and Robinson, 1990)); anorthosite and gabbroic anorthosite gneiss (best examples are bluish-gray glacial boulders with attached Monson Gneiss); composite mafic and intermediate blocks with gabbro, amphibolite, hornblendite, and biotite gneiss; rare hornblende-gedrite-garnet-plagioclase-biotite rock; and stratified amphibolite that looks most like metamorphosed layered mafic tuffs. **(C)** Felsic rocks in small intrusions probably or possibly synchronous with Acadian metamorphism such as coarse white biotite-tonalite pegmatite; medium-grained white to pink biotite-garnet tonalite gneiss similar to migmatitic layers in adjacent cover rocks; medium-grained, white to pink biotite-hornblende tonalite gneiss; and fine medium-gray biotite-epidote tonalite. The ultimate source of these melts is the Monson Gneiss itself (Hollocher, 1988).

Field, petrologic, and geochemical studies by Hollocher show the Monson to be a complex of metamorphosed intrusive rock types with only a miniscule supracrustal component. If it is part of a plutonic complex, then the well layered character of many outcrops must be attributed to deformation, either Acadian or earlier. Petrologically and geochemically Monson is distinct from metavolcanics in the physically overlying Ammonoosuc Volcanics and Partridge Fm. but closely resembles rocks in other orogens described as in the roots of a calc-alkaline island arc.

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Road log resumes at junction. **83.7** Bear right, Poor Farm Rd. **85.1** Junction, Rt. 9, Ware Center. Turn left (east) on Rt. 9. **85.5** cut on left; amphibolite in Monson. **86.9** Junction with Rt. 32, Ware; stay on Rt. 9. **87.3** Bridge, Ware River. **88.5** Rt. 32 turns left; stay right on 9. **89.4** Sulfidic calc-silicate, Frankestown Fm. **89.6** Turn across traffic to dirt pull-out on left. Exit on traffic side; walk west on old Rt. 9 past large exposures of Coys Hill granite. Right (north) on trail to picnic site and eastward overhanging outcrops of Coys Hill Granite.

**STOP 9. COYS HILL PORPHYRITIC GRANITE** (20 minutes) See Figure 2.8. This stop is to examine briefly the Coys Hill Granite, a major lithic unit in the Merrimack synclinorium of Mass., equivalent to the Kinsman Granite of New Hampshire. The Coys Hill forms a key marker across central Mass. (see blue unit on State Map). Based on contact relations in the Monadnock area, Peter J. Thompson (1985) has postulated a major early thrust, the Chesham Pond thrust, carrying Kinsman Granite westward over already folded strata. By analogy, it is suggested that the west margin of the Coys Hill Granite is also an early W-directed thrust. Here as in most of Massachusetts the Coys Hill is a coarse gneiss in which K-feldspar phenocrysts have tectonically rounded ends. Abundant garnets are probably metamorphic porphyroblasts. One might mistake the granite for a pegmatite formed during metamorphism. However here and elsewhere, it is cut by later fine-grained tonalite dikes, and both granite and tonalite were deformed and metamorphosed together. A garnet phenocryst age of 413 Ma in New Hampshire (Barreiro and Aleinikoff, 1985), the geochemistry (Clark and Lyons, 1986), the occurrence of gabbro xenoliths, and the consistent textural features regardless of metamorphic grade show that the Coys Hill and Kinsman were major peraluminous intrusions produced by crustal melting induced by mantle-derived melts early in the Acadian orogeny.

Continue east on Rt. 9. **92.7** Rt. 67 enters; stay on Rt. 9. **93.7** Stoplight in West Brookfield. **93.8** Bear left on Rt. 67 at West Brookfield common. **94.6** White farm house (registered Holsteins, Talvy Farm) left. Turn into barnyard to disembark. Bus reverses direction. Walk north on farm road along southeast side of long outcrop.

**STOP 10: SMALLS FALLS FORMATION, SULFIDE-RICH SCHIST** (30 minutes) See Figure 2.8; for more detailed description see Robinson et al. (1989). The outcrop has an irregular smoothed surface covered by a thick crust of iron oxides and sulfates. The surface contains many 3-5 cm pits inside which fresh-looking pyrite is typically visible. We have found only traces of megascopic pyrrhotite in the outcrop but still believe much of the outcrop's character is due to its weathering. Partly weathered rock just beneath the crust looks white because of the abundance of colorless silicates; fresh rock has a bluish look. The outcrop consists of two main rock types, sillimanite and biotite-bearing quartzites, and aluminous schists with variable proportions of sillimanite, biotite and cordierite. Biotites vary from very pale reddish-brown Fe-bearing to colorless Mg end-members. We first thought these were muscovites in a retrograded fault zone, but optics and probe analyses show they are biotites. The assemblage in two analysed samples from this outcrop (FW-882E and FW-882Y of Tracy and Robinson, 1988) is quartz-orthoclase-plagioclase-biotite-cordierite-sillimanite-rutile-pyrite-pyrrhotite-graphite. The XMg of biotite in these samples is 0.995 and 0.999 (less than 0.05 weight % FeO) and they contain 0.065 and 0.074 Ti per 11 oxygens. The cordierites, which are full of quartz, sulfide and graphite inclusions, appear as black to bluish lumps. They are essentially pure Mg end-members (very low to undetectable FeO and MnO) and lack pleochroic haloes around monazite inclusions, presumably due to lack of iron to be oxidized by alpha bombardment.

From barnyard go west (right) on Rt. 67. **95.5** Rejoin Rt. 9 at West Brookfield common. **95.6** Stoplights, continue west, Rts. 9 and 67. **96.6** Left (southwest) on Rt. 67. **98.3** General Henry Knox Monument, left. **99.1** Center of Warren. **101.4** Stoplight, West Warren. **101.8** Junction with Ware Rd.; stay on 67. **103.2** Right on Old Warren Rd. **104.4** Large outcrop, both sides. Park on right (north), beyond first large outcrop; disembark. Bus proceeds about 0.2 miles to turn around, returns on eastbound side for pick up.

**STOP 11: BACK-FOLD AND DOME-STAGE SHEAR FABRICS, RANGELEY FORMATION, CONANT BROOK SHEAR ZONE** (20 minutes) See Figure 2.8. This outcrop (P100 of Peterson and Robinson, 1993) is described in detail in Peterson (1992). It offers a 3-dimensional cross-strike exposure of graphitic and sulfidic schists within the Conant Brook shear zone. It preserves domains in which each of two Acadian lineations and related fabrics are dominant, an E-W trending, steep, backfold-stage lineation, and a later shallow N-S trending dome-stage lineation. Schist about 8-10 m from the west end of the outcrop is dominated by the steep lineation and there is little or no evidence of the shallow lineation. Steep lineation is evident on foliation surfaces in the field and in thin section, defined by abundant coarse sillimanite, elongate quartz grains and the stretching direction of porphyroblast tails. On the front of the outcrop, parallel to this lineation, numerous kinematic indicators suggest a consistent west-side-up sense. Farther west foliation surfaces commonly show both lineations. At the west end of the outcrop, a strong shallow south-plunging mullion-like lineation is evident on the foliation surface, also defined by aligned sillimanite and stretched quartz. Sections cut perpendicular to each finite

strain axis here show no evidence for the steep lineation. On top of the outcrop at the west end, numerous kinematic indicators show a dextral or west-side-north sense of shear. The principal assemblage in the shear zone is quartz-sillimanite-orthoclase-plagioclase-biotite-garnet, and garnet-biotite thermometry suggests temperatures were as high as 750°C. Garnets are chemically homogeneous, except where retrograde ion exchange has occurred near adjacent biotites, suggesting the absence of late metamorphic fluid. A few less-deformed (younger?) pegmatites contain coarse muscovite, otherwise unstable in these schists, and around these the schists are heavily sericitized.

**104.8** Go to bus on south side of road. **106.1** Rejoin Rt. 67; turn right (south). **108.4** Mass. Pike underpass. **110.8** Junction with Rt. 20 East; turn left. **111.3** Brimfield town line. **114.7** Right turn on Hollow Rd. **118.1** Stop sign; turn right (south). **118.2** Stop sign; bear right (south) onto Rt. 19 in Wales. **118.4** Sharp right off Rt.19 onto Monson Rd. **119.2** Bear right onto McBride Rd.; Monson Rd. swings left. **119.6** 3-way junction. Bear right (north) onto Mt. Hitchcock Road. **119.7** Tennessee Gas Pipeline. Site of trench in fall 1985.

**STOP 12. WELL BEDDED COARSE GRAY PELITIC GNEISS OF THE RANGELEY (?) FORMATION, MT. PISGAH** (40 minutes) See Figure 2.8; for more complete description see Thomson et al., 1992, stop 4. The attraction here is very fresh highly deformed granulite-facies pelitic gneiss rich in sillimanite and cordierite with coarse garnets that appear to have formed within felsic melt segregations. Best outcrops and samples occur east along the pipeline over the brow of the hill to the steepest east-facing slope. Gneiss samples from this vicinity typically have garnet Alm 67, Pyr 29, Spes 1, Gro 3, biotite  $X_{Mg} = .56-.63$  and Ti/11 oxygens = .252-.300, cordierite  $X_{Mg} = .76$ , plagioclase An34, estimated temperatures of 710-785°C, and estimated pressures of 5.6-6.9 kbar. A tiny, now obliterated outcrop nearby contained sillimanite pseudomorphs after andalusite up to 3 cm in diameter and 9 cm long in veins of K-feldspar, quartz, and biotite with euhedral garnets up to 2 cm across.

Pegmatites consisting of K-feldspar, quartz, plagioclase, cordierite (0.5mm to 10 cm), sillimanite, garnet, and biotite appear to be the product of in situ partial melting. Pegmatite cordierite typically has  $X_{Mg} = 0.65-0.70$ , except adjacent to garnet where it is 0.80-0.84; in contrast to cordierite within gneisses where  $X_{Mg} = 0.76-0.78$ , except up to 0.82 adjacent to garnet. The unusually Fe-rich cordierites suggest pegmatite melting took place at lower pressure than peak granulite-facies conditions recorded in the gneisses. Estimated conditions of pegmatite genesis are: gar/crd = 714°C; gar/bio = 700°C; qz-sill-gar-crd = 5.5-6.1 kbar (Tracy et al., 1976) or 4.8-5.2 kbar (Bhattacharya, 1986); GAsP = 5.3 kbar. The pegmatite cordierites show internal symplectic intergrowths of garnet (0.25-3 mm)+sillimanite+quartz, skeletal sillimanite+quartz intergrowths against garnet, and symplectites of pale-green low-Ti biotite (Ti/11 oxygens < 0.02)+sillimanite+quartz against large K-feldspar grains. Symplectite garnets in contact with cordierite  $X_{Mg} .79-.84$  are Alm 62.6-68.2, Pyr 26.7-33.2, Spe 1.8-2.2, Gro 2.4-3.1 suggesting gar/crd = 562-601°C, qz-sill-gar-crd = 7-7.5 kbar. Apparently large Fe-rich pegmatite cordierites did not re-equilibrate during peak granulite facies conditions recorded in the gneisses, but did respond by symplectite formation during cooling. Together these rocks record part of a P-T path in which compression with heating appears to have been followed by further compression with cooling, or at least isobaric cooling if newer pressure calibrations are employed.

Turn around and return south on Mt. Hitchcock Rd. Return to Rt. 19 via McBride and Monson Rds. **121.1** Turn left (north) on Rt. 19. **122.3** Bear right (east) on Holland Rd. at blind intersection. **123.2** Wales/Holland town line. **124.1** T junction. Turn right (south) on Brimfield Rd., Holland. **126.2** Crossroads and stoplight, Holland with windmill! Go straight. **127.6** Causeway, Hamilton Reservoir. **129.3** Connecticut State Line Monument. **129.5** Beginning of interchange at Mashapaug Road beyond truck depot. Park on grass strip to right and cross north to south end of large outcrop.

**STOP 13. LEADMINE POND GNEISS AT MASHAPAUG ROAD, PYROXENE GRANULITE OF TONALITE COMPOSITION** (20 minutes) See Figure 2.8; for detailed description and illustration see Robinson et al. (1989) The dominant rock is a dark-gray two-pyroxene granulite in the central part of the outcrop. On the northwestern end are felsic garnet-biotite tonalitic gneiss and intermediate garnet-orthopyroxene-biotite gneiss. On the southeast end are calc-silicate granulites and calcite-diopside-quartz-sphene-scapolite-apatite-biotite pegmatite. Berry (1989) has assigned these rocks to the pre-Silurian basement within an early Acadian thrust slice.

Bear right (south) from parking place. **129.6** Stop sign. Turn left (east) on bridge across Interstate 84. **129.7** Turn left (north) on entrance ramp to Interstate 84 North and re-enter Massachusetts. **133.8** Take Exit 1, Mashapaug Road. Turn right (southwest) toward Southbridge. **134.2** Underpass beneath northbound lane of Interstate 84. **134.6** Turn left (south) into entrance of Sturbridge Isle Truck Stop and large outcrop at back.

**STOP 14. CORDIERITE PEGMATITE IN BEDDED SCHIST AND GRANULITE OF PAXTON FORMATION AT STURBRIDGE ISLE** (optional; 15 minutes) See Figure 2.8; outcrop description adapted from Thomson et al. (1992). It is a cordierite±garnet-bearing pegmatite resulting from fluid-absent, biotite-dehydration melting of surrounding gneisses. The host consists of quartz-sillimanite-garnet-cordierite schists and interbedded biotite and calc-silicate granulites of the Sil.(?) Paxton Fm. The pegmatite assemblage includes quartz, plagioclase, orthoclase, sillimanite, biotite, cordierite, and garnet. The cordierite is blue to dark lavender and up to 8 cm across, locally with large dark patches up to 15 cm across that are pinitic alteration. Some samples show evidence of cordierite breakdown to aggregates of garnet, sillimanite, and quartz. Within 1 mm of these aggregates cordierite is zoned from a normal X<sub>Fe</sub> of 0.30-0.31 to as low as 0.22 near garnet. Garnet is also slightly zoned. Cordierite breakdown appears to have begun at 760°C/5 kbar and continued to 615°C/6 kbar.

**134.9** Re-enter Mashpaug Road from truck stop and turn right (northeast). **135.5** Turn left into Interstate 84 East. **136.9** Take Exit 2, Old Sturbridge Village Road. **137.2** End of ramp. Turn left (west) on Old Sturbridge Village Road. **137.6** Wide space in road with outcrop on right at back gate to Village.

**STOP 15. ISOCLINAL FOLDS IN BEDDED GRANULITE OF PAXTON FORMATION** (optional; 10 minutes) See Figure 2.8. Gray biotite-garnet granulites, and calc-silicate granulites of the Paxton Formation with abundant pegmatite and tight isoclinal folds.

After Stop 15 return to Interstate 84 East which leads to Massachusetts Turnpike Interstate 90 East and Worcester via Auburn Exit for overnight.

### DAY 3: EASTERN MASSACHUSETTS TERRANES

by J. Christopher Hepburn

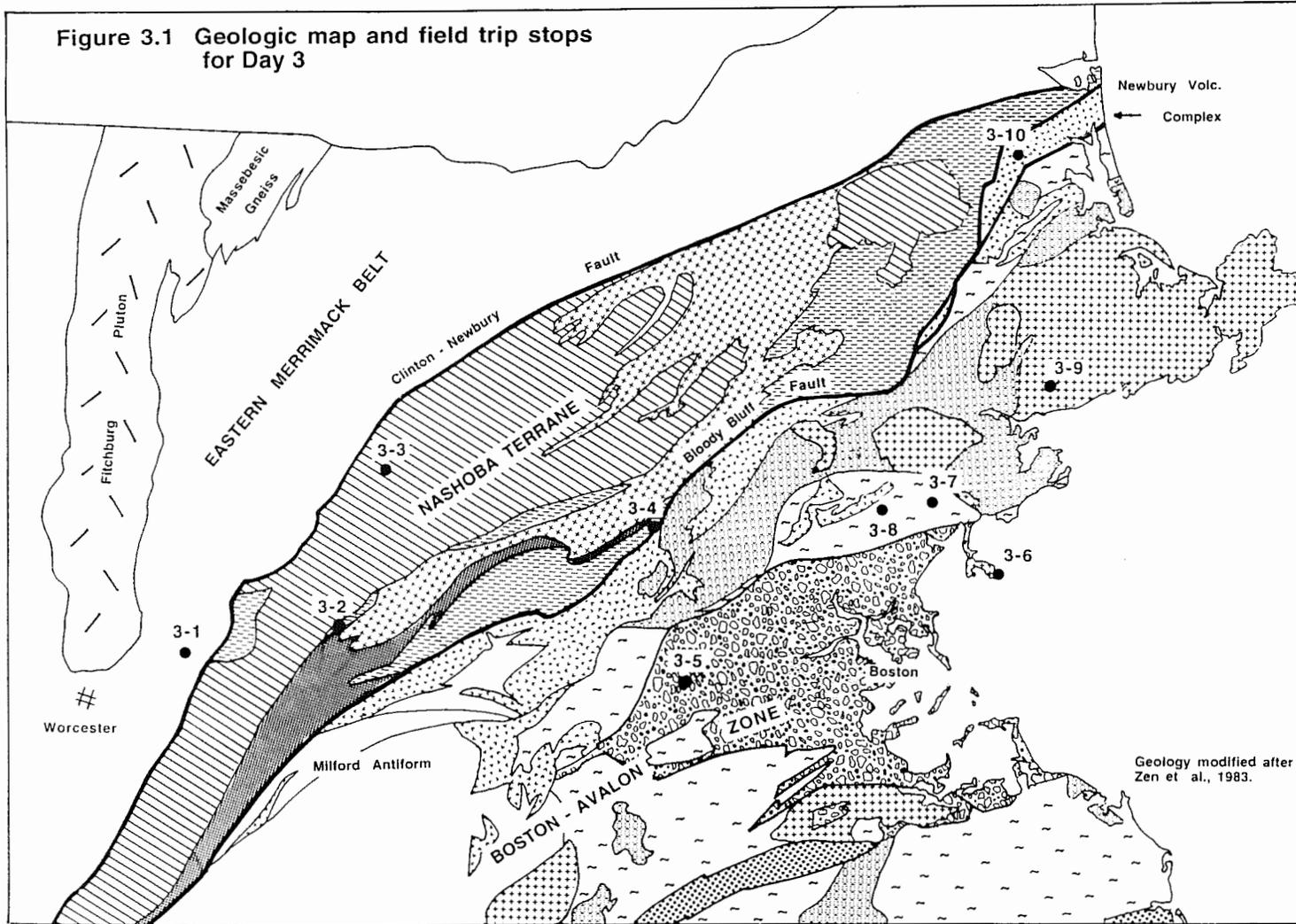
#### INTRODUCTION

The eastern side of the Appalachian Orogenic belt in Massachusetts is formed by three fault-bounded terranes or potential terranes (Figure 3.1), exotic to the North American margin prior to the mid-Paleozoic. The easternmost of these, the Boston-Avalon Zone (shown as the Milford-Dedham Terrane on Fig. 0.1 and by Zen et al., 1983), was part of the Avalon "plate" and can be readily correlated with the type Avalon in eastern Newfoundland and other Avalonian fragments in eastern North America. The Nashoba Terrane, separated from the Boston-Avalon Zone by the Bloody Bluff Fault System, formed largely as part of an early Paleozoic (Cambro-Ordovician to Sil.?) arc and its associated sedimentary basin. While it is tempting to include the Nashoba Terrane as part of the Avalon Zone, we have no direct evidence of basement ages upon which to base this assignment. The Nashoba Terrane was intruded both by calc-alkaline intermediate plutons and peraluminous granites during the Late Ord. and Sil. In the Silurian (c. 425 Ma, see Hepburn et al., this Vol., Chapt. X) this terrane experienced deformation and upper amphibolite facies metamorphism, which signaled the beginning of final closure of the ocean between these eastern terranes and the already combined Laurentia-Craton X. Thus, the Acadian Orogeny (if that name is still appropriate for events at 425 Ma) started in the Silurian on the eastern side of the belt and telescoped westward during the Devonian.

The eastern part of the Merrimack Belt in Massachusetts (Zen et al., 1983), located between the western edge of the Nashoba Terrane and, approximately, the Fitchburg Pluton (Fig. 3.1), is enigmatic. The age of the sedimentary rocks within this belt, as well as the age of the deformation, are controversial. Whether it is part of the rest of the Merrimack Belt (Robinson and Goldsmith, 1991), a lower grade faulted portion of the Nashoba Terrane (Hepburn et al., 1987a) or a separate terrane (Zen, 1989; Lyons et al., 1982) remains to be determined.

Because of page constraints, descriptions of the eastern terranes must be brief. Since Boston is the "hub" for the meeting, several other field trips will be visiting these terranes more extensively than we are able to do on our excursion across the whole of the orogenic belt. Thus, participants should refer to other articles in the guidebooks for this meeting for more details on these terranes. Of particular interest will be the field excursions of Hepburn et al., Chapter X; Hon et al., Chapter Q; Rast et al., Chapter S; Bailey and Ross, Chapter Y; and Mosher et al., Chapter BB. A thorough account of the geology of eastern Massachusetts with extensive references is found in "The Bedrock Geology of Massachusetts": U.S. Geol. Survey Prof. Paper 1366, Chapters E-J, edited by Hatch (1991), which is a companion to the Mass. Geologic Map. Other recent field guides to eastern Massachusetts include those

Figure 3.1 Geologic map and field trip stops for Day 3



**NASHOBA TERRANE**

-  Andover Granite
-  Intermediate & Granite Plutons
-  Marlboro Formation
-  Schists, Gneisses  
(Nashoba, Fishbrook, Shawsheen Fms)

**BOSTON - AVALON ZONE**

-  Carb. Seds. (Norfolk, Narr. Basins)
-  O-Dev. Peralkalic Gran. & Volcs.
-  PeZ & E. Pz Sed. Fms.  
(Boston Bay, Braintree, Weymouth)

-  PeZ Granite & Volcanics
-  PeZ - Pz Mafic Plutons
-  Pre-630 Ma Mafic Volcs. & Sed.

Geology modified after  
Zen et al., 1983.

of the New England Intercollegiate Geological Conferences for 1964, 1981, 1984, and 1986; also Zen's (1989) trip for the 28th Internat. Geol. Congress followed a route across Massachusetts similar to ours.

### EASTERN MERRIMACK BELT

The Merrimack Belt in Massachusetts east of roughly the Fitchburg Pluton (Fig. 3.1) is underlain by a sequence of calcareous metasiltsstones (Stop 3-1), pelites and impure quartzites (Zen et al., 1983; Robinson and Goldsmith, 1991). These are metamorphosed to the lower greenschist facies along the eastern side of the belt, but increase in grade toward the west. On the Massachusetts Geological Map (Zen et al., 1983) they are shown as Silurian or Ordovician, based on potential correlation with fossiliferous strata in Maine (Robinson and Goldsmith, 1991). If this assignment is correct, then the deformation and metamorphism of this belt is Acadian and similar to that of the rest of the Merrimack Belt. However, Lyons et al. (1982) pointed out the distinct geological characteristics of this eastern belt and resurrected the name Merrimack Trough for it. If dates on cross-cutting plutons (455 Ma, Newburyport Quartz Diorite, Zartman and Naylor, 1984; 473 Ma, Exeter Diorite (NH), Gaudette et al., 1984; Hon et al., 1986) are correct, then the sediments must be pre-Middle Ordovician. Bothner et al. (1984) and Hon et al. (1986) note that the Exeter Diorite is undeformed but cuts deformed and metamorphosed rocks and has a contact aureole developed adjacent to it, indicating that the metamorphism and deformation in this belt must also be pre-Middle Ordovician. The Massabesec Gneiss Complex (Fig. 3.1) is associated with the Merrimack Trough and contains two ages of orthogneiss dated at 650 and 475 Ma (Besancon et al., 1977; Aleinikoff et al., 1979; Kelly et al., 1980). Bothner et al. (1984) suggest that paragneiss cut by the older of these dated orthogneisses is a high-grade equivalent of sediments in the Merrimack Trough. If this is true, then the sedimentary rocks are Precambrian. If this belt of rocks is in fact a separate terrane, as this evidence suggests, its boundary with the high-grade Siluro-Devonian rocks of the Merrimack Belt to the west is not clearly differentiated. Zen (1989) suggests that such a boundary may be hidden beneath the Devonian Fitchburg Pluton.

### NASHOBA TERRANE

The high-grade Nashoba Terrane is everywhere fault-bounded (Clinton-Newbury Fault Zone on the NW, Bloody Bluff Fault Zone on the SE, Fig. 3.1) and clearly differs from adjoining terranes. Fig. 3.2 summarizes the major events in the history of this terrane. Its stratigraphy is dominated by two units, the Marlboro Fm., largely mafic volcanic rocks, and the Nashoba Fm. (Stop 3-3), volcanogenic sedimentary rocks, pelites and minor carbonates of Early Paleozoic age (Ordovician, but also may include some Cambrian or Early Silurian). These rocks were metamorphosed and polydeformed during the Silurian (see Hepburn et al., this Vol., Chapt. X) under sillimanite and sillimanite-K-feldspar zone conditions to amphibolites, biotite-feldspar gneisses, schists and calc-silicate granulites (Bell and Alvord, 1976). Migmatites are developed in rocks of the appropriate composition toward the northeast. Geochemistry indicates the Marlboro volcanic rocks are generally high-alumina basalts and have trace element signatures compatible with an arc or marginal basin tectonic setting (DiNitto et al., 1984).

The oldest rock yet dated in the terrane is the Fish Brook Gneiss, a leucocratic feldspar-biotite gneiss once thought to represent possible Late Proterozoic basement (Olszewski, 1980). This rock has been recently redated by Dunning (see Hepburn, this Vol., Chapt. X); he obtained a U-Pb zircon age of  $520 \pm 14$ -11 Ma for the time of the original igneous crystallization. While no Precambrian rocks have been found, detrital zircons (Olszewski, 1980) and Nd isotopes in the Andover Granite (Hill et al., 1984) indicate the presence of Early Proterozoic crustal material in the terrane, at least as detritus.

The Nashoba Terrane experienced widespread plutonism from the mid-Ordovician through at least the Silurian, with intrusions of both calc-alkaline intermediate and approximately contemporaneous peraluminous granitic magmas (Zartman and Naylor, 1984; Hepburn et al., 1987a). The intermediate composition plutons, typified by the Straw Hollow Diorite (Stop 3-2) and Sharpners Pond Diorite, are little deformed and range in composition from gabbroic cumulates to hornblende and biotite diorites and tonalites with sphene as a nearly ubiquitous accessory phase (Hill et al., 1984; Hon et al., 1986). The granites vary from foliated biotite-muscovite granite to unfoliated garnet-bearing muscovite granite and pegmatite. They likely were intruded over at least a 50 Ma period (Zartman and Naylor, 1984; Hill et al., 1984). The Andover Granite (Fig. 3.1; Stop 3-2) is thought to have been at least partially generated by anatexis of the Nashoba Fm. during the Silurian metamorphism. A recent U-Pb date on the unfoliated, younger phase of the Andover of  $412 \pm 2$  Ma by Dunning (Hepburn et al., this Vol., Chapt X) supports this conclusion. Magmatic pillows and other features (Hon et al., 1986; Hon et al., this Vol., Chapt Q) indicate the co-existence of the granitic and intermediate magmas. Geochemistry indicates that these two magma types are not

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co-genetic. It is likely that heat from the calc-alkaline plutons may have induced melting in the metasedimentary rocks that led to the granite formation.

Metamorphism in the Nashoba Terrane is generally of the lower pressure-higher temperature andalusite-sillimanite type, although early kyanite pseudomorphs have been found in a few locations (Stop 3-2). A  $425 \pm 3$  Ma age on monazite from the Fish Brook Gneiss (see Hepburn et al., this Vol., Chapt. X) is interpreted to represent the time of metamorphism and is supported by dates from plutons.  $^{40}\text{Ar} / ^{39}\text{Ar}$  ages on hornblendes from amphibolites of the Nashoba Terrane give a range of 354-325 Ma; an Ar plateau age on biotite gives 308 Ma (Hepburn et al., 1987b). These indicate the time of uplift and cooling of the terrane. They also indicate that the Nashoba Terrane in Massachusetts did not experience an Alleghanian thermal event sufficient to affect the Ar/Ar systematics of hornblende or biotite.

**BOSTON - AVALON ZONE**

The Boston-Avalon Zone is likely a composite terrane. It contains many features similar to the type Avalon of eastern Newfoundland but has been eroded to a somewhat deeper level. Indeed, plutonic rocks are areally the most important at current levels of erosion (Zen et al., 1983). The bulk of these are related to the main Avalonian magmatic pulse of Late Proterozoic (c. 600-630 Ma; Zartman and Naylor, 1984) arc magmatism. These plutons include the calc-alkaline Dedham (Stop 3-8), Milford, Westwood and associated granites (Fig. 3.3). The Lynn (Stop 3-7) and Mattapan Volcanics are extrusive equivalents (Smith and Hon, 1984). A poorly preserved record of

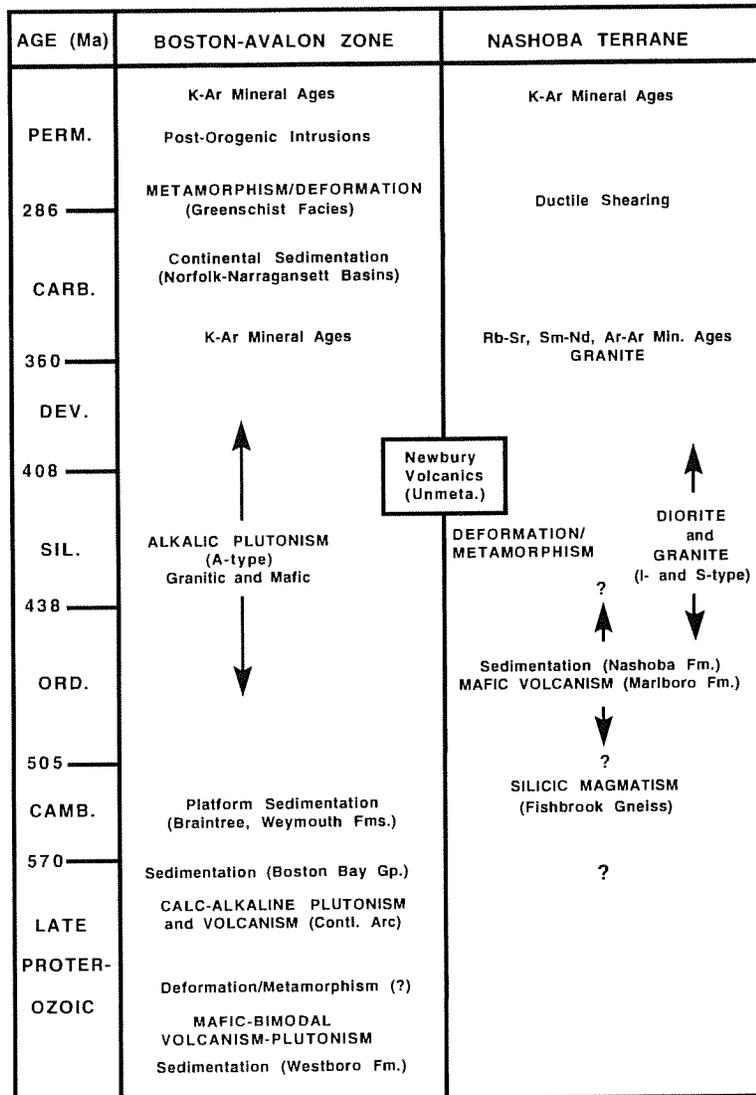


Figure 3.2 Comparison of geologic events in the Boston-Avalon and Nashoba terranes, eastern Massachusetts.

quartzitic sedimentation (Westboro Quartzite) and mafic volcanism (Middlesex Fells Volcanics) that predate this magmatism is preserved as blocks and pendants in the granitic plutons. Bailey et al. (1987) suggest that the quartzites formed as olistostromal deposits during a period of rifting and collapse of a cratonal margin. Geochemistry indicates the alkalic nature of the Middlesex Fells Volcanics (Cardoza et al., 1990), which supports this idea.

The Boston Basin developed as an intra-arc rift basin (Bailey, 1984; Nance, 1990; Socci and Smith, 1990) during the waning stages of the arc magmatism. Sedimentary rocks in the Boston Bay Group are largely conglomerates (Roxbury Conglomerate; Stop 3-5) and shales (Cambridge Argillite) dated on the basis of acritarchs as Late Proterozoic (Lenk et al., 1982). Whether the Boston Bay Group sediments, including the Squantum Tillite/Tilloid are of glacial, fluvial or submarine mass-wasting origin continues to be debated (Rehmer and Roy, 1976; Socci and Smith, 1990; Newman et al., this Vol., Chapt. U). Early and Middle Cambrian platformal sedimentation (Stop 3-6) followed the deposition of the Boston Bay Group with at most a minor break. An Acado-Baltic trilobite fauna in these rocks allow the correlation of the Boston area rocks with other fragments of Avalonia (Rast and Skehan, 1983). No Paleozoic sedimentary rocks younger than the Cambrian are found in the Boston-Avalon Zone of Massachusetts (excluding those associated with the Sil.-Dev. Newbury Volcanics; see below) except for Pennsylvanian nonmarine fluvial deposits in the Norfolk and Narragansett basins. From Ordovician to Devonian, perhaps in distinct pulses (Hermes and Zartman, 1985), the area experienced sporadic but rather extensive A-type peralkaline granitic (Stop 3-9) and mafic plutonism with local felsic volcanism.

### Metamorphism and Deformation

In the Boston area, the Boston-Avalon Zone has been metamorphosed to no higher than lower greenschist facies, with brittle faulting the most common type of deformation (Zen et al., 1983). Most of the metamorphism is thought to be Alleghanian in age. Alleghanian metamorphism increases to the amphibolite facies to the south and west in the Milford Antiform (Fig. 3.1) and in southern Rhode Island where the Narragansett basin is polydeformed and metamorphosed to the amphibolite facies (see discussions by Mosher et al., Chapt. BB and Wintsch et al., Chapt. H of this Vol.). It is likely that the Boston-Avalon Zone in Massachusetts also experienced a Precambrian deformation and a low-grade metamorphism that precede the main Avalonian magmatic pulse. Folded metasedimentary xenoliths are found in the Dedham Granite north of Boston. A more complete record of Precambrian deformation and greenschist facies metamorphism is preserved in Rhode Island (Rast and Skehan, 1983; Skehan and Rast, 1983)

### NEWBURY VOLCANIC COMPLEX

The Newbury Volcanic Complex (Stop 3-10) occurs only in fault-bounded slivers (Fig. 3.1) directly between the Boston-Avalon and Nashoba Terranes. Whether it belongs to either of these terranes has yet to be established (Zen, 1989). It is composed of a bimodal sequence of andesitic and rhyolitic volcanics and shallow intrusives with

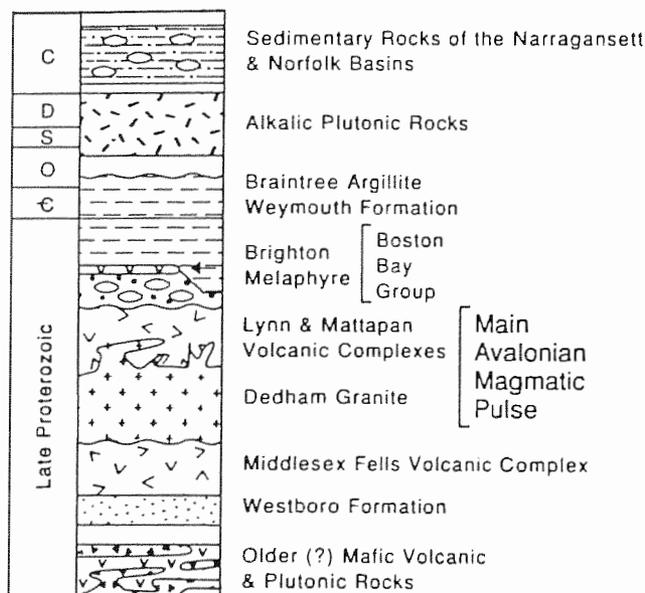


Figure 3.3 Simplified stratigraphic column for the Boston-Avalon Zone in the Boston area, Massachusetts.

interbedded sedimentary rocks that contain Late Silurian to Early Devonian fossils (Shride, 1976a). These rocks are tilted but not deformed or metamorphosed. McKenna et al. (1993) indicate that these rocks are calc-alkaline and formed as part of a continental arc. Hon et al. (1986) note the similarity of the composition of the Newbury to intrusive diorites and granites of the Nashoba Terrane and suggest that the Newbury may be a near surface volcanic expression of them, preserved in a down-dropped fault block. Volcanic rocks similar in both composition and age also occur in the coastal volcanic belt in eastern Maine (Gates and Moench, 1981)

### TECTONIC DISCUSSION

Differences in the geologic histories between the Nashoba Terrane and the Boston-Avalon Zone are shown in Fig. 3.2. The Boston-Avalon Zone was a continental arc during the Late Proterozoic, but by Early Paleozoic arc magmatism ceased and the area developed into a stable shelf or platform. The Nashoba Terrane is interpreted to have formed in an arc or marginal basin setting in the Lower Paleozoic. It was intruded by both calc-alkaline and granitic plutons in the Ordovician-Silurian and was extensively deformed and metamorphosed in the Silurian. The calc-alkaline plutonism is believed to be related to an Ordovician-Silurian east-dipping subduction zone beneath the Nashoba Terrane and Eastern Merrimack Belt. The Newbury Volcanics may be preserved remnants of this magmatism. As the Nashoba impinged on the Merrimack Belt to the west, perhaps obliquely, it was metamorphosed and deformed. This initiated the Acadian orogenic cycle on the eastern side of the orogenic belt. Tectonism spread westward as the combined Avalon-Nashoba block and the Laurentia-Craton X block collided. The Ordovician-Devonian peralkaline magmatism of the Boston-Avalon Zone likely represents behind-the-arc magmas originating from the same east-dipping subduction zone and interacting with mature crust in Avalonia (see Hon et al., this vol., Chapt. Q). If this is so, the Nashoba Terrane developed and was along the leading edge of Avalon by at least the Ordovician. The large faults between and within these terranes have clearly altered their original relationships, with foreshortening and an unknown amount of oblique movement. These faults have had a long and complex history of motion. Faulting in the area ranges in age from Precambrian (Skehan and Rast, 1991; Rast et al., Chapt. S, this Vol.) to Alleghanian (e.g. Goldstein, 1992) and even Mesozoic (Zen et al., 1983).

### ROAD LOG: DAY 3

**Start:** From Worcester, take I-290 east to exit #23N, Rt. 140, Shrewsbury. Starting mileage (0.0) is on Rt. 140 N at bridge over I-290. Proceed north on Rt. 140. 2.0 junction of Rts. 140 & 70, continue NW on Rt. 140 for 0.3 (2.3) to outcrops on both sides of the road. Park on right shoulder.

**STOP 3-1. OAKDALE FM., EASTERN MERRIMACK BELT** (Boylston, MA; 20 minutes). The Oakdale Fm. here is typical of the lower greenschist facies calcareous metasiltstones, impure quartzites and pelitic rocks of the eastern Merrimack Belt (Zen et al., 1983) just west of the Clinton-Newbury Fault Zone. Based on ages of cross-cutting plutons, these rocks are Ordovician or older. At this stop the Oakdale Fm. consists of light gray to purplish tan-weathering calcareous metasiltstone with ankerite weathering spots and interbedded gray phyllites. Tight folds with axial surfaces approximately parallel to bedding indicate recumbent folding.

Continue NW on Rt. 140 to end of Wachusett Reservoir, (2.6), turn around and retrace route to junction Rt. 140 & I-290. 5.3 Enter I-290 East and proceed to interchange of I-495, I-290 and "To 85" (13.0) noting outcrops of Nashoba Fm. along the way. Continue east on "To 85" to just beyond (east of) the point where the ramp from I-495 North joins it (13.5). Park with care on the right shoulder just past the electrical box. Walk to exposures along the north side of the ramp from "To 85" to I-495 West, the center ramp in the interchange.

**STOP 3-2. STRAW HOLLOW DIORITE & ANDOVER GRANITE, NASHOBA TERRANE** (Marlborough, MA; 40 minutes). Examples of the two principal types of plutonic rocks in the Nashoba Terrane are seen here. The Straw Hollow Diorite is typical of a number of the intermediate composition calc-alkaline dioritic to tonalitic bodies in the terrane. The Straw Hollow contains at least two phases: a finer grained, more foliated phase and a coarser grained, lighter colored, less foliated rock that intrudes it. The Straw Hollow Diorite is assumed to be similar in age to the Sharpners Pond Diorite to the northeast, dated at 430 Ma (Zartman and Naylor, 1984). At the east end of this outcrop the diorite intrudes pelitic schist of the Nashoba Fm. containing kyanite pseudomorphs that have been replaced by sillimanite. (Please do not hammer on these.) Contemporaneous with the intermediate composition plutons, the Nashoba Terrane was intruded by a series of foliated to unfoliated, metaluminous to peraluminous granites and pegmatites. The Andover Granite seen here is characteristic, although somewhat sheared. If time permits we will briefly examine blastomylonite associated with the Assebet River Fault Zone (one of several within the Nashoba Terrane) along the south side of "To 85" near where it is joined by the ramp from I-495 North.

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Continue east on "To 85" to Fitchburg St. (14.5). Turn right and in 0.2 miles turn around by the High School and return to the traffic light at "To 85." 14.9 Turn west on "To 85" and return to I-495 cloverleaf. 15.8 Take I-495 North. Note excellent exposures of biotite schists and gneisses of the Nashoba Fm. along I-495 between here and the next stop. 26.2 Exit I-495 at junction with Rt. 111 West (exit #28, Boxborough). 26.4 Turn right onto Rt. 111 west and cross over I-495. 26.7 Park on right (north) at large exposures opposite the entrance to I-495 South.

**STOP 3-3. NASHOBA FORMATION, NASHOBA TERRANE** (Harvard, MA; 30 minutes). The Nashoba Fm. is typified by various biotite and feldspar gneisses, schists and lesser amounts of calc-silicate granulite. These have been metamorphosed to sillimanite and sillimanite-K-feldspar zone conditions. In general, the amount of migmatite increases to the northeast until the Nashoba is entirely cut out by the Andover Granite. It is believed that melting in the Nashoba Fm. contributed to the formation of this granite.

This outcrop is typical of the migmatized Nashoba: biotite gneisses are interlayered with migmatitic gneisses and pegmatites. Sillimanite is commonly present with biotite in selvages along the rims of melted material. Muscovite is present in large, retrograded? flakes. It is believed most of the pegmatite and granite here is more or less locally generated by anatexis of layers with the appropriate composition (Hepburn and Munn, 1984). Note how the percentage of melting changes with the composition of the original layer.

Proceed west on Rt. 111 for 0.1 to Swanson Rd., turn around and return to junction with I-495 North. 27.1 Enter I-495 North and proceed to junction with Rt. 2 East. 30.0 Take exit #29A to Rt. 2 and follow it east. 37.5 Concord rotary, continue east on Rt. 2. 41.0 Woods by Walden Pond, where Thoreau lived in the wilderness. 42.2 Follow Rt. 2A straight ahead where Rt. 2 turns right. 42.8 Bear right on Rt. 2A. We are following approximately the route of Paul Revere's ride. 45.4 Turn left onto Mass. Ave., opposite Minuteman Tech. 45.6 Stop on right shoulder and walk to light-colored outcrops on the north side of the road.

**STOP 3-4. BLOODY BLUFF FAULT ZONE, NEAR BOUNDARY BETWEEN BOSTON-AVALON AND NASHOBA TERRANES** (Minute Man Natl. Hist. Park, Lexington, MA; 30 minutes). The Bloody Bluff Fault Zone separates the Nashoba Terrane on the west from the Boston-Avalon Zone on the east and undoubtedly has had a long and complex history. Here the zone is on the order of a kilometer wide. Nelson (1987) described these exposures, the type area of this fault zone, in detail. Mylonitized granite (Dedham Granite? of the Boston-Avalon Terrane) and sheared mafic rocks along the north side of the road have a well-developed NW-dipping mylonitic foliation.

Continue on Mass. Ave. 0.1. 45.8 Turn left onto Wood St. (do not cross I-95). Proceed 0.05 and turn left onto Old Mass. Ave. (45.85). Follow for 0.5 mi. to junction with Rt. 2A (46.4). The rusty weathering exposure to the right of the road junction is Bloody Bluff itself, where the colonists skirmished with the British retreating from Concord on April 19, 1775 at the beginning of the Revolutionary War. The rock is an altered and shattered granite. We will stop if there is time. Turn left on Rt. 2A and follow to I-95. 46.7 Turn right and enter I-95 South. Follow I-95 to exit #20A (56.8) and exit onto Rt. 9 East. 59.8 Turn right on Florence St., just before Atrium Mall. 60.4 Turn left at light onto Hammond Pond Parkway. 60.6 Turn right into Chestnut Hill Mall. Continue along the left side of the Mall by the pond to exposures at the far end, behind the Star Market. Massive conglomerate in the Roxbury Fm. of the Boston Bay Group. We will observe this from the bus. Return to Hammond Pond Parkway (61.2). Turn right. 62.1 Pull off on right shoulder with care just before road junct. (or park in pull-off on left). Cross the road and walk the short path on the west into Newton Webster Conservation Area to outcrops. Beware of Poison Ivy!

**STOP 3-5. ROXBURY CONGLOMERATE, BROOKLINE MEMBER, BOSTON BAY GROUP, BOSTON-AVALON ZONE** (Newton, MA; 40 minutes). The Roxbury Conglomerate, one of whose three members is the infamous Squantum Tillite or Tilloid, is the most visible of the two major sedimentary units of the Late Proterozoic Boston Bay Group; the uncommonly exposed Cambridge Argillite is the other. Various models for the depositional environment for these facies have been advanced over the years and are still debated (Socci and Smith, 1990). Rehmer and Roy (1976) provide an excellent short summary of the problem (as well as a detailed description of this particular exposure). They discuss the three most common sedimentation models for the Boston Bay Group: (1) terrestrial glacio-alluvial-lacustrine deposition (Remember that Avalonia was supposedly near the south pole in the Proterozoic); (2) marine mass movement in a eugeosynclinal setting (Dott, 1961); and (3) subaqueous debris flow-fan deposition.

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Much of the Roxbury Conglomerate is a massive, coarse conglomerate such as that we just drove by in the shopping mall. Here, however, sedimentological features can be distinguished. Four intervals of roundstone pebble/cobble conglomerate alternate with intervals of sandstone and shale. Large channels with erosional relief of up to several meters can be seen truncating sandstone beds (Rehmer and Roy, 1976). The conglomerate in the Roxbury is generally clast supported with the majority of the pebbles recognizable as felsites, volcanics and granites from the surrounding area. Note in some of the sandstones the "outsized" pebbles with overlying draped strata. These have been interpreted as glacial "dropstones" by the proponents of the glacial model. (Do you agree?)

The upper part of an altered basaltic lava flow, with overlying red-brown volcanogenic sandstone, is also present here. These Brighton Volcanics are found at several horizons within the Boston Bay Group. Cardoza et al. (1990) indicate that these volcanics have a calc-alkaline geochemical signature and are thus likely related to the waning stages of Avalonian arc magmatism.

Continue on Hammond Pond Parkway for 0.1. **62.2** Turn left (west) on Beacon St. and follow to Newton Center. **64.0** Turn right onto Centre St. Follow across Commonwealth Ave. 2.0 to T-junction by hotel. **66.0** Turn right and follow to entrance of Massachusetts Turnpike East. **66.2** Enter Mass. Turnpike east and follow to its end. **72.8** Follow signs to I-93 and Rt. 3 North. **73.0** Follow to Expressway North and Callahan Tunnel. **74.2** Take exit #24, Callahan Tunnel. Proceed through tunnel and continue on Rt. 1A north past the airport. **78.9** Veer right, stay on Rt. 1A north. **79.4** Rotary, stay on Rt. 1A north. **83.8** Bear right on Carroll Parkway, follow signs to Nahant and Swampscott. **84.5** At the rotary take the first right to Nahant; cross causeway (a tombolo). **86.0** Stay straight at the end of the causeway on Nahant Rd.; continue through the town of Nahant. **87.9** Bear left toward East Point. **88.2** Turn left into Northeastern Univ. Marine Science Center and park (permission needed). Walk to outcrops at East Point.

**STOP 3-6. LOWER CAMBRIAN WEYMOUTH FM., EAST POINT, NAHANT** (1 hour) In addition to a 600-650 Ma plutonic-volcanic episode, the most important component of Avalonian terranes in eastern North America is that they contain a Cambrian sequence with an Acado-Baltic trilobite fauna. The sea cliffs at this locality provide the best section through the Cambrian succession in the Boston area (with fossils, but no trilobites). At this locality the sequence has been intruded by the Early Paleozoic Nahant Gabbro and a multitude of dikes and sills. R. Bailey and M. Ross have worked extensively on these exposures and are leading a separate GSA trip just to this location. Please refer to the excellent map, descriptions and references for this locality in Bailey and Ross (Chapt. Y, this Vol.). Obviously we could spend days at these exposures. We will concentrate on the Cambrian sediments in the limited time that we have.

Return to the mainland by reversing the previous route. **91.8** Rotary. Return south on the the Lynnway, the second road as you go around the rotary, retracing our previous route. **92.4** Turn right on Market St. and follow signs to downtown Lynn. **92.8** Turn left and immediately veer right onto Franklin St. **93.4** Straight on North Franklin St. **93.6** Take sharp left onto Linwood Rd. **94.0** Turn right onto Walnut St. **94.4** Stay right on Walnut St. **94.6** Bear left on Walnut St. **94.7** Park on the north side of the road by the ball fields. Walk straight across the fields to the spillway of Breeds Pond. Examine exposures to the left of the spillway and in a small abandoned quarry about 750 feet northwest along the shore.

**STOP 3-7. LYNN VOLCANIC COMPLEX** (Lynn, MA; 45 minutes). The Lynn Volcanics formed as part of an intrusive-volcanic complex with the Dedham Granite north of Boston during the major period of Late Proterozoic magmatic arc activity. Dacites and rhyolites dominate within the crystal and lithic tuffs, breccias and subvolcanic rocks of the Lynn (Smith and Hon, 1984). The Lynn has been recently dated as  $596 \pm 3$  Ma by Dunning (Hepburn et al., Chapt. X, this Vol.) on zircons taken not far from here. In the abandoned quarry four rock types are exposed: welded tuffs, crystal tuffs, breccia and an intrusive dome facies (Smith and Hon, 1984). Plagioclase and hornblende form the phenocrysts here and in exposures of rhyolite next to the spillway.

Continue west (Walnut St.) **95.3** Turn left on Washington St. **95.4** Bear right onto W. Sigourney St. **95.5** Turn left on Garfield Ave. **95.6** Turn right on Sherman St. **95.7** Turn left at T-junction onto Fairmount Ave. **95.85** Turn right on Oakridge Dr. into Forest Highlands development. **95.9** Turn right onto Foliage Dr. **96.0** Turn right on Stonecrest Dr.; proceed to the end (**96.1**). Outcrop at cul-de-sac.

**STOP 3-7A. LYNN VOLCANIC COMPLEX; BRECCIA** (Saugus, MA; optional, 15 minutes). This volcanic breccia in the Lynn contains abundant clasts of granite and other rocks that pre-date the magmatic pulse, such as quartzites of the Westboro Fm. and mafic volcanics of the Middlesex Fells.

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Return to Fairmount Ave. **96.3** Turn left onto Fairmount Ave. **96.9** Turn left onto Walnut St. and follow to Rt. 1. **98.2** Turn right and enter Rt. 1 South. **99.7** Outcrop of Dedham Granite on right. **100.2** Turn into K-Mart/Stop and Shop Plaza just beyond Main St. Proceed to exposures beside and behind Super Stop & Shop store.

**STOP 3-8. DEDHAM NORTH GRANODIORITE** (Saugus, MA; 20 minutes). The Dedham Granite-Granodiorite is the most widespread rock of the Boston-Avalon Zone (Zen et al., 1983), reaching truly batholithic proportions. While varying from true granite to diorite, the Dedham exposed here is a medium- to coarse-grained calc-alkaline granodiorite (normative diopside up to 8%) containing hornblende, biotite, plagioclase, quartz and K-feldspar (Smith and Hon, 1984; Wones and Goldsmith, 1991). The Dedham North Granodiorite (i.e., north of Boston) was recently dated at  $607 \pm 4$  Ma (Hepburn et al., Chapt. X, this Vol.) from exposures along Rt. 1 just north of here. Wones and Goldsmith (1991) and Zartman and Marvin (1991) summarize other dates on the Dedham.

Proceed to the north exit from Plaza, to Main St. **100.4** Turn right onto Main St. **100.6** Turn right, enter Rt. 1 North and continue to Rt. 128. **105.2** Exit onto Rt. 128 North. Continue on Rt. 128 North past I-95 interchange. Note outcrops of Peabody Granite. **113.6** Exit at Brimball Ave. Beverly (exit #19). **113.8** Turn left. **113.9** Turn right onto Brimball Ave. **114.8** Stay straight toward Rt. 22 South. **114.9** Turn right onto Rt. 22, Essex St. **115.0** Turn left onto Corning St. **115.5** Turn right onto Victor Ave. **115.7** Turn left then immediately right onto Cedar St. **115.8** Park on the right by the entrance to a field (or in the field if the wire is not up) that is a partially filled-in quarry. Walk to exposures along the side and rear of the old quarry.

**STOP 3-9. CAPE ANN GRANITE** (Beverly, MA; 40 minutes). The Cape Ann Plutonic Complex is the largest and best known of the early to mid-Paleozoic peralkaline granitic (A-type) intrusions into the Boston-Avalon Zone. The Cape Ann Complex is Late Ordovician to Early Silurian (Zartman and Marvin, 1991; Wones and Goldsmith, 1991). The main phase of the Cape Ann Granite consists of alkalic granite to quartz syenite (Wones and Goldsmith, 1991). In this quarry porphyritic microgranite is exposed in addition to medium-grained Cape Ann granite or syenite. Interesting mafic dikes cutting the granite have been broken and pulled apart, yet the space between them is filled with the very granite they cut. Thus, they likely intruded shortly after the emplacement of the granite when the granite was solid enough to crack but before it was completely solidified (Toulmin, 1964 a,b).

Continue on Cedar St. 0.1. **115.9** Turn right on Essex St. (Rt. 22). **116.6** Turn left onto Brimball Ave. and follow to Rt. 128. **117.8** Enter Rt. 128 South and follow to first exit. **118.4** Exit onto Rt. 1A North, exit #20A. **118.6** Turn right at end of ramp onto Rt. 1A North. In 0.1 (**118.7**) Turn left onto Conant St. **118.9** Turn right onto Rt. 97 and follow to Rt. 1. **124.2** Turn right onto Rt. 1 North. **131.8** Turn right at the light onto Central St. and park as soon as possible. Examine outcrops on both sides of Central St. by the junction with Rt. 1.

**STOP 3-10. NEWBURY VOLCANIC COMPLEX, PORPHYRITIC ANDESITE.** (Rowley, MA; optional, 20 minutes). The Newbury Volcanic Complex consists of a series of andesitic and rhyolitic volcanic rocks and associated sediments that lie entirely within fault slivers along the Bloody Bluff fault zone, directly between the Boston-Avalon and Nashoba Terranes. It is well dated as latest Silurian to possibly Early Devonian on the basis of shelly fossils found in associated sediments (Shride, 1976b). Exposures here are tuffs and flows of the porphyritic andesite member (Shride, 1976a, 1976b). The top of each flow is recognizable by the presence of a vesicular band. Note differences in the phenocryst content between the flows.

Tectonically the Newbury, while tilted, is important because it demonstrates that here the Acadian and subsequent orogenies created no penetrative deformation and metamorphism no higher than the lowest greenschist facies. Note the undeformed nature of the plagioclase phenocrysts and filled amygdules in the outcrop. Contrast this to the Silurian deformation seen earlier today in the adjacent Nashoba Terrane; and also to the Acadian-deformed Silurian rocks to the west that we saw on Day 2. Geochemically the Newbury Volcanic Complex consists of a bimodal volcanic suite of andesites and rhyolites (Shride, 1976a; McKenna et al., 1993). The andesites are high-alumina calc-alkaline rocks with trace element signatures indicative of formation on a continental margin above a subduction zone. Thus, these rocks give evidence of a subduction zone adjacent to Avalon in the late Silurian-early Devonian. Gates and Moench (1981) indicate that the Newbury Volcanic Complex was once likely continuous with similar aged volcanics in the Coastal Volcanic Belt of eastern Maine.

**END OF TRIP.** Turn around. Head south on Rt. 1. **134.4** Turn right (west) onto Rt. 133 and follow to I-95. Follow I-95 south to I-93 and Boston.

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## ROBINSON, RATCLIFFE AND HEPBURN

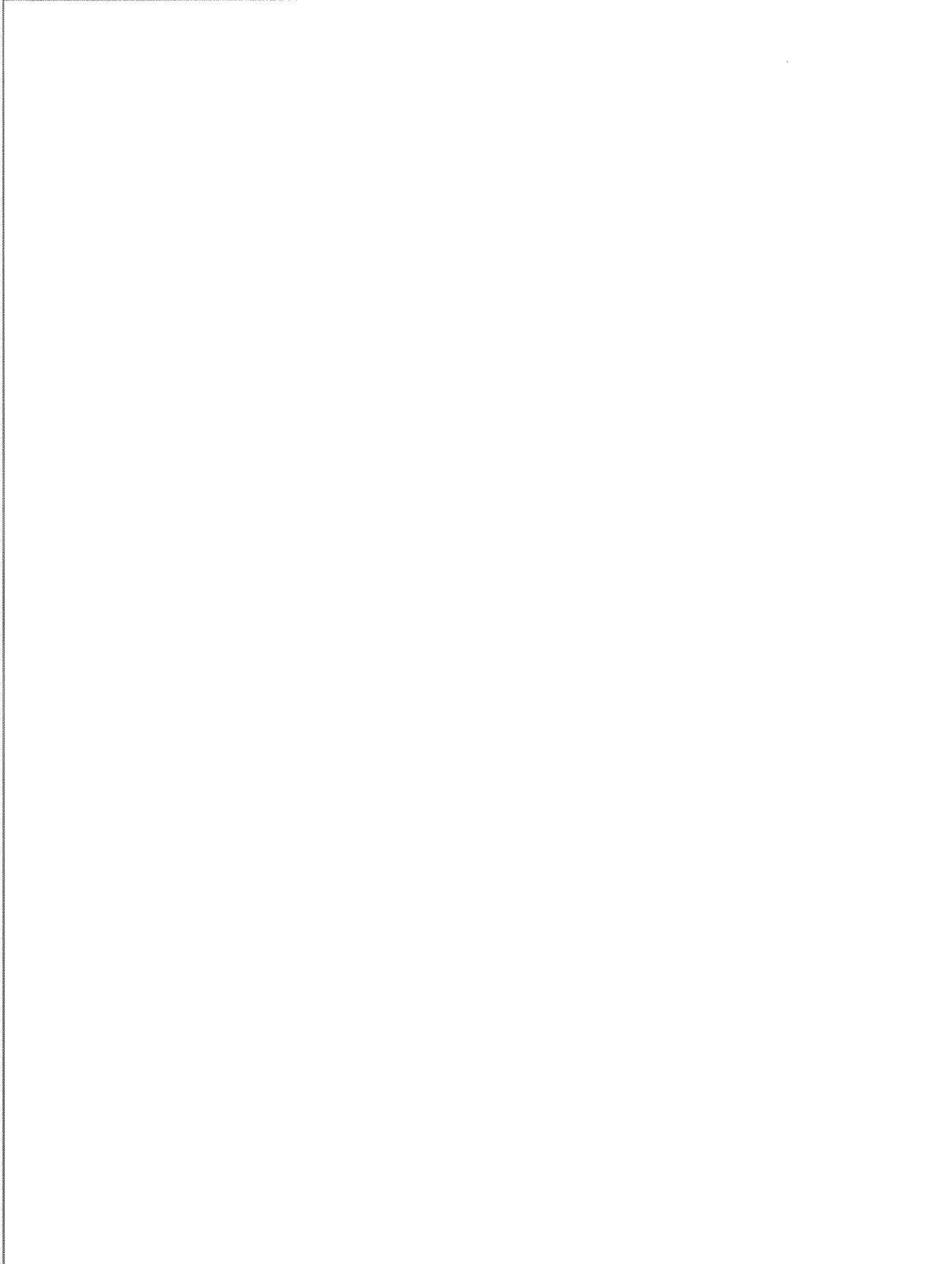
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Chapter D

# Petrologic and Isotopic Studies in Metamorphic Rocks of Eastern Vermont and Western New Hampshire

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## PETROLOGIC AND ISOTOPIC STUDIES IN METAMORPHIC ROCKS OF EASTERN VERMONT AND WESTERN NEW HAMPSHIRE

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### FLUID FLOW DURING REGIONAL METAMORPHISM AT MID-CRUSTAL DEPTHS, EAST-CENTRAL VERMONT

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#### INTRODUCTION

The purpose of the trip is to examine the mineralogical effects of progressive Barrovian regional metamorphism and contemporaneous fluid flow on the Siluro-Devonian Waits River and Gile Mountain Formations. The trip consists of 4 stops with relatively long drives in-between (30-45 minutes). The long drives are necessary to visit low-grade exposures of both formations and their higher-grade equivalents. On a more positive note, the long drives provide tangible evidence of the enormous area that was infiltrated by reactive aqueous fluids during the regional metamorphic event (at least 3000 km<sup>2</sup> in Vermont alone). The rocks, especially at Stops 1-3, are utterly unremarkable in outcrop and hand specimen. These outcrops demonstrate that completely ordinary-looking metamorphic rocks may contain a rich history of infiltration-driven metamorphic mineral reactions.

Locations of all stops are shown on Figure 1, a simplified geologic map of east-central Vermont. Stop 1 is located at 43° 57.41' N latitude and 72° 7.95' W longitude, just off the eastern margin of the map, precisely at the contact between the Gile Mountain Formation and the pre-Silurian Orfordville Formation (see Fig. 1 of Ferry, 1988). Basic geologic relations in the area of Figure 1 have been established by Doll (1944), White and Jahns (1950), Lyons (1955), Doll and others (1961), Woodland (1977), Fisher and Karabinos (1980), and Hatch (1988a). Hatch (1988b) and Hueber et al. (1990) present the most recent summary of the lithology, stratigraphy, and age of the formations. The Waits River Formation lies stratigraphically beneath the Gile Mountain Formation and is primarily composed of interbedded micaceous limestone and pelitic schist. Individual lithologic layers vary greatly in thickness from decimeters to meters. There is a general increase in the amount of carbonate rock from ≈30-50% at the western exposure of the formation to >80% in the east, and Hatch (1988b) subdivided the formation into two facies based on the proportion of schist and carbonate rock. The Gile Mountain Formation is typically composed of interbedded pelitic schist, micaceous sandstone, and minor (≤ 5%) micaceous carbonate rock. Hatch (1988b) mapped four facies of the Gile Mountain Formation differentiated by the relative proportions of sandstone and schist. The easternmost margin of the Gile Mountain Formation consists of a thin unit of pelitic schist traditionally mapped as the Meetinghouse Slate member (not shown at the scale of Fig. 1). The Northfield Formation overlies or is laterally equivalent to the youngest portions of the Gile Mountain Formation (Hatch, 1988b) and is primarily composed of pelitic schist. The Standing Pond Volcanics (not shown at the scale of Fig. 1), composed of metamorphosed mafic flows, tuffs, and volcani-clastic sediments, separates the Gile Mountain and Waits River Formations along their eastern contact.

The stratigraphic age of the Gile Mountain and Waits River Formations is controversial. Spear and Harrison (1989) report one Ordovician, one Silurian, and 4 Devonian <sup>40</sup>Ar/<sup>39</sup>Ar ages for hornblende from the Gile Mountain Formation and Standing Pond Volcanics. The occurrence of plant fossils in the Gile Mountain Formation, however, requires a Devonian age for at least part of the formation (Hueber et al., 1990). While recognizing the problematic nature of the Ordovician and Silurian hornblende ages, Hueber et al. concluded that field, paleontological, and radiometric data are most consistent with a Siluro-Devonian age for the Waits River, Gile Mountain, and Northfield Formations. All radiometric dates from hornblendes collected in and near the domes

in Figure 1, are 350-397 Ma (Spear and Harrison, 1989). The highest grade regional metamorphism of Siluro-Devonian units in the area of Figure 1 therefore is unequivocally Devonian in age.

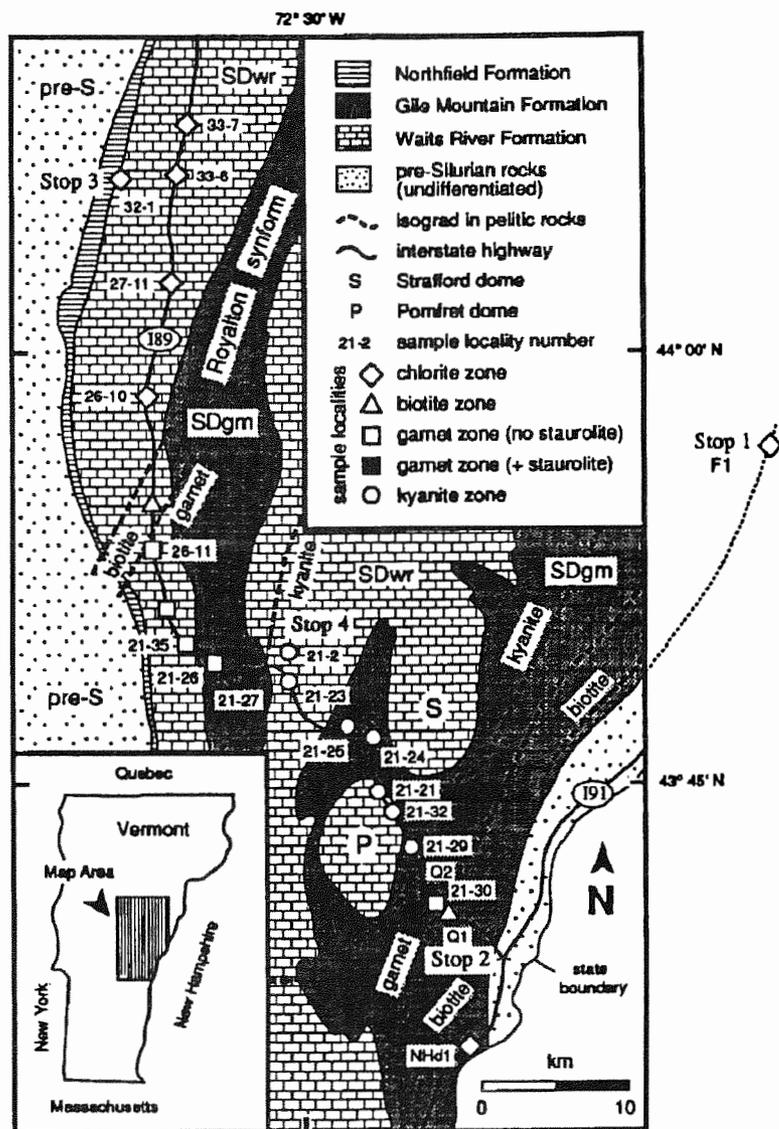


FIG. 1 Geologic sketch map of east-central Vermont. Data from Lyons (1955) and Doll et al. (1961). Sample location numbers are those from Ferry (1988, 1994).

The Waits River and Gile Mountain Formations were folded into recumbent folds with westward transport direction (nappe stage of deformation) and later into structural domes (dome stage). Deformation and regional metamorphism occurred during the Devonian Acadian orogeny (Thompson et al., 1968; Thompson and Norton, 1968; Osberg et al., 1989). The Stafford and Pomfret domes lie along the axis of the Willoughby-Chester antiform (Osberg et al., 1989; the Stafford-Willoughby arch of Doll et al., 1961). With respect to minerals in pelitic schists, rocks in the area were metamorphosed to conditions between those of the chlorite zone and those of the kyanite zone. The kyanite and eastern biotite and garnet isograds in Figure 1 were taken from Lyons (1955) and from the metamorphic map of Doll et al. (1961). Positions of the western biotite and garnet isograds are based on mineral assemblages observed in pelitic schists from the Waits River Formation exposed along Interstate 89 highway. Isograds (not shown) have also been mapped in the area based on the appearance of oligoclase, biotite, and amphibole in metacarbonate rocks from the Waits River Formation (Ferry, 1992). The 3 isograds separate the

ankerite-albite, ankerite-oligoclase, biotite, and amphibole zones which correlate approximately with the chlorite, biotite, garnet, and kyanite zones, respectively, in pelitic schists. The axis of the Willoughby-Chester antiform is coincident with the long axis of the metamorphic high defined by the kyanite zone in Figure 1. The geometry of inclusion trails in garnets on the flanks of the antiform south of the area in Figure 1 indicates that garnet growth in the Waits River Formation initiated during the nappe stage of deformation and continued during the dome stage with no long hiatus in-between (Rosenfeld, 1968). The region where the antiform's axis was later to develop therefore was already hot relative to adjacent regions at the nappe stage. Mineral equilibria suggest that the peak of regional metamorphism post-dated recumbent folding and was contemporaneous with or slightly predated development of the domes and antiform (Menard and Spear, 1990; Barnett, 1990).

### Description of Localities

- 0.0 miles Trip starts at Exit 15, Interstate 91 (Fairlee, Vermont). Turn west after exit from I - 91
- 0.1 miles Turn north on Lake Morey Road East.
- 2.6 miles Turn north on Maurice A. Roberts Memorial Highway.
- 4.1 miles STOP 1 (30 minutes). Roadcut on Maurice A. Roberts Memorial Highway (Fairlee, Vermont, 7.5' Quadrangle; Location Fl of Ferry, 1988).

The roadcut exposes the Meetinghouse Slate member of the Gile Mountain Formation in contact with the Ordovician Orfordville Formation. The Meetinghouse Slate is represented by the grey phyllites at the west end of the exposure. The Orfordville Formation is represented by black graphitic, sulfidic phyllites interlayered with felsic and mafic and metavolcanic rocks at the east end of the exposure which are equivalent to metavolcanic rocks of the Partridge Formation and Ammonoosuc Volcanics elsewhere in eastern Vermont and western New Hampshire. There is continuing debate whether the contact between the Gile Mountain Formation and Ordovician units to the east, the so-called Monroe Line, is a fault or an unconformity. There is evidence here to support both sides of the argument. Recent opinion favors the interpretation of the Monroe Line not just as a fault (Hatch, 1988a) but as a terrane boundary separating the terrane hosting the Gile Mountain and Waits River Formations to the west from the Central Maine terrane to the east.

Metamorphic grade at this stop is that of the chlorite zone. Mineralogically the Meetinghouse Slate is composed of muscovite + chlorite + albite + quartz + ankerite + rutile + pyrite + chalcopyrite  $\pm$  siderite. Ankerite and siderite occur as porphyroblasts that weather as mm-size rusty spots on rock surfaces. Pelitic schists from the chlorite zone everywhere in east-central Vermont contain ankerite+siderite. The phyllites are the low-grade equivalents of biotite schists of the Gile Mountain Formation at the next stop on the trip and at higher grades. Metavolcanics of the Orfordville Formation are composed of chlorite + ankerite + albite + quartz + rutile  $\pm$  muscovite  $\pm$  calcite  $\pm$  pyrite  $\pm$  chalcopyrite. Rusty-weathering ankerite porphyroblasts are visible on rock surfaces. The metavolcanics are the low-grade equivalents of greenschists that we will drive by on the way to the next stop.

Make U-turn and return to I - 91.

- 5.6 miles Turn east on Lake Morey Road East.
- 8.1 miles Junction I - 91. Take southbound I - 91 to White River Junction.
- 29.9 miles Take Exit 10N from I - 91 to northbound Interstate 89.

Roadcuts at the junction of I - 89 and I - 91 expose the Ordovician Ammonoosuc Volcanics, lithologic equivalents of the metavolcanic rocks at Stop 1. Metamorphic grade is that of the biotite zone and metavolcanics rocks mineralogically are respectable greenschists composed of actinolite+ chlorite + plagioclase + quartz + epidote + calcite + sphene  $\pm$  biotite  $\pm$  pyrite  $\pm$  pyrrhotite  $\pm$  chalcopyrite. The transition from the chlorite zone to the biotite zone was associated with decarbonation-dehydration reactions in metavolcanics such as

$0.01 \text{ muscovite} + 1.23 \text{ chlorite} + 8.60 \text{ quartz} + 4.86 \text{ calcite} + 0.15 \text{ rutile} = 1.03 \text{ actinolite} + 1.25 \text{ epidote} + 0.02 \text{ plagioclase} + 0.13 \text{ sphene} + 4.86 \text{ CO}_2 + 3.38 \text{ H}_2\text{O}$

(per liter rock). During progress of the reaction, minerals were in equilibrium with a CO<sub>2</sub>-H<sub>2</sub>O fluid with low mole fraction CO<sub>2</sub>, X<sub>CO<sub>2</sub></sub> = 0.04-0.05. The difference between the CO<sub>2</sub>-rich volatile mixture released by the

rocks during reaction and the H<sub>2</sub>O-rich fluid with which the rocks were in equilibrium can best be explained by infiltration of the greenschists by chemically reactive aqueous fluid as the reaction proceeded (Ferry, 1988).

32.8 miles      Take Exit 1 from I - 89 (Route 4).  
 33.1 miles      Turn northwest onto Route 4 West.  
 33.4 miles      STOP 2 (30 minutes). (Quechee, Vermont, 7.5' Quadrangle;  
 Location Q1 of Ferry, 1988, 1994). Park on southwest side of  
 overpass over I - 89. Climb down to roadcut on southbound I - 89  
 leading up to southbound Exit 1 from I - 89.

The roadcut exposes typical Gile Mountain Formation consisting of interbedded pelitic schist, micaceous sandstone, and micaceous limestone. Metamorphic grade is that of the biotite zone. Schists typically contain the mineral assemblage biotite + muscovite + chlorite + oligoclase + quartz + ankerite + ilmenite + pyrrhotite + chalcopyrite ± pyrite that developed from assemblages in equivalent rocks from the chlorite zone (Stop 1) by reactions such as

$$1.38 \text{ muscovite} + 0.05 \text{ chlorite} + 0.33 \text{ quartz} + 0.49 \text{ ankerite} + 2.64 \text{ siderite} + 0.34 \text{ rutile} = 1.49 \text{ biotite} + 0.45 \text{ plagioclase} + 0.25 \text{ ilmenite} + 3.61 \text{ CO}_2 + 0.08 \text{ H}_2\text{O}$$

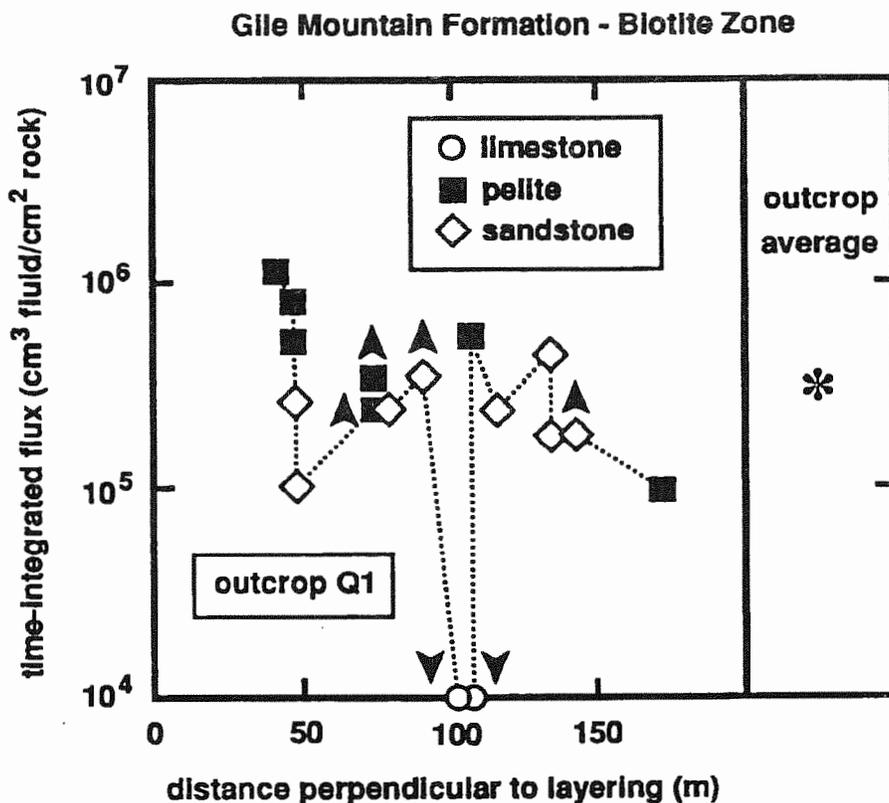
(per liter rock). Micaceous sandstones typically contain the same mineral assemblage as schists with and without calcite, and the assemblage developed from minerals in equivalent rocks from the chlorite zone by reactions such as

$$0.62 \text{ muscovite} + 0.29 \text{ chlorite} + 0.31 \text{ quartz} + 0.20 \text{ ankerite} + 0.44 \text{ calcite} + 0.17 \text{ rutile} = 0.65 \text{ biotite} + 0.63 \text{ plagioclase} + 0.13 \text{ ilmenite} + 0.83 \text{ CO}_2 + 1.11 \text{ H}_2\text{O}$$

(per liter rock). Micaceous limestones typically contain the assemblage muscovite + ankerite + calcite + quartz + albite + rutile + pyrrhotite ± pyrite ± chalcopyrite which is the same as that observed in equivalent lithologies in the chlorite zone.

Mineral equilibria in rocks from the garnet and kyanite zone record pressures of 7-8 kb. The preferred peak temperature of 475°C in the biotite zone is based on an extrapolation of peak temperatures recorded by mineral equilibria in the kyanite zone along an apparent ground surface temperature gradient of ~10°C/km to the biotite zone. At these P-T conditions minerals participating in the biotite-forming reactions in pelites and sandstones would have been in equilibrium with CO<sub>2</sub>-H<sub>2</sub>O fluids with X<sub>CO<sub>2</sub></sub> in the narrow range 0.02-0.03 (Ferry, 1994). The significant difference in composition between the CO<sub>2</sub>-rich mixture of volatiles released by the biotite-forming reactions and the H<sub>2</sub>O-rich fluids with which the rocks were in equilibrium is evidence that the sandstones and schists were infiltrated by reactive aqueous fluids as the reactions progressed. Amounts of fluid involved, calculated as a time-integrated flux, were estimated using measured reaction progress and the mass continuity equation (see Baumgartner and Ferry, 1991, for details). Results for 17 samples of metasediment collected from this outcrop are shown in Figure 2. An average value for the outcrop as a whole was estimated by weighting the average value for each lithology by its measured abundance in outcrop. Progress of the biotite-forming reaction in schists and sandstones record passage of 10<sup>5</sup>-10<sup>6</sup> cm<sup>3</sup> fluid/cm<sup>2</sup> rock. As a group, schists record slightly larger values than sandstones. The amount of fluid is astonishingly large, corresponding to a column of fluid with 1 cm<sup>2</sup> cross-sectional area and a length of 1-10 km flowing through each cm<sup>2</sup> of sandstone and schist. Two limestone samples contain no biotite, show no evidence for devolatilization, and therefore record a zero value of time-integrated fluid flux. Taken together results in Figure 2 indicate that most or all metamorphic fluid flow was parallel to lithologic layering. The apparent lack of cross-bed flow is testimony to the first-order control that lithology exerts on the geometry of fluid flow during regional metamorphism. Some layers, such as the schists, were relatively permeable and acted as metamorphic aquifers. Other layers, such as the sandstones, were less permeable and acted as aquitards. Still other layers, such as the micaceous limestones evidently were impermeable and acted as metamorphic aquicludes. The prograde decarbonation reactions that sandstones and schists experienced, along with the evidence for layer-parallel flow, indicate that fluid flow during metamorphism was in the direction of increasing temperature, east to west at this exposure.

The petrologic evidence for interaction of rocks at this outcrop with chemically reactive fluids during metamorphism is independently substantiated by stable isotope data for schists and sandstones from the exposure (Chamberlain et al., 1990). The whole-rock δ<sup>18</sup>O and δ<sup>13</sup>C of 3 sandstones and the whole-rock δ<sup>13</sup>C of one schist collected from this exposure are significantly different from the carbon- and oxygen-isotope composition of equivalent rocks in the chlorite zone (collected from outcrop NHdl, Fig. 1). The differences are larger than can be



**FIG. 2** Calculated time-integrated flux plotted as a function of distance perpendicular to lithologic layering in outcrop Q1 from the biotite zone of the Gile Mountain Formation. Upward pointing arrowheads flag carbonate-free samples that register minimum values; downward pointing arrowheads flag two samples that record zero time-integrated flux and which plot off the bottom of the diagram. Asterisk denotes average time-integrated flux for the outcrop as a whole taking into account the measured relative proportions of different rock types. Fluid flow during metamorphism was spatially heterogeneous at the scale of outcrop Q1. From Ferry (1994).

explained by the observed progress of the biotite-forming reaction in each sample, and are attributed to interaction of sandstone and schist at this outcrop with metamorphic fluid with which they were out of C- and O-isotopic exchange equilibrium.

Make U-turn and return to northbound I - 89.

33.7 miles      Entrance to northbound I - 89 from Route 4. Take I - 89 north towards Barre and Montpelier.

In the next 20 miles I - 89 crosses the stratigraphic contact between the Gile Mountain and Waits River Formations 5 times (Fig. 1). The Waits River Formation is easily recognized at high speed by its m-thick brown-weathering massive limestone beds. Between the first two crossings the Waits River Formation is exposed by uplift in the Pomfret dome. The first rest area that we pass is located in the Waits River Formation in the dome (location 21-32, Fig. 1). The Pomfret dome lies along the axis of the Willoughby-Chester antiform, a structural and metamorphic high that strikes over 200 km southwest-northeast from southern Vermont to southern Quebec. The route after Stop 2 initially traverses the biotite, garnet, and kyanite zones in the direction of increasing metamorphic grade. Between the fourth and fifth crossings of the contact, the Gile Mountain Formation reappears by down-folding in the Royalton synform. The maximum in metamorphic grade along the route occurs between

the Pomfret dome and the Royalton synform. Once in the Royalton synform, metamorphic grade decreases along the route from conditions of the kyanite zone to those of the chlorite zone.

73.4 miles      Take Exit 5 from I - 89 (Route 64).  
 73.5 miles      Turn west on Route 64 West.  
 75.5 miles      STOP 3 (20 minutes). Roadcut on Route 64 (Roxbury, Vermont,  
 7.5' Quadrangle).

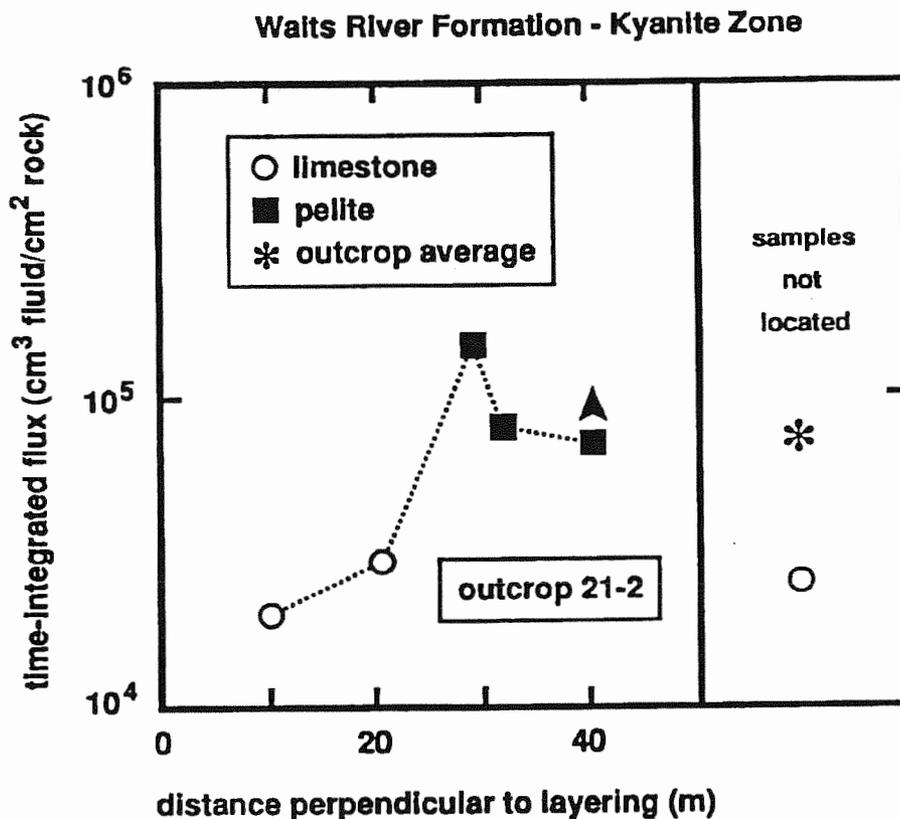


FIG. 3 Calculated time-integrated flux plotted as a function of distance perpendicular to lithologic layering in outcrop 21-2 from the kyanite zone of the Waits River Formation. Symbol in the right-hand panel refers to results for a sample whose position within the outcrop was not recorded. Other features as in Figure 2. Pelites record higher values of time-integrated flux than do interbedded limestones. From Ferry (1994).

We have traveled across the axis of the Willoughby-Chester antiform and are now at the westernmost margin of the Waits River Formation. The Waits River Formation is composed lithologically of interbedded pelitic phyllites and micaceous limestones. The proportion of limestone to pelite here is  $\approx 3:7$ ; due to facies changes, the ratio increases from west to east.

Metamorphic grade at this stop is that of the chlorite zone. Mineralogically the phyllites are composed of muscovite + chlorite + albite + quartz + ankerite + pyrrhotite + chalcopyrite + rutile or ilmenite  $\pm$  pargonite  $\pm$  calcite. All pelitic rocks in the chlorite zone of the Waits River Formation contain carbonate as do chlorite-zone pelites east of the Willoughby-Chester antiform. Limestones contain muscovite + calcite + ankerite + albite + quartz + rutile  $\pm$  pargonite  $\pm$  pyrite  $\pm$  pyrrhotite  $\pm$  chalcopyrite. The phyllites and limestones here are the low-grade equivalents of garnet schists and calc-silicate granofelses of the Waits River Formation at the next stop on the trip.

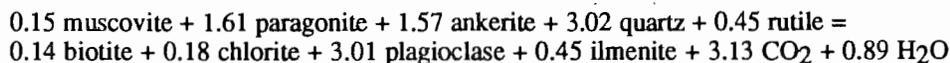
Note the small abundance of quartz veins in outcrop. The volume concentration, estimated by traversing the outcrop with a measuring tape, is  $\approx 0.04\%$ . In general, quartz-veins are much less abundant west of the Willoughby-Chester antiform than in the axial region.

Make U-turn and return to I - 89.

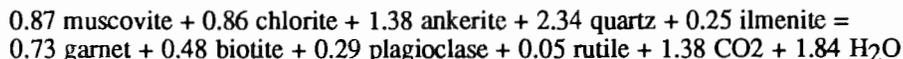
77.5 miles	Junction I - 89. Take southbound I - 89 back towards White River Junction
98.4 miles	Take Exit 3 from I - 89 (Route 107).
98.6 miles	Turn east on Route 107 East.
99.4 miles	Turn south on Route 14 South.
103.0 miles	Turn north on Route 110 North.
103.5 miles	Stop 4 (40 minutes). Roadcut on Route 110 (South Royalton, Vermont, 7.5' Quadrangle; Location 21-2 of Ferry, 1992, 1994).

Since the last stop we traversed the chlorite, biotite, and garnet zones and are now located in the Waits River Formation in the axial region of the Willoughby-Chester antiform. The ratio of limestone to schist interbeds is larger than at the last stop,  $\approx 4:6$ .

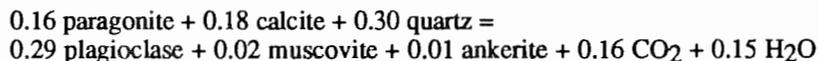
Metamorphic grade at this stop is that of the kyanite zone. Mineralogically the schists are composed of biotite + garnet + muscovite + chlorite + plagioclase + quartz + ilmenite + rutile + pyrrhotite + ankerite or calcite or both  $\pm$  clinozoisite  $\pm$  chalcopyrite. Only rarely do schists in the kyanite zone of the Waits River Formation have bulk compositions that develop staurolite or kyanite. The carbonate rocks contain biotite + chlorite + calcite + ankerite + plagioclase + quartz + rutile + pyrrhotite + amphibole or muscovite  $\pm$  chalcopyrite. Assemblages in the pelites developed from minerals like those observed at the last stop in two steps. One step produced biotite by a reaction such as



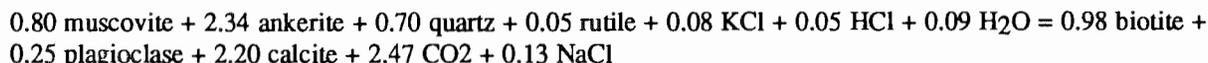
(per liter rock) and the other produced garnet by a reaction such as



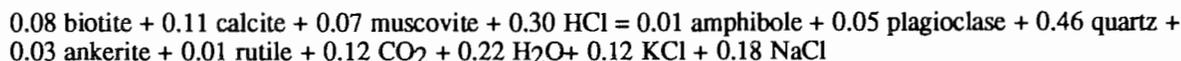
(per liter rock). Assemblages in the metacarbonate rocks developed in three steps. One step produced plagioclase by a reaction such as



(per liter rock), a second produced biotite by a reaction such as



(per liter rock), and a third produced calcic amphibole by a reaction such as



(per liter rock). All prograde plagioclase-, biotite-, garnet-, and amphibole-producing reactions in schists and carbonate rocks of the Waits River Formation involved decarbonation. Biotite- and amphibole-forming reactions in the metacarbonates commonly were allochemical involving mass transfer of K and Na.

This outcrop is well-endowed with mineral thermometers and mineral barometers. Biotite-garnet and calcite-ankerite pairs record T in the range  $490^\circ$ - $545^\circ\text{C}$  and  $520^\circ$ - $545^\circ\text{C}$ , respectively. Occurrences of garnet-rutile-ilmenite-plagioclase-quartz (GRIPS) record P = 7.1-7.6 kb; muscovite-biotite-garnet-plagioclase records P = 6.9-7.9 kb; clinozoisite-calcite-quartz-garnet-plagioclase-CO<sub>2</sub>/H<sub>2</sub>O fluid records P = 6.8 kb; and muscovite-ankerite-quartz-

biotite-calcite-plagioclase-clinozoisite-CO<sub>2</sub>/H<sub>2</sub>O fluid records P = 6.8 kb. These pressures and temperatures are typical of those recorded by mineral equilibria in schists, sandstones, and carbonate rocks from the Waits River and Gile Mountain Formations in the kyanite zone (P = 7-8 kb; maximum T ≈ 550°C). Temperature recorded by geothermometers in the garnet zone are lower, ≈ 500°C. At peak metamorphic P-T conditions inferred for the biotite, garnet, and kyanite zones, reactants and products of the prograde reactions in schists and carbonate rocks from the Waits River Formation were in equilibrium with H<sub>2</sub>O rich CO<sub>2</sub>-H<sub>2</sub>O fluids with the following compositions: biotite-forming reaction in schists, X<sub>CO</sub> ≈ 0.02; garnet-forming reaction in schists, X<sub>CO</sub> ≈ 0.08-0.09; plagioclase-forming reaction in metacarbonates, X<sub>CO<sub>2</sub></sub> ≈ 0.02; biotite-forming reaction in metacarbonates, X<sub>CO<sub>2</sub></sub> ≈ 0.08; amphibole-forming reaction in metacarbonates, X<sub>CO<sub>2</sub></sub> ≈ 0.070.08.

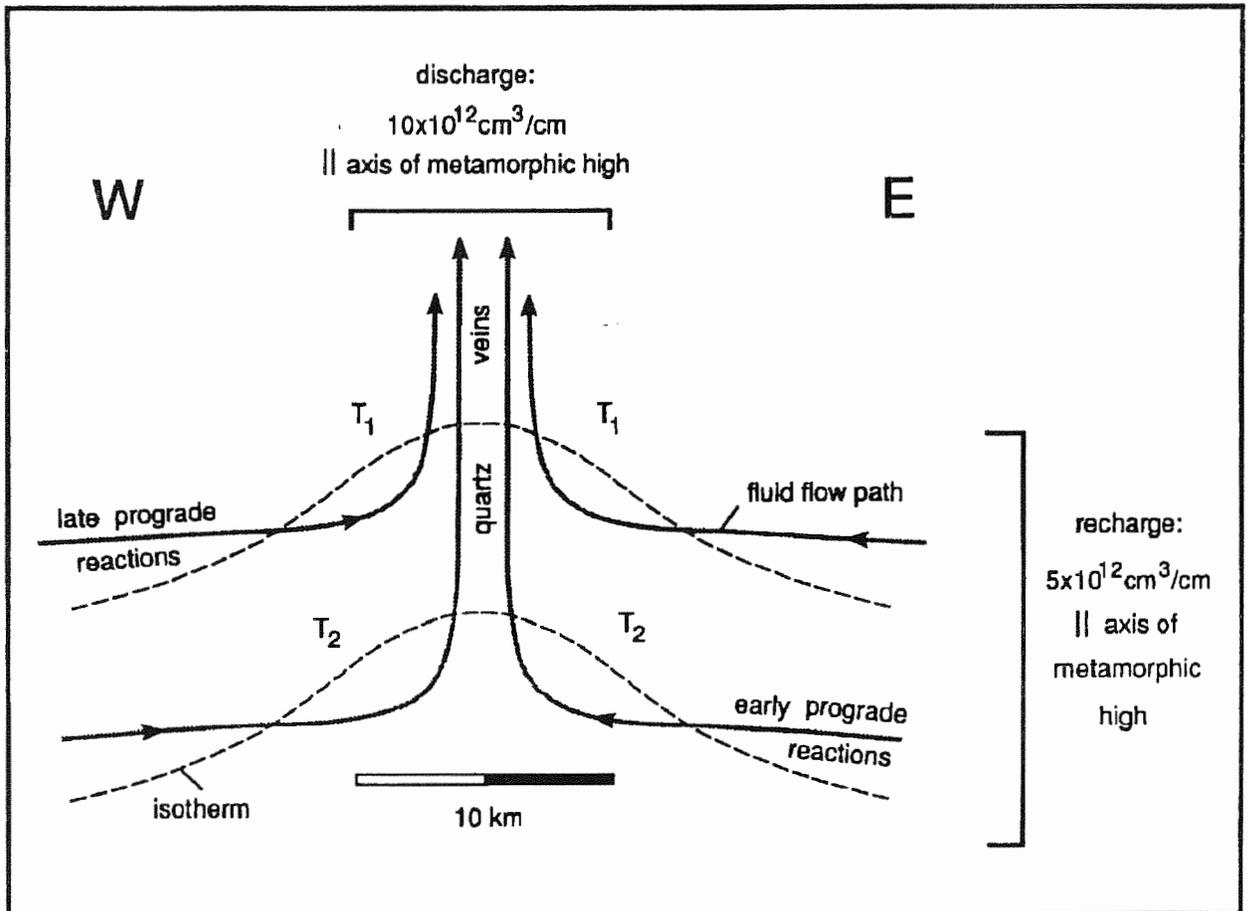
As at Stop 2 (location Q1), the significant difference in composition between the CO<sub>2</sub>-rich mixture of volatiles released by the prograde metamorphic reactions and the H<sub>2</sub>O-rich fluids with which the rocks at the same time were in equilibrium is evidence that both schists and limestones were infiltrated by reactive aqueous fluids as the reactions progressed. Amounts of fluid involved, calculated for 6 samples from the outcrop, are shown in Figure 3. Plotted values correspond to the sum of time-integrated flux recorded by each different prograde reaction in the two rock types. An average value for the outcrop as a whole was estimated by weighting the average value for each lithology by its measured abundance in outcrop. Schists record passage of (0.7-1.4) 10<sup>5</sup> cm<sup>3</sup> fluid/cm<sup>2</sup> rock; metacarbonate rocks record lower values, (0.2-0.3) 10<sup>5</sup> cm<sup>3</sup>. Results in Figure 3, as do those in Figure 2, indicate that most or all metamorphic fluid flow was parallel to lithologic layering. The prograde decarbonation reactions that limestones and schists experienced, along with the evidence for layer-parallel flow, indicate that fluid flow during metamorphism was in the direction of increasing temperature, west to east at this exposure.

The petrologic evidence for interaction of rocks at this outcrop with chemically reactive fluids during metamorphism is independently substantiated by stable isotope data for metacarbonate rocks from the Waits River Formation (Barnett and Chamberlain, 1991; Stern et al., 1992). Many carbonate rocks in the biotite, garnet, and kyanite zones are significantly depleted in <sup>18</sup>O by up to 9‰ δ<sup>18</sup>O, relative to equivalent rocks in the chlorite zone. The depletions are larger than can be explained by the observed progress of the prograde devolatilization reactions in each sample and are attributed to interaction of rocks with metamorphic fluid with which they were out of isotopic exchange equilibrium. Furthermore, both <sup>18</sup>O and δ<sup>13</sup>C in individual outcrops is progressively more uniform at higher grades than at lower grades of metamorphism. The homogenization of O and C isotopes is attributed to isotope exchange between rock and pervasively through-flowing CO<sub>2</sub>-H<sub>2</sub>O fluid during metamorphism. <sup>18</sup>O-depletion fronts are observed at the contact between the Waits River and Gile Mountain Formations. The magnitude and direction of displacement of the fronts record a time-integrated fluid flux of 10<sup>5</sup>-10<sup>6</sup> cm<sup>3</sup> fluid/cm<sup>2</sup> rock in the direction of increasing temperature, in good agreement with amounts and direction of flow inferred from petrologic data.

If fluid flowed from the cold flanks to the axis of the nascent Willoughby-Chester antiform during metamorphism, the fluid had to somehow later escape. Notice the much greater abundance of quartz veins in this outcrop (11%) compared to the last (<0.1%). The quartz vein abundance here is typical of that in the axial region of the antiform as a whole. Quartz veins inevitably form by fracture flow of quartz-saturated fluids in the direction of decreasing temperature and pressure. At least some quartz veins here and throughout the kyanite zone are inferred to represent the discharge paths along which fluids escaped flowing down-temperature from the metamorphic system.

A variety of field, textural, and isotopic data bear on the timing, geometry, and duration of fluid flow in eastern Vermont during the Acadian orogeny. The foliation at this outcrop and throughout the kyanite zone is a product of nappe-stage deformation (Woodland, 1977). Most quartz veins in the outcrop therefore formed prior to or synchronous with the nappe-stage deformation. A few quartz veins, however, cross-cut the foliation and post-date the nappe stage. In thin sections of schists from the kyanite zone garnets commonly contain inclusion trails that define a foliation at a high angle to that of micaceous minerals in the matrix. Garnet growth therefore initiated prior to or during nappe-stage deformation. Mineral equilibria involving garnet in schists collected around the Strafford dome record uplift (Menard and Spear, 1990). Garnet growth evidently continued into the dome stage of deformation. Both pervasive fluid flow in the direction of increasing temperature (recorded by garnet growth) and fracture flow in the direction of decreasing temperature (recorded by quartz veins) occurred during the nappe and dome stages. Observations at this outcrop and nearby are corroborated by studies along the Willoughby-Chester antiform to the south. Patterns of oxygen isotope zoning in garnets on the west limb of the Athens dome record

time-integrated fluxes of  $10^4$ - $10^5$   $\text{cm}^3$  fluid/ $\text{cm}^2$  rock during nappe-stage deformation with flow towards the dome's long axis (Chamberlain and Conrad, 1993). The classic studies of rotated garnets by Rosenfeld (1968, 1970) indicate that garnet growth (and hence fluid flow) occurred in the Waits River Formation around the Athens and Chester domes during both nappe and dome-stage deformation with little or no hiatus in-between. (See Hayward, 1992, however, for an alternative interpretation of garnets from the same area.) Rosenfeld (1968) further concluded that lithologic layering was probably near-horizontal both prior to the nappe stage and at the initiation of doming. Given the evidence for layer-parallel fluid flow at Stops 2 and 4 (locations Q1 and 21-2), pervasive up-temperature fluid flow recorded by the prograde biotite- and garnet-forming reactions therefore must have been near-horizontal as well. The duration of flow, as estimated by the duration of garnet growth, was probably  $\approx 10 \pm 4$  million years (Christensen et al., 1989).



**FIG. 4** Schematic cross-section through the metamorphic hydrothermal system centered on the axis of the Willoughby-Chester antiform in east-central Vermont. The section is perpendicular to the axis of the antiform. Isotherms are represented by dashed curves ( $T_2 > T_1$ ). Fluid flow paths are represented by thick grey curves with arrows. From Ferry (1992).

Field, petrologic, textural, and isotopic evidence therefore converge on an overall geometry of regional metamorphic fluid flow like that illustrated in Figure 4. Subhorizontal, pervasive, layer-parallel, east-west flow drove prograde decarbonation reactions and effected stable isotopic alteration in schists, sandstones and limestones of the Waits River and Gile Mountain Formations. Flow began during the nappe stage of deformation and persisted into the dome stage. The terrane average amount of flow at the present level of exposure was  $\approx 5 \cdot 10^5$   $\text{cm}^3$  fluid/ $\text{cm}^2$  rock (Ferry, 1994); integrated over 50 km of crust, the amount is  $\approx 5 \cdot 10^{12}$   $\text{cm}^3$  fluid per cm rock

parallel to the long axis of the flow system. Flow was likely buoyancy-driven associated with the thermal anomaly which developed along what was later to become the axis of the Willoughby-Chester antiformal. Fluid escaped along subvertical fractures now preserved as quartz veins. A 10% concentration of quartz veins in the axial region records a time-integrated exit flux of  $\approx 10 \cdot 10^{12}$  cm<sup>3</sup> fluid per cm rock parallel to the long axis of the flow system (Ferry and Dipple, 1992; Ferry, 1992). The fairly good quantitative agreement between independent estimates of the recharge and discharge fluxes provides additional support to the general model of flow proposed in Figure 4.

End of trip. Retrace route back to I - 89. Take I - 89 south to White River Junction.

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## MULTIPLE AMPHIBOLE ASSEMBLAGES AND WONESITE (Na-PHLOGOPITE), MT. CUBE QUADRANGLE, NH AND VT

by

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### INTRODUCTION

In the vicinity of Orfordville, New Hampshire, a sequence of meta-volcanic rocks (the Ammonoosuc and Post Pond volcanics) are exposed that display a wide variety of mineral assemblages. This variety arises from the diversity of bulk compositions. Original rock types range from basalts to rhyolites and many units were hydrothermally altered following deposition, presumably in a sub-seafloor environment.

The purpose of this trip is to examine various aspects of the mineralogy and petrology of these rocks. Multiple amphibole assemblages are common, and a rare, Na-rich trioctahedral mica (wonesite) was discovered here.

#### Mileage

0.0 Fairlee, VT. Intersection of US Route 5 and bridge to Orford, NH. Proceed south on US Route 5.  
10.9 Road cuts on right (west) side of US Route 5. Park on right just south of the outcrop.

**Stop 1. US Route 5 Locality (45 MINUTES).** (Mt. Cube 15 minute, Lyme 7 1/2 minute quadrangle). The purpose of this stop is to examine multiple amphibole assemblages in the Ammonoosuc (= Post Pond) volcanics. Interesting assemblages include (1) four coexisting amphiboles (anthophyllite + gedrite + cummingtonite + hornblende) ± garnet ± plagioclase + quartz + ilmenite; (2) staurolite + hornblende ± gedrite ± garnet + quartz + plagioclase + ilmenite; and (3) two coexisting plagioclase feldspars (An<sub>40</sub> + An<sub>87-92</sub>). A generalized phase diagram for this area is shown in Figure 5. Details of the phase petrology of these rocks are contained in Spear (1980a, 1980b, 1981, 1982), Grove et al. (1983), and Spear and Rumble (1986a, 1986b). P-T paths from this locality are discussed in Spear and Rumble (1986a) and Kohn et al. (1992).

73-18. The rocks at this station are relatively calcic and contain hornblende + plagioclase + quartz + epidote + dolomite + calcite + biotite ± chlorite ± scapolite.

73-19. The north end of this station contains calcic bulk compositions similar to 73-18 with assemblages such as hornblende + plagioclase + epidote + dolomite + calcite + scapolite. In the middle and south end of this station are Mg-rich, calcic bulk compositions with the assemblage hornblende + quartz + plagioclase + cummingtonite + chlorite + actinolite ± dolomite + calcite ± biotite. This is the only locality in the area that contains actinolite + hornblende assemblages.

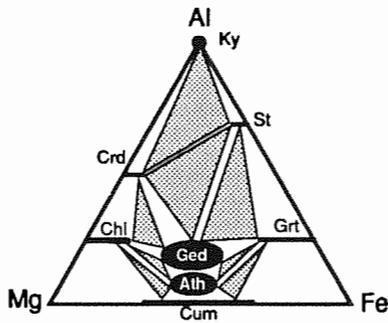


FIG. 5. Phase diagram showing the assemblages found at the US Route 5 locality.

*US Route 5 locality walking log.* This text is to accompany the sketch map of sample stations shown in Figure 6. Start at the north end of the roadcut.

73-20. One sample from this station contains calcic bulk compositions with hornblende + mixed plagioclases (An<sub>45</sub>, An<sub>65</sub>, An<sub>85</sub>) + quartz + epidote + chlorite. The plagioclase is too small to be sure that the analyses are representative of single phases. The significant feature of this station is the presence of the most Fe-rich, 4-amphibole rocks in the area. The representative assemblage is hornblende + cummingtonite + gedrite + anthophyllite + garnet + plagioclase + quartz + ilmenite. Note that this Fe-rich, 4-amphibole rock contains garnet. One sample

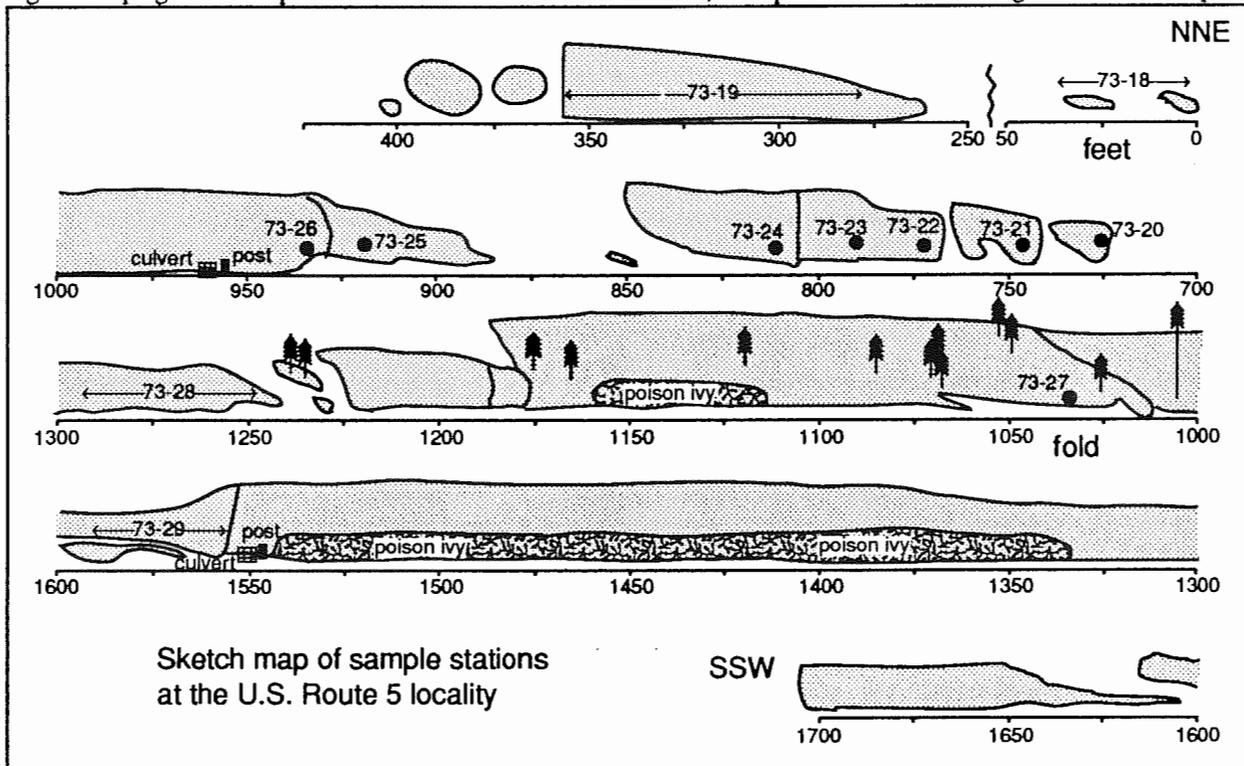


FIG. 6. Sketch map showing sample stations at the US Route 5 locality.

from this station (73-20C) was the subject of P-T path calculations based on the zoning observed in garnet (Spear and Rumble, 1986a).

73-21 This station contains rocks with hornblende + garnet + gedrite + quartz + plagioclase.

73-22 Samples here contain hornblende + garnet + gedrite + plagioclase + quartz + staurolite ± cummingtonite ± anthophyllite. Note that staurolite only occurs as inclusions within garnet and gedrite.

73-23 Assemblages include (1) hornblende + garnet + gedrite + quartz + plagioclase ± biotite + staurolite (as inclusions within garnet, gedrite and hornblende) and (2) hornblende + plagioclase + quartz + biotite + cummingtonite + gedrite + garnet. Note the flattened lenticular shapes within the layering from stations 73-21 to 73-23. These are suspected to be primary volcanic features.

73-24 Hornblende + plagioclase + chlorite + biotite ± staurolite + ilmenite ± cummingtonite.

73-25 There are some calcic bulk compositions here that display considerable compositional heterogeneity. Assemblages in the calcic layers include hornblende + garnet + quartz + 2 plagioclases (An<sub>40</sub> and An<sub>87</sub>) + epidote ± calcite ± dolomite. More "normal" amphibolites containing hornblende + garnet + cummingtonite + plagioclase + quartz + ilmenite and hornblende + garnet + plagioclase ± staurolite + chlorite are also present.

73-26 This station contains rocks with the assemblage hornblende + one or two plagioclases (An<sub>40</sub> and An<sub>87</sub>) + quartz + chlorite ± staurolite + ilmenite.

73-27 This station contains a calcic bulk composition with the assemblage hornblende + plagioclase + quartz + calcite + chlorite + garnet + epidote and the carbonate-free assemblage hornblende + plagioclase + garnet + biotite + chlorite. There is a fold at this station, which can be viewed by looking south. The fold has a vergence of west over east and deforms an earlier foliation. The early foliation is interpreted as having formed during the nappe stage and the fold as having been produced during the formation of the Orfordville anticlinorium in the dome stage. Watch out for poison ivy from here on.

73-28 At the north end of this station is the interesting assemblage hornblende + garnet + chlorite + quartz + 2 plagioclases (An<sub>40</sub> and An<sub>90</sub>) + staurolite. Staurolite occurs as a matrix phase and also rimmed by plagioclase, suggesting a reaction texture. Toward the middle of this station are some calcic bulk compositions with epidote and carbonate-bearing assemblages. Toward the south end of this station are staurolite-bearing rocks with the assemblage hornblende + garnet + gedrite + staurolite ± biotite + plagioclase + quartz + ilmenite.

73-29 Some of the best staurolite-bearing assemblages occur here. One sample contains hornblende + gedrite + chlorite + staurolite + plagioclase ± quartz + biotite + ilmenite.

- 10.9 Continue south on US Route 5
- 12.2 Pompanoosuc. Turn right (west) on Route 132
- 12.6 Hogback Road. Turn right (north) up hill.
- 12.8 Drew Farm (permission required). Drive to upper quarry.

**Stop 3A. Upper Quarry in Ammonoosuc Volcanics.** (20 MINUTES) (Mt. Cube 15 minute, Lyme 7 1/2 minute quadrangle). The Ammonoosuc volcanics here display cross cutting dikes and agglomerate textures as evidence of their original volcanic nature. The thickness of the Ammonoosuc Volcanics here suggests that this locality may be near an ancient volcanic center. Assemblages at this outcrop are largely hornblende + plagioclase + ilmenite ± cummingtonite ± quartz ± epidote and a few assemblages of four coexisting amphiboles ± garnet have also been found. These assemblages have Fe/(Fe+Mg) values slightly lower than those at the US Route 5 locality.

**Stop 3B. Wonesite locality.** (2 HOURS). (Mt. Cube 15 minute, Lyme 7 1/2 minute quadrangle). A sodium-rich trioctahedral mica with a composition intermediate between Na-phlogopite (NaMg<sub>3</sub>AlSi<sub>3</sub>O<sub>10</sub>(OH)<sub>2</sub>) and talc (Mg<sub>3</sub>Si<sub>4</sub>O<sub>10</sub>(OH)<sub>2</sub>) (Fig. 7) was discovered in the Ammonoosuc (Post Pond) Volcanics in the SW corner of the Mt Cube quadrangle in 1979 and described by Spear, et al. (1981). The mineral was named wonesite in honor of David R. Wones.

Wonesite occurs in rocks that have been interpreted as altered volcanics and is associated with the minerals phlogopite, talc, chlorite, cordierite, anthophyllite and quartz. Wonesite displays extinction behavior characteristic of the non-brittle micas, and pleochroism similar to, although slightly paler than, coexisting phlogopite. A TEM study by Veblen (1983) showed that wonesite contains lamellae of talc and a more Na-rich phlogopite, with lamellae ranging in size from a few hundred angstroms to about 0.5 μm.

79-137. (Fig. 8) At this station the contact between relatively unaltered hornblende-plagioclase amphibolites and altered, orthoamphibole-bearing amphibolites is exposed. The contact is quite sharp and dips to the north west with



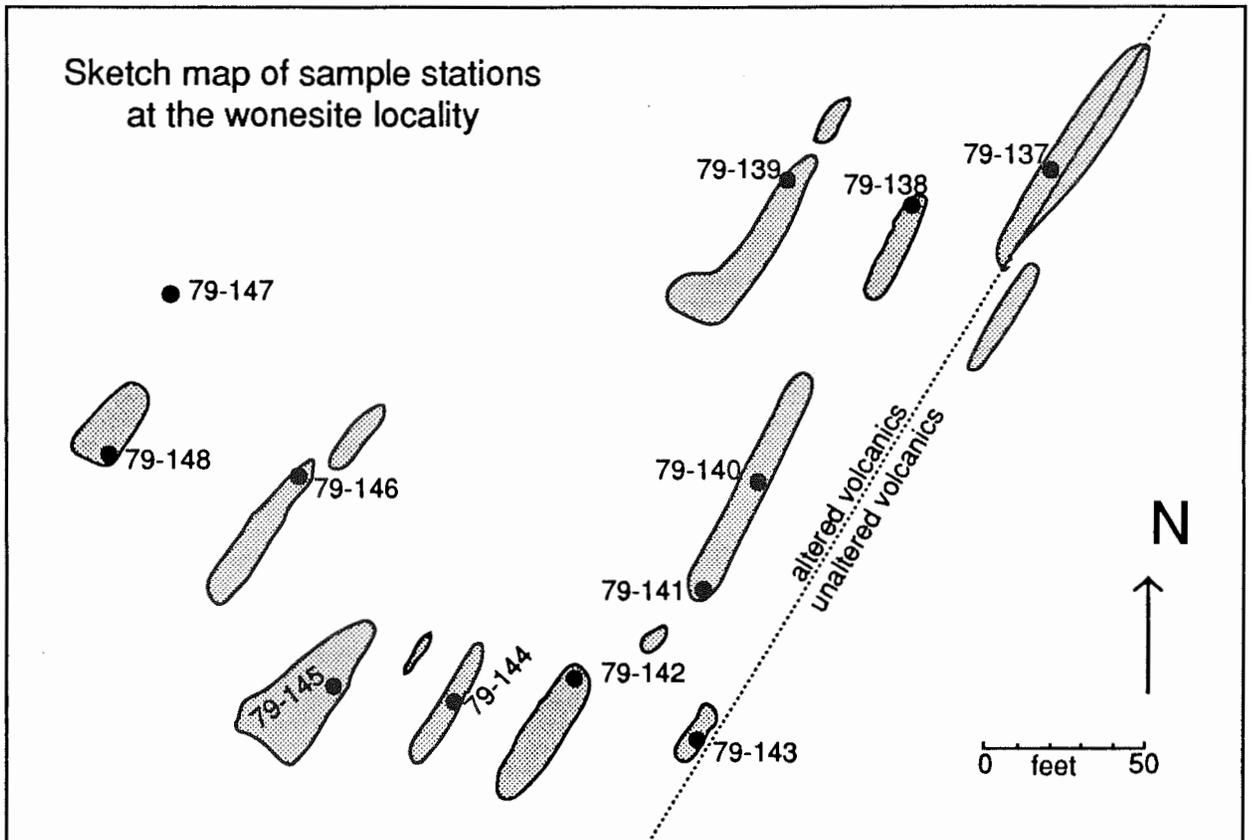


FIG. 8. Map showing outcrops and sample locations at the wonesite locality.

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## BLACK MOUNTAIN AND BEAVER BROOK, MT. MOOSILAUKE AREA, NEW HAMPSHIRE

by

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### INTRODUCTION

The purpose of the field trip to Black Mountain and Beaver Brook is to study two localities that show well exposed evidence of different types of fluid-rock interactions. The rocks of Black Mountain show examples of the bed-by-bed control of fluid compositions by mineral assemblage buffering. Infiltration, in contrast, is the prevalent mode of fluid-rock behavior at Beaver Brook. An important question to keep in mind throughout the field trip is "Why did Beaver Brook experience infiltration during metamorphism while Black Mountain did not?"

### MAPS AND LITERATURE

The area of the field trip is covered by two 7-1/2', 1/24,000, topographic maps: East Haverhill, NH and Mt. Moosilauke, NH (Fig. C-1). A key reference on the geology and petrology of the area is the seminal work of Marland P. Billings "Regional metamorphism of the Littleton-Moosilauke area, NH" (1937). The distribution of  $Al_2SiO_5$  minerals in the region is described by Rumble (1973). Pressure-temperature (P-T) estimates are given by K. V. Hodges and F. S. Spear (1982). The Black Mtn. and Beaver Brook localities are discussed, respectively, by Rumble (1978) and by Rumble, Ferry, Hoering, and Boucot (1982).

### STRATIGRAPHY AND GEOLOGIC HISTORY

The oldest stratified rocks in the Mt. Moosilauke region are the Ammonoosuc Volcanics of Ordovician age (Fig. 9). There are at least several kilometers of felsic and mafic volcanics in the Ammonoosuc. A more accurate estimate of thickness cannot be made because the bottom of the unit is cut out of magnetite-bearing granitoids of the Ordovician Oliverian magma series. Intrusive rocks of the Oliverian series occupy the cores of domal uplifts and are mantled by Ammonoosuc volcanics and younger rocks. An Ordovician magmatic arc composed of Ammonoosuc volcanics and Oliverian plutons is known to extend parallel to regional strike trends from Connecticut, across Massachusetts, along the western edge of New Hampshire, and finally curving eastward into Maine. The arc is now exposed within a Devonian structure, the Bronson Hill anticlinorium.

Rocks of the Ordovician magmatic arc are overlain unconformably by (1) Silurian conglomerates and quartzites of the Clough Quartzite; (2) marbles and calc-silicates of the Silurian Fitch Formation; and (3) pelitic mica schists and micaceous quartzites of the Lower Devonian Littleton Formation. The thicknesses of the Clough and Fitch are highly variable: in some areas the units are absent because of lack of deposition or boudinage; in other areas, as at Black Mountain, the exposed outcrop breadth of the Clough is hundreds of meters, however, some of this apparent thickness may be due to repetition by isoclinal folding. The Littleton Formation may be several kilometers thick but its top is not exposed owing to uplift and erosion. Sections of the Littleton show well developed graded beds, 10-20 cm thick; in other outcrops only massive mica schists are seen. The Littleton is intruded by foliated sill-like plutons of ilmenite-bearing granitoids, the Bethlehem Gneiss and Kinsman Quartz Monzonite of the New Hampshire series. The plutons show neither a contact metamorphic aureole nor chill zones. These features, combined with the presence of primary igneous muscovite, suggest emplacement during active regional metamorphism at depths of 10 km, or more.

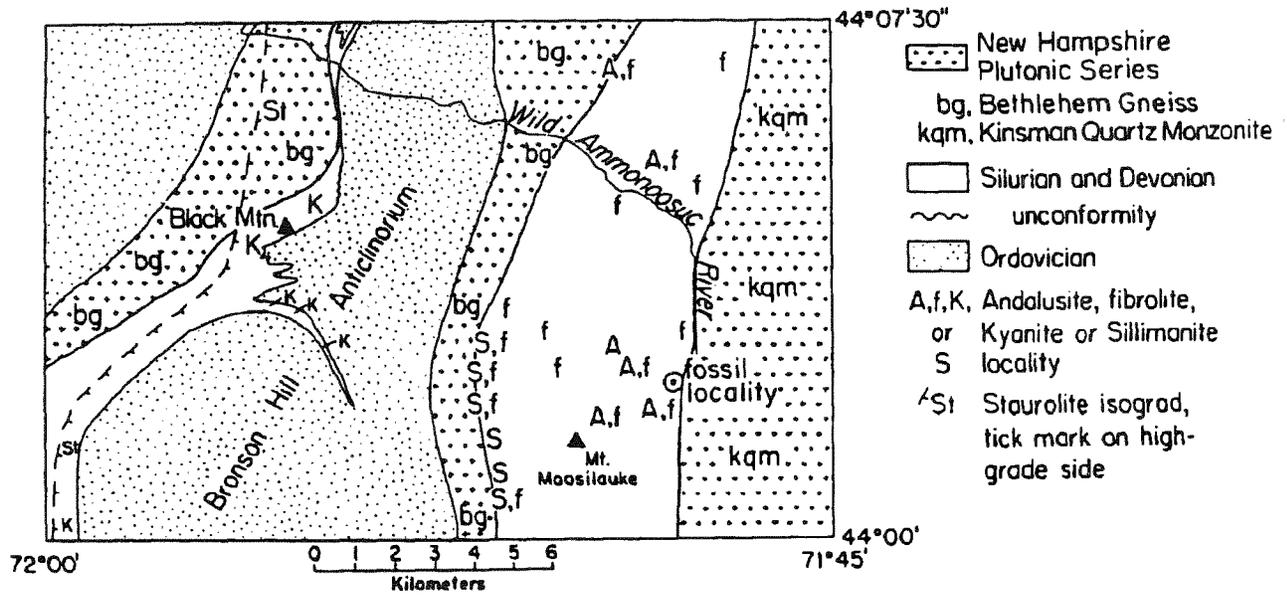


FIG. 9 Location of Black Mountain, Beaver Brook fossil locality, and distribution of  $Al_2SiO_5$  polymorphs in southern half of Mt. Moosilauke 15' quadrangle. (Rumble et al., 1982)

The age of regional metamorphism and deformation is constrained by the presence of unaffected Mesozoic intrusives and volcanics of the White Mountain Magma series in nearby Franconia Notch. Metamorphism and deformation were traditionally assigned to the Devonian, an age confirmed by the Rb-Sr dating of the syn-metamorphic Bethlehem and Kinsman at 405–415 m.y. Deformation and metamorphism were synchronous during early stages of tectonism: porphyroblasts grew and the preferred orientation of schistosity was established during emplacement of large nappes. Later, metamorphic intensity waned but folding continued as the Bronson Hill anticlinorium and its component domes were built. The sense of transport during the nappe episode was from E to W; in consequence, hotter rocks were stacked on top of cooler rocks. The domes of the Bronson Hill not only refold the nappes but also deform isograds.

#### Access

Black Mtn. may be reached by following New Hampshire route 25 to the village of East Haverhill (about 5 miles E of the intersection of route 10 and route 25). A short detour, a few hundred meters E. of the village on rt. 25, reveals an impressive view of Black Mtn., Sugarloaf Mtn., and the Hogsback, all composed of white Clough quartzite. The arcuate quartzite cliffs of the Hogsback delineate the northerly plunge of the Owls Head dome. The thick accumulations of quartzite at the Sugarloaf and Black Mtn. illustrate the effects of isoclinal folding in the nappes. Returning to East Haverhill village, turn N. and follow Lime Kiln Road to Lime Kiln Camps. The trip will follow Chippewa Trail to the top of Black Mtn. The Lime Kiln Camps are now privately owned. Parking for Chippewa Trail is at the first turn-out south of the entrance to the Camps on Lime Kiln Road.

#### Hiking Route

The group will proceed directly to the summit of Black Mtn. We will begin detailed study of outcrops at the summit and then retrace our steps slowly downward so that we can stay together for good discussions. The rock types that will be seen are conglomerates, quartzites, and quartz-mica schists of the Clough Quartzite.

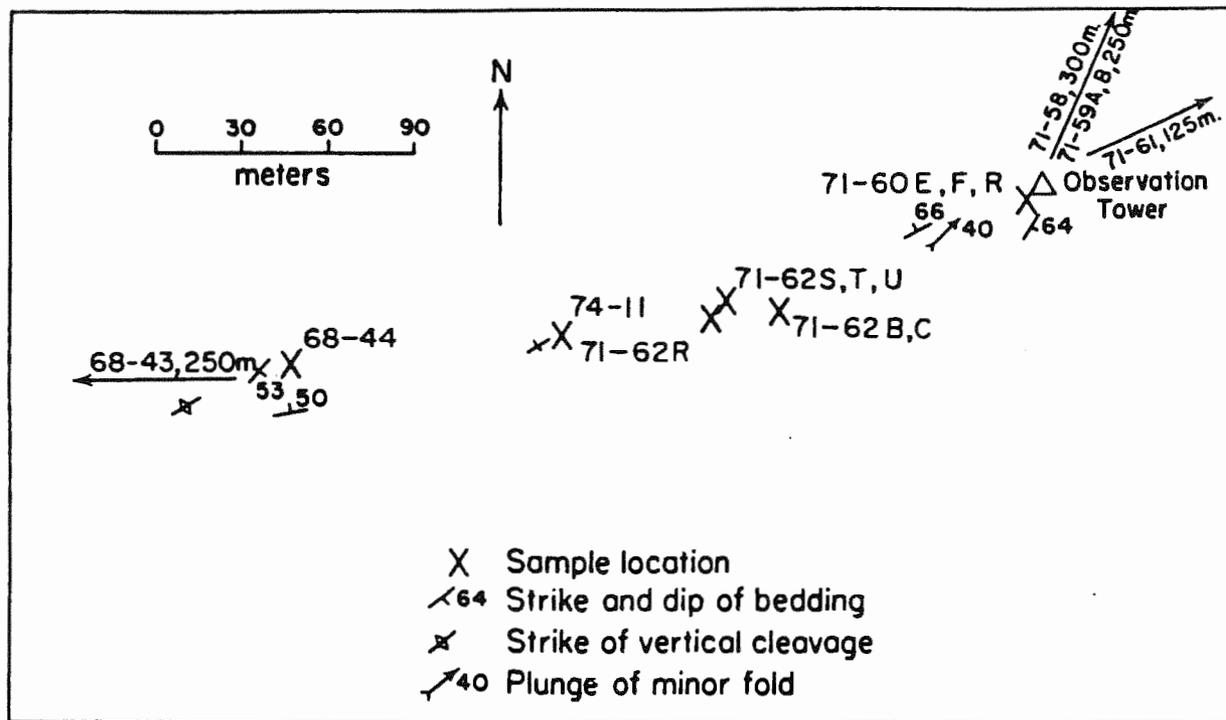


FIG. 10 Location of samples collected along summit ridge of Black Mtn. Field trip route begins at site of observation tower and proceeds SW, towards 68-44 (Rumble, 1978).

The summit was formerly marked by an observation tower (Fig. 10). One can still find the location of its concrete foundation. Facing S. from the site of the tower, the back sides of Hogsback and Sugarloaf are visible. The northerly plunge of the Owls Head Dome beneath Black Mtn. can be easily visualized. The structure to the N. is not well displayed but mapping shows a southerly plunging dome. Thus, Black Mtn. lies in a structural saddle formed by a plunge depression between two domes. Other summits underlain by Clough Quartzite stand to the S. including Piermont Mtn., Mt. Cube, and Croydon Peak. They trace the continuation of the Bronson Hill anticlinorium.

Oxygen isotope geothermometry of Black Mtn. rocks (quartz-magnetite) gives a temperature estimate of 495°C ( $\pm 10^\circ\text{C}$ ). Pressure is estimated to lie within 0.5 kbars of the  $\text{Al}_2\text{SiO}_5$  triple point, i.e. 3.8 kbars according to Holdaway.

In ledges a few meters S. of the tower, there is abundant kyanite occurring as felted layers defining schistosity. Note how kyanite has been folded. The minor folds that fold kyanite are parasitic to the domes.

Turn back N. and look down at the tower foundations. Here you will find chloritoid-staurolite-chlorite assemblages of 71-60E and extremely Fe-rich assemblages with garnet (71-60F, 71-60R). The Fe-rich, garnet-bearing assemblages have reduced oxides with  $\text{FeTiO}_3$  in ilmenite of 95%, or more. (Data on mineral assemblages summarized in figures 11 through 18).

Walk just to the E., beyond the tower site and you will find tourmaline-filled joints. Careful searching may find beds of staurolite-chloritoid-garnet assemblages that have been almost completely replaced by tourmaline, within a few cm of the joints.

Before beginning the descent, look SE. Mt. Moosilauke dominates the skyline. Beaver Brook, the second stop of today's trip, lies on the E. slopes of Mt. Moosilauke. Mica schists on the slopes of the mountain contain abundant andalusite and sillimanite. The site of the  $\text{Al}_2\text{SiO}_5$  triple point is between where we are standing on Black Mtn. and Mt. Moosilauke.

Turn towards the SW and begin to descend towards the cars. Remember to keep along the rocky spine of the ridge in order to see the best outcrops. The first steep scramble downward, just a few 10's of meters from the tower site, leads past abundant dome-stage minor folds and kyanite schists. The kyanite schists typically have the assemblage quartz-muscovite-kyanite-ilmenohematite-magnetite: they are highly oxidized.

Beyond the steep scramble one reaches a knob on the ridge. Just S. of the knob is a gently-dipping, tourmaline-coated joint. The knob is the site of a chlorite schist with staurolite-chloritoid-chlorite-magnetite-ilmenite ( $\text{FeTiO}_3 = 98\%$ ). Adjacent to the chlorite schist are outcrops of kyanite-staurolite-chloritoid-chlorite-magnetite-ilmenohematite ( $\text{FeTiO}_3 = 28\%$ ) schist (71-62B, 71-62S, 71-62T, 71-62U). Here, separated by a meter or so, are rocks whose fluids differed by 20-30% in  $X_{\text{H}_2\text{O}}$  and by the same amount (antipathetically) in  $X_{\text{H}_2}$ . Oxygen isotope analyses of co-existing quartz and magnetite show the minerals to have the same partitioning of  $^{18}\text{O}/^{16}\text{O}$  but the quartz and magnetite of 71-62U and 71-62B are 1‰ enriched in  $\delta^{18}\text{O}$  relative to 71-62R. Thus, inferences of fluid inhomogeneity based on mineral assemblages are supported by oxygen isotope analyses.

Turn SW from the knob and continue to descend, following blazed trail through the scrub to avoid a steep place on the ridge. Shortly, the trail emerges on broad, open bare-rock ledges. The kyanite-chloritoid-staurolite-paragonite quartzite of 68-44 is located in low ledges N of the trail. In thin sections it may be seen that kyanite, chloritoid, and staurolite are intergrown epitaxially, presumably along the chains of aluminum octahedra they all share.

Turn SW and continue to follow the blazed trail through the ledges of quartzite. Be on the lookout for current bedding, tourmaline-covered joints, and quartz-mica schist with very  $\text{Fe}^{2+}$ -rich garnets. The garnets may be readily recognized because of their tendency towards rusty weathering.

Return to bus for the drive to Beaver Brook.

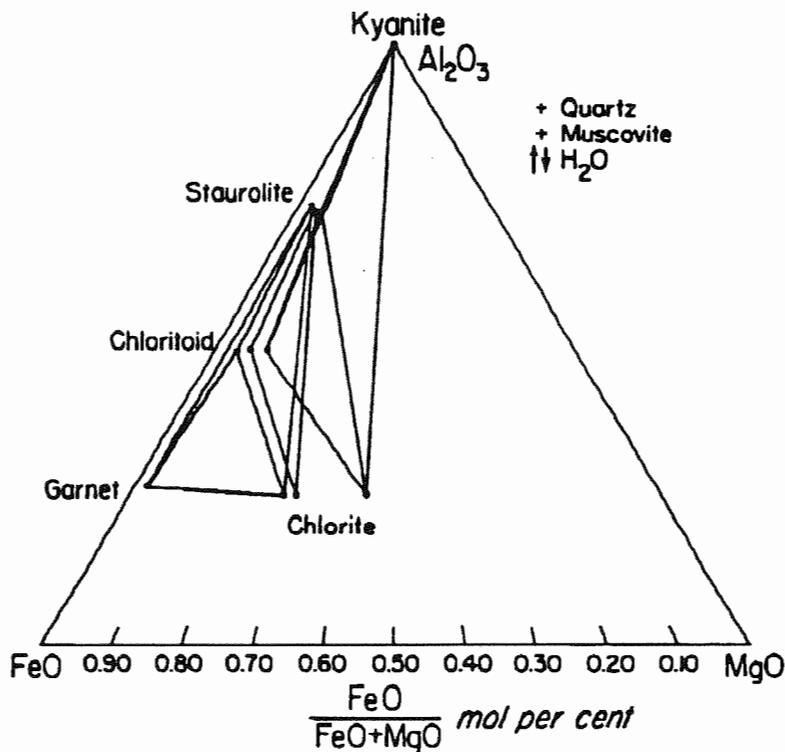


FIG. 11 Thompson projection of kyanite, staurolite, chloritoid, garnet, and chlorite compositions (ubiquitous minerals quartz and muscovite; also, projection through  $\text{H}_2\text{O}$ ) (Rumble, 1978).

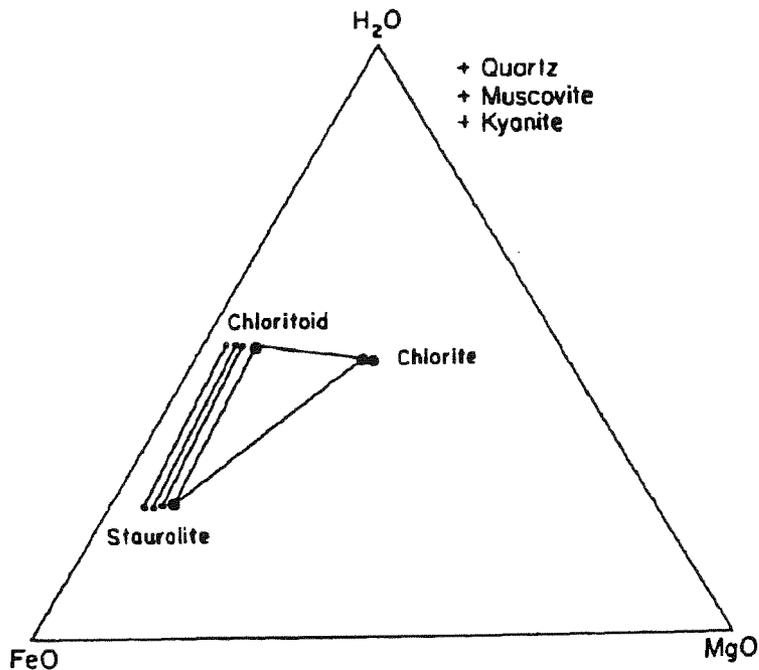


FIG. 12 Projection of chloritoid-chlorite-staurolite compositions from the ubiquitous minerals quartz-muscovite-kyanite. (Rumble, 1978)

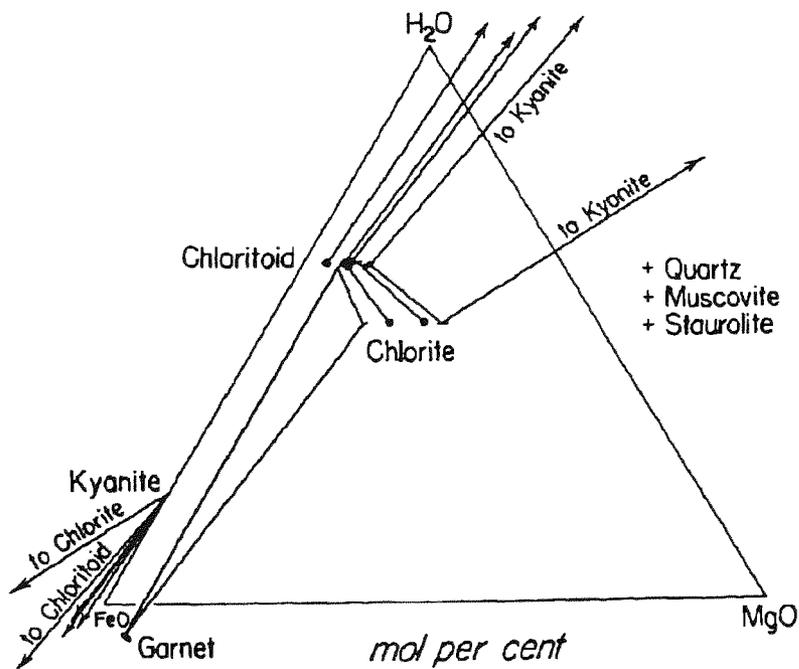


FIG. 13 Projection of chloritoid-chlorite-garnet-kyanite from the ubiquitous minerals quartz-muscovite-staurolite. (Rumble, 1978)

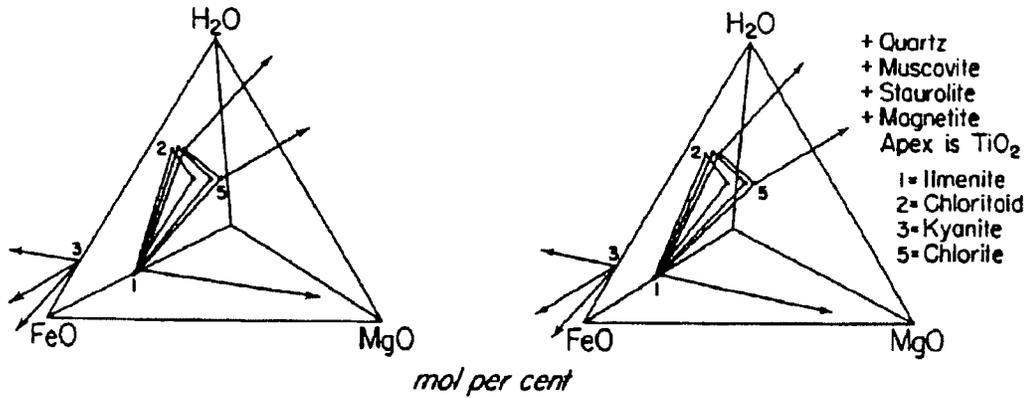


FIG. 14 Stereographic pair of tetrahedra showing projection of ilmenite-chloritoid-chlorite-kyanite from ubiquitous minerals quartz-muscovite-staurolite-magnetite. Apex is  $TiO_2$ . (Rumble, 1978)

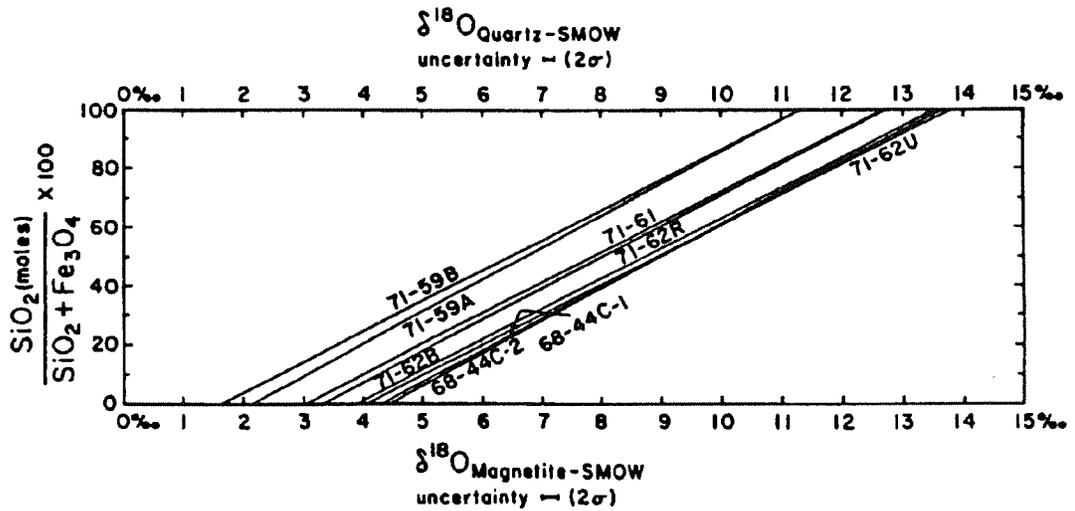


FIG. 15 Oxygen isotope exchange equilibrium diagram for coexisting quartz and magnetite.

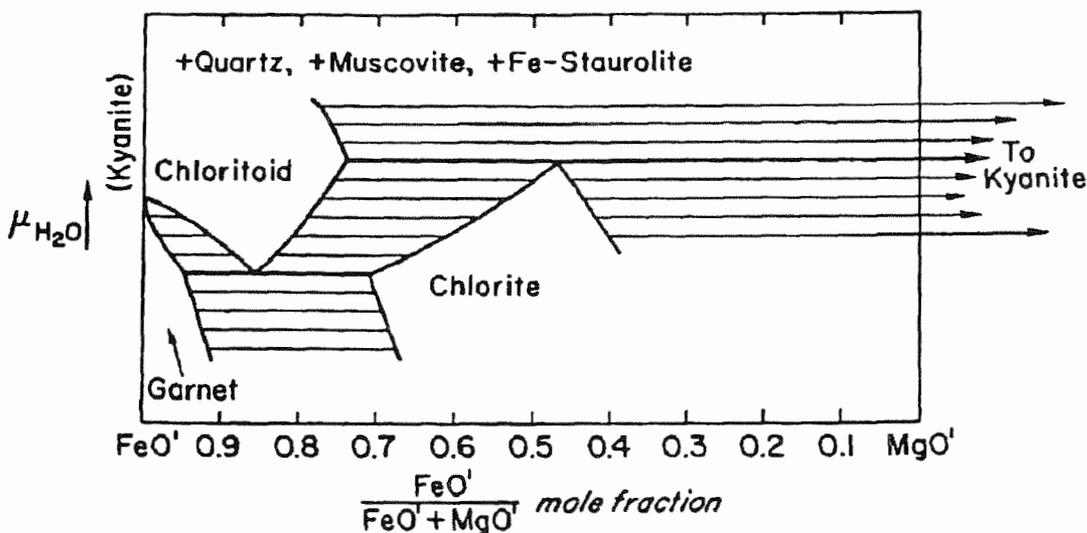


FIG. 16 Isothermal, isobaric variation of  $\mu_{H_2O}$  with mineral assemblage and mineral composition. (cf. Fig. 13).

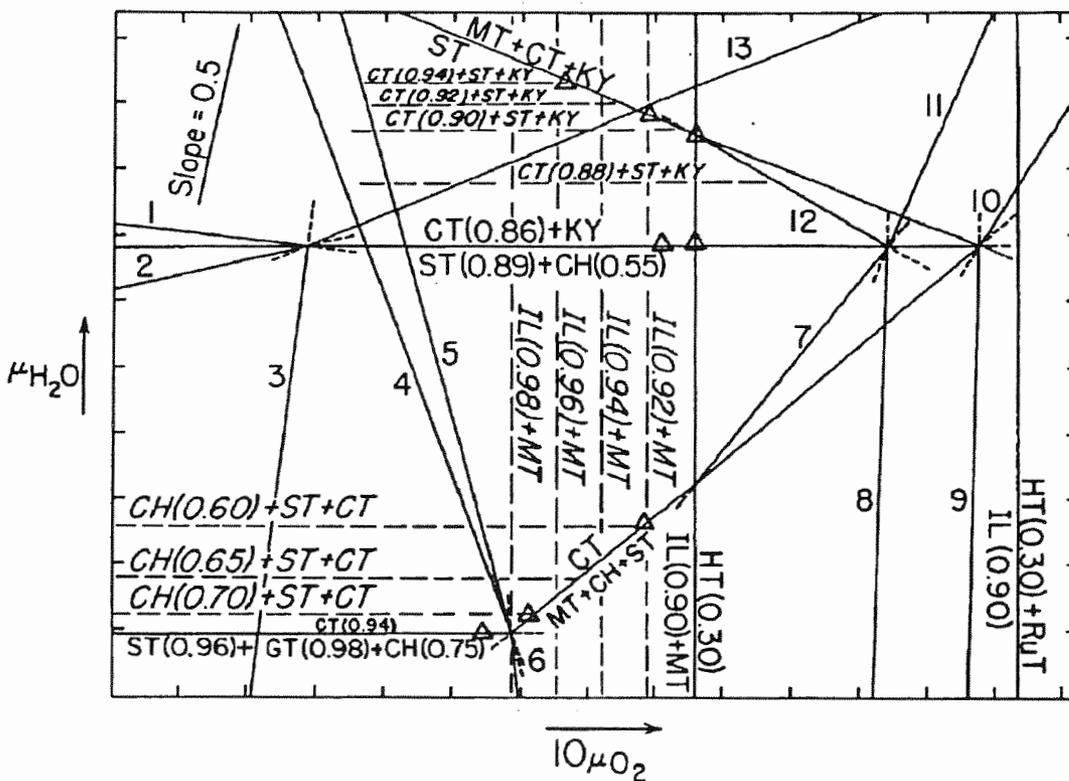


FIG. 17 Variation of  $\mu_{H_2O}$  and  $\mu_{O_2}$  with mineral assemblage and mineral composition. Triangles show relative positions of observed assemblages.

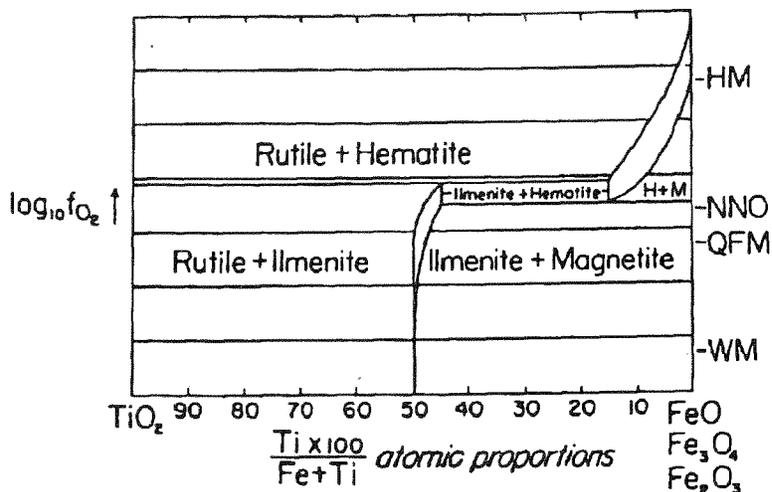


FIG. 18 Isothermal, isobaric variation of  $fO_2$  with mineral assemblage and mineral composition.

### BEAVER BROOK FOSSIL LOCALITY

#### Access

Drive along New Hampshire rt. 112 to a point approximately 0.75 km W. of Lost River. Here there are signs marking the Appalachian Trail. Park cars under the trees, on the S. side of the road, near Appalachian Trail signs. The Beaver Brook fossil locality may be reached by following the white-blazed, Appalachian trail S., across the floor of Kinsman Notch, and climbing 0.5 km up the lower slopes of Mt. Moosilauke.

#### Hiking Route

**WARNING: CASCADES OF BEAVER BROOK ARE VERY SLIPPERY. EXERCISE EXTREME CAUTION!**

Hike S. on the trail (white blazes) for about 0.5 km to the first cascades of Beaver Brook. At your feet are large feldspar phenocrysts (up to 10 cm) in the Kinsman Quartz Monzonite. Visible in the cascades are mica schists of the Littleton Formation cross-cut by aplite dikes. A knife-sharp contact between Kinsman and Littleton may be found at the base of the ledge, down in the brook. (CAUTION: SLIPPERY WHEN WET!)

Continue following the white-blazed trail climbing alongside the cascades. Do not attempt to climb straight up the cascades because the rocks are very slippery. Over the course of the next 0.5 km you will notice that the Littleton mica schists are not very aluminous. There are some muscovite-garnet schists but most rocks lack muscovite, have biotite with intermediate  $Fe/(Fe + Mg)$ , and contain calcic-plagioclase. In at least one case (sample 70-184-N) are two coexisting plagioclases: An 83 and An 67. The fossiliferous calc-silicate is usually interbedded with calcite mica schists rather than with aluminous pelites.

There are no obvious landmarks to help locate the fossil locality. It will be marked for the trip so you can't miss it. We will have to cross the Brook, then climb a gently sloping ledge to the base of a cascade. **USE EXTREME CAUTION: ROCKS ARE VERY SLIPPERY!**

Fluid-rock behavior at Beaver Brook was dramatically different from that at Black Mountain. You will be able to see two different types of evidence for fluid infiltration with your own naked eyes. Still other evidence for infiltration is given in the accompanying figures and will be discussed on the outcrop. Take a moment for orientation and note the color contrasts between pink-orange and green banded calc-silicates, dark gray mica schists,

and pale gray-dirty white dikes of aplite and quartz monzonite. Thermobarometry for these rocks gives metamorphic conditions of 3.5 kbars and 600°C (Fig. 19).

There are two pieces of evidence for infiltration that may be seen in the ledge before you. These are (1) a sharp, infiltration metasomatic front of anorthite in the margin of the quartz monzonite dike, on the left side of the ledge; and (2) fossil brachiopods replaced by wollastonite, on the right side.

A close inspection of the quartz monzonite dike shows a creamy, translucent zone (2-4 cm thick) in the dike margin where it truncates calc-silicate beds. The zone is absent where the dike crosscuts mica schist. Microscopy shows that the marginal zone consists of myrmekitic anorthite (An 95-98) - quartz pseudomorphs after the quartz monzonite's primary oligoclase and microcline. A microprobe traverse across the replacement front reveals a change in plagioclase composition from An 95 to An 19 over a distance of 1-2 mm. The anorthite zone has features diagnostic of infiltration metasomatism (as discussed by Korzhinskii and Hofmann) including (1) one-way mass transfer (i.e. Ca from calc-silicate to dike rock) and (2) a sharp replacement front separating mineral assemblages that are not in chemical equilibrium. The anorthite zone documents a component of the direction of fluid flow. Calcic aqueous solutions flowed from calc-silicate wall rocks into the quartz monzonite dike.

Fossil brachiopods replaced by wollastonite are located on the right side of the ledge. The original discovery of fossils by (C.V. Guidotti) was made on an exposed portion of the ledge where weathering had etched out their distinctive forms. Weathered material has since been quarried away for examination by A. J. Boucot who identified four genera of Early Devonian brachiopods including *Acrospirifer*, *Leptocoelia*, *Atrypa*, and either *Leptostrophia* or *Protoleptostrophia*. The present outcrop surface has not as yet weathered very deeply. You should be able to see, however, the characteristic cross-section of brachiopod shells, e.g. a shape like a flattened parenthesis.

The presence of the assemblage wollastonite-quartz-calcite in the fossil shells is evidence of fluid infiltration during metamorphism. Fluid/rock ratios may be calculated with the model of J. M. Ferry. Consider the decarbonation reaction  $\text{Quartz} + \text{Calcite} = \text{Wollastonite} + \text{CO}_2$ . At the conditions of metamorphism at Beaver Brook, 3.5 kbars and 600°C, the equilibrium  $X_{\text{CO}_2}$  for this reaction is 0.09. But, the reaction produces pure  $\text{CO}_2$ . The decarbonation reaction would saturate a rock of normal porosity with  $\text{CO}_2$  after only a minuscule amount of reaction. Thus, reaction would cease after formation of an undetectable amount of wollastonite. We are faced with a problem: How does one make wollastonite in detectable amounts at temperatures well below its stability limit in pure  $\text{CO}_2$ ? The answer is to infiltrate the rock with a fluid whose  $X_{\text{CO}_2}$  is less than the equilibrium value. The infiltrating fluid sweeps away product  $\text{CO}_2$  and makes it possible for decarbonation reactions to operate until reactants are exhausted. Calculated fluid/rock ratios at Beaver Brook range from 1.5 to 5.0, with a value of 4.0 (by volume) in the wollastonite bed.

Additional evidence of infiltration include possible hydrothermal graphite in the quartz monzonite dike and changes in  $\delta^{18}\text{O}$  inferred to have occurred during metamorphism. These features cannot be seen in outcrop but will be discussed with the aid of the accompanying figures (Figs. 20 through 21).

Why did Beaver Brook experience infiltration during metamorphism while Black Mountain did not?

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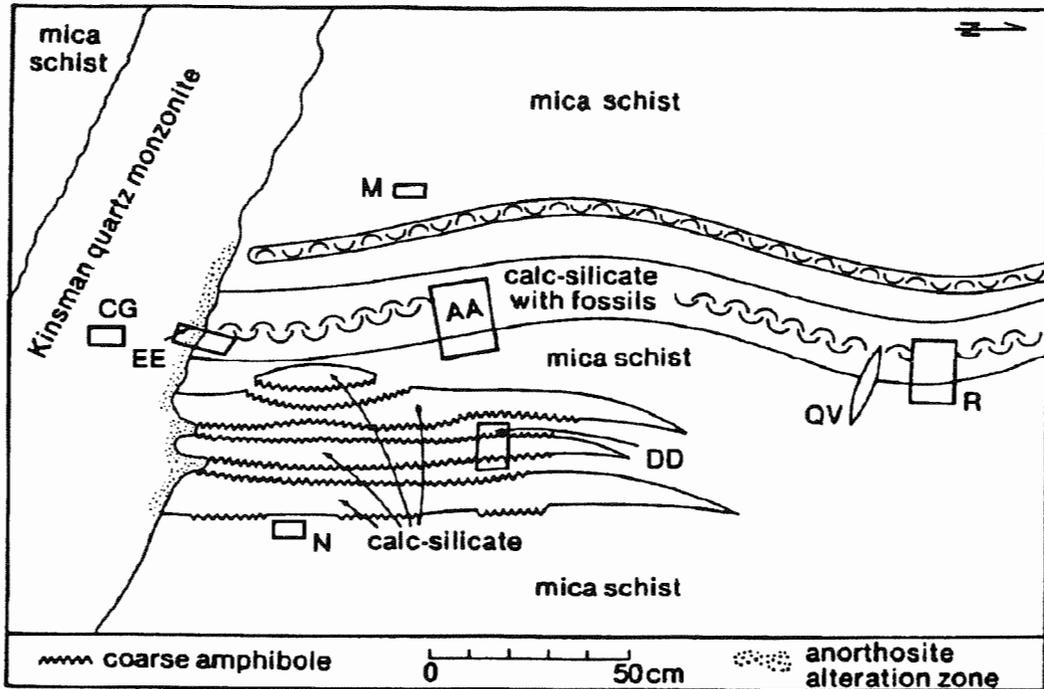


FIG. 19 Outcrop map of Beaver Bk. fossil locality. Sample locations given by letter symbols. (Rumble, et al., 1982)

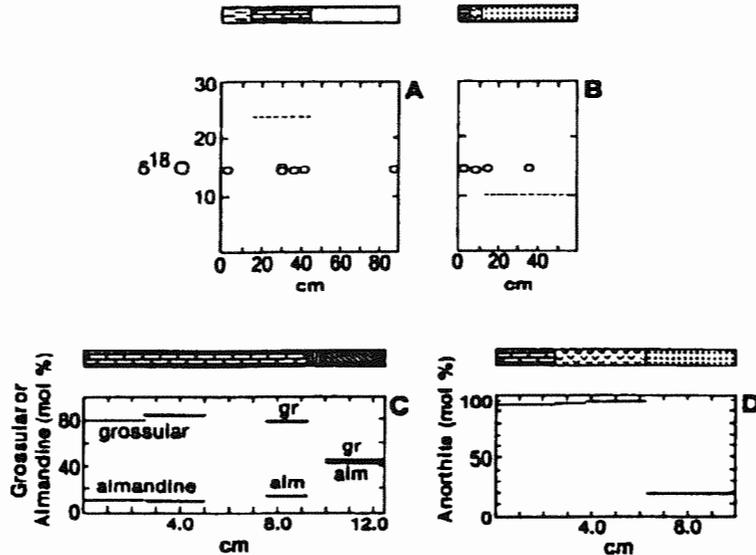


FIG. 20 (A) shows  $\delta^{18}\text{O}$  values vs. distance perpendicular to bedding for traverses AA and R, combined, and samples M and N. Circles show measured data. Dashed line gives inferred pre-metamorphic isotopic composition. Dashed pattern denotes mica schist, brick work is calc-silicate. (B) is a plot of  $\delta^{18}\text{O}$  vs. distance perpendicular to dike/country rock contact for traverse EE and CG. The "V" pattern is anorthite replacement zone, the "+" pattern is Kinsman Quartz Monzonite. (C) shows garnet composition vs. distance perpendicular to bedding for traverse AA. Vertical ruled pattern is amphibolite. (D) is a plot of plagioclase composition vs. distance perpendicular to the dike/country rock contact for traverse EE and CG. (Rumble, et al., 1982)

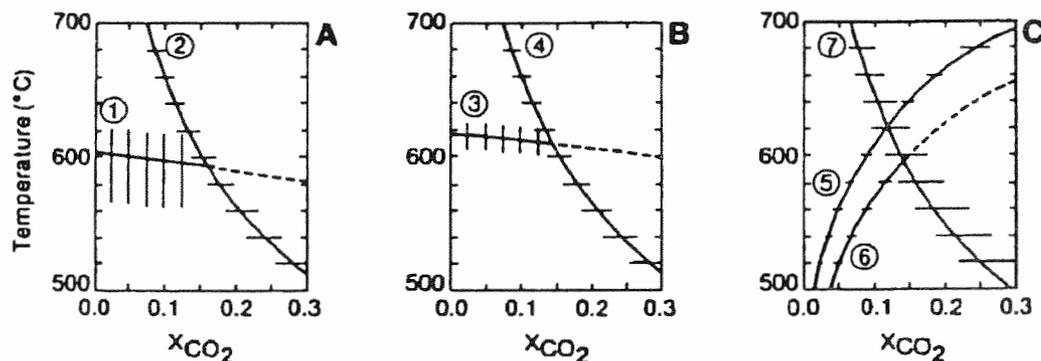


FIG. 21 Thermobarometry on T-  $X_{CO_2}$  diagrams at 3.5 kbars. Short horizontal and vertical bars give uncertainties in location of equilibrium curves for a range of  $2\sigma$  about average values of equilibrium constants. In (A) equilibrium 1 is  $Amp + Czo + 2Q = 5 Cpx + 9 Pl + 4H_2O$ , (samples AA, R, DD, and EE) and 2 is  $Czo + CO_2 = 3 Pl + Cc + H_2O$ . (B) shows equilibrium 3,  $4 Czo + Q = 5 Pl + Gt + 2H_2O$  (sample AA) and equilibrium 4,  $Czo + CO_2 = 3Pl + Cc + H_2O$ . In (C) equilibrium 5 is  $Cc + Q = Wo + CO_2$  (sample R) equilibrium 6 is  $2 Czo + 5 Cc + 5 Cc + 3Q = 3Gt + 5 CO_2 + H_2O$ . Dashed lines are metastable extensions. (Amp = amphibole, Cco = clinozoisite, Q = quartz, Cpx = clinopyroxene, Pl = plagioclase, Cc = calcite, Gt = garnet, Wo = wollastonite).

### TOWNSEND DAM GARNET LOCALITY

by

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**Stop X Townshend Dam:** Uninterrupted exposures of schists and amphibolites crop out along the approximately 400 meter section of outcrop in the spillway and highway (Rt. 30). These schists contain spectacular rolled garnets that have been the subject of detailed structural (Rosenfeld, 1968; 1970), strontium isotopic (Christensen et al., 1989); and oxygen isotopic (Chamberlain and Conrad, 1993) studies. These studies have shown that the garnets at Townshend Dam grew over an approximately 10 Ma interval during Acadian-aged thrusting and nappe emplacement and during a period of intense hydrothermal activity.

The rocks at Townshend Dam belong to four stratigraphic units. At the far eastern edge of the spillway are well banded quartz-albite gneisses belonging to the Hoosac Formation (Precambrian-Cambrian). Immediately west of the Hoosac are paragonite-chlorite-garnet schists interbedded with epidote amphibolites belonging to the Pinney Hollow Formation (Cambrian). These schists are overlain by a rusty weathered, chlorite-garnet schist with interbedded amphibolites, collectively called the Ottaquechee Formation. At the northwest end of the spillway are rocks assigned to the Moretown member of the Missisquoi Fm. (Ordovician). These rocks are gray weathered, muscovite-chlorite-epidote schists with characteristic biotite-chlorite-epidote pseudomorphs of hornblende.

The garnets studied by Christensen et al. (1989) and Chamberlain and Conrad (1993) were collected from the Pinney Hollow Formation that crops out at the far eastern edge of the spillway. The duration of garnet growth was determined using Rb-Sr isotope systematics of garnet (Christensen et al.; 1989). These authors analyzed core and rim segments of the garnets and the schist matrix. From these data, they concluded that the garnets grew over an average time interval of  $10.5 Ma \pm 4.2$  myrs. The paper by Christensen et al. (1989) marks an important advance in our understanding of tectonometamorphic processes, for they were the first to show that rates of metamorphism could be directly determined from isotopic zoning in garnet. However, the utility of their method rests largely on the assumption that the evolution of  $^{87}Sr/^{86}Sr$  ratio of the matrix during garnet growth is due solely to decay of  $^{87}Rb$

and is not influenced by the transport of Sr by infiltrating fluids. It is, therefore, of interest to determine whether the outcrops studied by Christensen et al. (1989) were infiltrated by fluids during metamorphism and, if so, whether the Rb-Sr systematics were affected by the infiltration.

With this in mind, we used the CO<sub>2</sub> laser extraction system at Dartmouth to determine if the garnets were zoned in oxygen isotopes. Our studies show that the the garnets are strongly zoned in oxygen isotopes (Fig. 22). The magnitude and nature of  $\delta^{18}\text{O}$  zoning depends upon the garnet's location in the outcrop studied. The garnets examined in this study come from the isotopically low  $\delta^{18}\text{O}$  schists of the Pinney Hollow Fm. ( $\delta^{18}\text{O}$  whole rock  $\sim 9\text{‰}$ ) which is adjacent to the relatively high  $\delta^{18}\text{O}$  schists of the Ottaquechee Fm. ( $\delta^{18}\text{O}$  whole rock  $\sim 12.5\text{‰}$ ). Garnets from the Pinney Hollow within 10 meters of the contact with the Ottaquechee have relatively homogeneous  $\delta^{18}\text{O}$  values varying from 9.5‰ in the core to 10.5‰ at the rim. Garnet in the Pinney Hollow 85 meters from the contact are more strongly zoned, with  $\delta^{18}\text{O}$  ranging from a low of  $\sim 6.0\text{‰}$  in the cores to a high of  $\sim 9.0\text{‰}$  at the rims of the garnet. These zoning patterns were produced by continuous infiltration of relatively high  $\delta^{18}\text{O}$  waters derived from the subjacent schists into the paragonitic schists during garnet grade metamorphism.

It is possible to determine the time-integrated fluid fluxes by comparison of observed  $\delta^{18}\text{O}$  zoning profiles in garnet with those calculated from the equation describing combined advection-diffusion of a tracer. Using this method, we calculate time-integrated fluid fluxes of  $\sim 1.5 \times 10^4 \text{ cm}^3/\text{cm}^2$ . These calculated fluid fluxes have several interesting implications. First, using these fluid fluxes we estimate that Sr was transported 1.5 cm to 1.5 m during fluid flow. Therefore, the assumption that the matrix is closed to Sr during metamorphism (see Christensen et al., 1989) is a reasonable approximation for most of the rocks at Townshend Dam. Second, the fluid fluxes that we calculate are consistent with the fluxes that Ferry (1992) determined from petrologic studies of rocks adjacent the Chester and Athens gneiss domes. Ferry (1992) proposed that the gneiss domes were centers of regional-scale hydrothermal systems during the Acadian metamorphism. The garnet isotopic data, in part, support this hypothesis and suggest that flow was directed toward structurally deeper rocks now exposed in the cores of the gneiss domes.

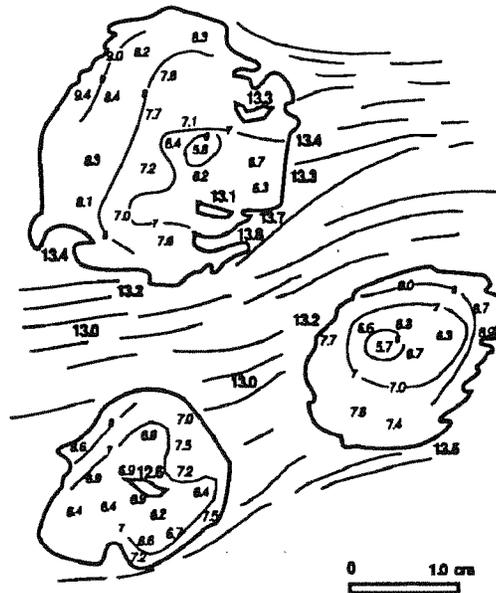


FIG. 22 Oxygen isotope values of garnet and quartz. Smaller numbers are  $\delta^{18}\text{O}$  in garnet (note contours); larger numbers give analyses of quartz.

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## ROAD LOG

- 1.9 miles West of the village of Townshend on route 30. Park cars in the parking area on left at Townshend Control Dam. The rocks that we will be examining are exposed in the spillway below the dam.

## INVERTED METAMORPHISM, P-T PATHS AND TECTONIC HISTORY OF WEST-CENTRAL NEW HAMPSHIRE

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## INTRODUCTION

The existence of an isolated area of high-grade metamorphic schists on Fall Mountain, New Hampshire was noted by Kruger (1946). This, in addition to stratigraphic and structural considerations, led Thompson et al. (1968) to propose the existence of the Fall Mountain nappe, which was interpreted to have carried hot rocks of the Merrimack synclinorium westward over the Bronson Hill anticlinorium to their present position on Fall Mountain.

The purpose of this field trip is to examine the interrelationship between metamorphism and structural level in a traverse across the Bronson Hill anticlinorium that includes Fall Mountain. The trip begins in the Connecticut Valley metamorphic low along the Chicken Yard line and proceeds eastward up metamorphic grade and up structural level (Fig. 23). Details of the petrology of rocks to be examined on this trip are found in Spear (1993) and Spear et al. (1990).

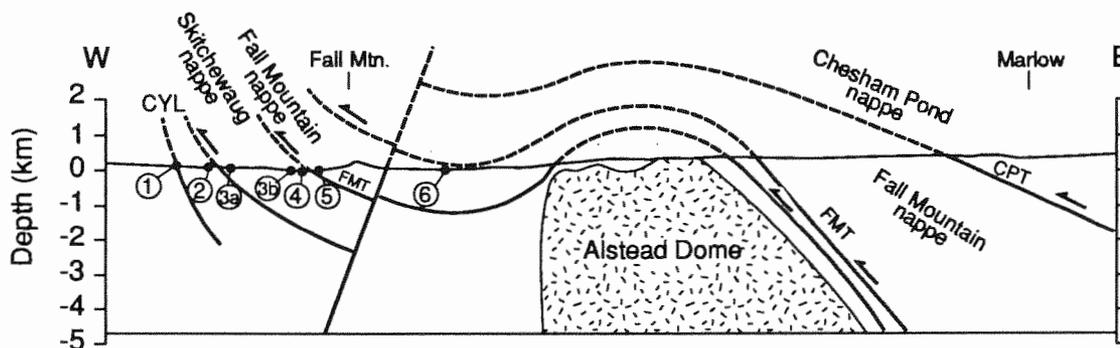


FIG. 23. Schematic cross section (adapted and modified from Thompson et al, 1968) from the Chicken Yard Line (CYL) eastward along the field trip route. CPT = Chesham Pond thrust; FMT = Fall Mountain thrust. Numbered dots show structural position of field trip stops.

## ROAD LOG

Assemble in Bellows Falls, Vermont at the intersections of Routes 5 and 121 in the parking lot of Athens Pizza and Family Restaurant (formerly a grocery store).

## Mileage

- 0.0 Parking lot of Athens Pizza and Family Restaurant. Exit parking lot and turn right (west) onto Route 121. Proceed to stop light (50 meters). Go straight through light up hill (west) on Route 121.
- 1.5 Cross Saxtons River.
- 1.7 Pass under I-91.
- 1.95 Pass road on left.
- 2.3 Westminster town line.
- 2.7 Turn right (north) onto bridge that crosses Saxtons River.
- 2.75 Turn left (west) at "T" junction.
- 2.85 Park on left or right as room permits. Warning – shoulder on right (south) is soft in places.

**Warning – poison ivy.**

**Stop 1. Chicken yard line – Chlorite zone.** (30 minutes). (Bellows Falls 15' quadrangle; Bellows Falls 71/2 minute quadrangle; Sample location BF-55 and BF-56). The purpose of this stop is to examine the low grade rocks that crop out along the Chicken Yard line and to discuss the possible significance of the "Connecticut Valley metamorphic low".

Rocks on the west end of the series of outcrops are Gile Mountain Formation (gray, well laminated with sandy layers on the order of a few mm to a cm thick separated by shaly layers). The intermediate and easternmost outcrops are Littleton Formation (gray shaly appearance without pin striping). The Littleton here is highly sheared and in thin section is characterized by abundant mica fish and small scale shear zones. In other words, this is a mylonite.

The metamorphic grade here is biotite zone. Typical assemblages include **chlorite + muscovite + quartz**. To the west the grade increases to garnet zone at the town of Saxtons River and to the east the garnet zone is encountered in 0.25 km (field trip stop 2). The metamorphic evolution is not symmetrical, however. Metamorphic pressures in the Vermont sequence are generally higher (6-8 kbar) than those in the Bronson Hill anticlinorium in New Hampshire (4-6 kbar). Moreover, P-T paths of rocks in Vermont are generally clockwise whereas P-T paths from New Hampshire rocks are generally counterclockwise.

- 2.85 Continue straight (west).
- 3.1 Outcrops on right (north) are well laminated Gile Mountain formation.
- 3.2 "T" junction. Go left.
- 3.3 Covered bridge. Cross Saxtons River and return to route 121.
- 3.35 Route 121. Turn left (east).
- 3.9 Park on left of road in pull out. Outcrops on south side of road.

**Stop 2. Garnet zone.** (20 minutes). (Bellows Falls 15' quadrangle; Bellows Falls 71/2 minute quadrangle. Sample location BF-53). The purpose of this stop is to examine rocks of the garnet zone on the east flank of the Connecticut Valley metamorphic low.

The rocks here are well bedded, graphitic schists of the Littleton Formation. Small garnets are abundant in some layers. The assemblage is **garnet + biotite + chlorite + muscovite + quartz**. Note that we are only 0.25 km east of stop 1 and the grade appears to be higher.

- 3.9 Continue east on Route 121.
- 4.6 Old Westminster Road on right (south). Continue on 121.
- 4.8 Pass under I-91.
- 5.1 Cross Saxtons River.
- 5.6 Turn right at circle with flagpole
- 5.7 Fork in road. Right fork crosses Saxtons River.
- There is no detailed road mileage to stop 3a. Turn right and cross Saxtons River. Continue south to the intersection with Route 5 (approximately 2 miles). Go south on RT 5 approximately 3.5

miles to where a small road passes under the interstate to the west (right). Turn right and go under the interstate. Stop at first outcrop on the right (100 yards).

**Stop 3a. Staurolite zone.** (20 minutes). (Bellows Falls 15' quadrangle; Walpole 71/2 minute quadrangle.) The purpose of this stop is to examine staurolite zone rocks that are structurally low in the nappe sequence.

The rocks here are in the Littleton Formation in the Skitchewaug nappe and the most obvious feature is the large staurolite crystals. The staurolites are late in the paragenetic sequence and overgrow early foliations as well as include garnets. The staurolite producing reaction is  $\text{garnet} + \text{chlorite} + \text{muscovite} = \text{staurolite} + \text{biotite} + \text{H}_2\text{O}$ .

Retrace steps to Saxtons River and pick up road log.

- 5.7 (0) Cross Saxtons River and turn right.  
 5.9 (0.2) Pull off onto side of road on left (west) and park. Walk down path to east to river.  
**Warning—Poison Ivy.**

**Stop 3b. Staurolite zone.** (40 minutes). (Bellows Falls 15' quadrangle; Walpole 71/2 minute quadrangle. Sample location 79-66). The purpose of this stop is to examine rocks of the staurolite zone that are structurally high in the Skitchewaug nappe. (Also see Thompson and Rosenfeld, 1979, Stop 1).

Rocks of this outcrop are the Littleton Formation displaying excellent graded beds. Tops can readily be determined by examining direction of fining (note that the large crystals have grown in the clay layers at the tops of the beds). The rocks here are isoclinally folded, which reverses the direction of tops. A particularly good example of an isoclinal fold hinge can be seen in the rocks immediately to the right (upstream) of the main pool. Stand facing the pool and note the topping direction, which is inverted to the west. Now note the topping direction of rocks to the right of the pool at head level (2 meters above the water). Here the rocks top up to the east. The fold hinge is exposed at the pool's edge and plunges into the pool.

Tension gashes filled with quartz are common and are pronounced in the sandy layers but die out in the shaly layers, attesting to the difference in mechanical properties of the different rocks. Also note large boudins in places.

The assemblage here is staurolite + garnet + biotite + muscovite + quartz. Retrograde chlorite is present in the matrix. Most of the staurolite is gone in these rocks, being replaced by pseudomorphs of muscovite. Locally, one can find relict staurolite cores in the pseudomorphs. The pseudomorph-producing reaction is interpreted to be the retrograde reaction  $\text{staurolite} + \text{biotite} + \text{H}_2\text{O} = \text{garnet} + \text{chlorite} + \text{muscovite}$ .

- 5.9 (0.2) Continue straight (north) on dirt road.  
 6.55 (0.85) Route 121. Turn right (east).  
 7.0 (1.3) Stop light. Intersection with Route 5. Go straight through stop light and proceed north on route 5.

As we drive through Bellows Falls, note the Connecticut river on your right. The mountain above the river is Fall Mountain and is comprised of sillimanite + K-feldspar grade rocks of the Fall Mountain nappe. We will stop and examine these rocks in stop 5.

- 7.4 (1.7) Bellows Falls center. Continue north on Route 5.  
 7.55 (1.85) Miss Bellows Falls Diner on the left.  
 7.7 (2.0) Bridge across Connecticut River (to Rt. 12 and Walpole). Turn right and cross river to New Hampshire.  
 7.75 (2.05) Stop light at Route 12. Turn left (north) on Route 12.  
 9.1 (3.4) Road cuts on right (east) are of Bellows Falls pluton. The Bellows Falls pluton is a sheet approximately 100-200 meters thick with a very shallow contact on top and bottom. The pluton strikes northeast and dips south east so we are traversing down through the pluton as we drive north on Route 12.  
 9.45 (3.75) County line. Sullivan County.  
 10.4 (4.7) Large pullout on left (west). Pull off and park. Walk south 50 meters on route 12 and then cross road and climb grassy bank to railroad tracks. **Warning** – there is **poison ivy** hidden in the grassy bank. Outcrops in railroad cuts are good for sampling. First, examine small pavements of dark schist on east of railroad tracks approximately 30 meters to the north.

**Stop 4. Andalusite + biotite → garnet + chlorite + biotite zone.** (30 minutes). (Bellows Falls 15' quadrangle; Bellows Falls 7 1/2 minute quadrangle. Sample location BF-12). The purpose of this stop is to

examine rocks of the Skitchewaugh nappe immediately below the Bellows Falls pluton and the Fall Mountain nappe. Note that the best collecting is along the railroad cut.

A small outcrop of graded bedded schists occurs a few meters to the east of the railroad tracks (please do not hammer on this outcrop). The Bellows Falls pluton crops out a few tens of meters up the hill to the east. The rocks here are folded, presumably during the nappe stage. Visible in the outcrop are pseudomorphs after andalusite displaying chiastolitic crosses. Most of these pseudomorphs now consist of muscovite  $\pm$  chlorite but J. T. Cheney collected one sample that still contains relict andalusite. Matrix minerals include garnet + biotite + chlorite + muscovite + quartz + ilmenite. The earliest assemblage in this outcrop is andalusite + biotite + muscovite + quartz. Andalusite + biotite assemblages are replaced by garnet + chlorite assemblages and the P-T conditions recorded by the matrix assemblage is approximately 525 °C, 5-6 kbar (i.e. in the kyanite field). Chlorite is clearly late and overgrows the foliation.

The early andalusite + biotite assemblage is interpreted to represent a regional contact metamorphic episode with the heat source, in part, being the overlying Bellows Falls pluton. The P-T conditions for this "event" were approximately 525-600 °C, 2.5-3 kbar. The main fabric and folding was produced during the nappe stage and the rocks were loaded to approximately 5.5 kbar along a P-T path of approximately isothermal loading. The increase in pressure destabilized andalusite and stabilized garnet + chlorite assemblages via the reaction andalusite + biotite + H<sub>2</sub>O = garnet + chlorite + muscovite.

- 10.4 (4.7) Turn the vans around and return south on Route 12.
- 13.0 (7.3) Stoplight at bridge to Vermont. Turn left and continue south on Route 12 towards Walpole.
- 13.6 (7.9) Stoplight at bridge to Vermont. Note excellent outcrops in river. Proceed straight (south) through stoplight.
- 13.7 (8.0) Pull vans off onto shoulder on right (west) side of road. We will exit vans here so leave enough room to get out, but exercise caution, because this is a busy road. Climb down bank to outcrops in river below.

**Stop 5. Sillimanite + K-feldspar zone rocks of the Fall Mountain nappe.** (60 minutes. (Bellows Falls 15' quadrangle; Bellows Falls 7 1/2 minute quadrangle. Sample location BF-9, BF-14, 79-67). The purpose of this stop is to examine high grade rocks of the Fall Mountain nappe immediately above the Bellows Falls pluton. (Also see Thompson and Rosenfeld, 1979, Stop 2).

The rocks exposed in the river bed are the gray schist of the Rangeley Formation of the upper plate of the Fall Mountain nappe and the Bellows Falls pluton. Assemblages observed in the gray schist include quartz + plagioclase + garnet + biotite + sillimanite  $\pm$  spinel (only as inclusions within sillimanite)  $\pm$  chlorite (retrograde)  $\pm$  staurolite (retrograde)  $\pm$  K-feldspar (only as inclusions with garnet) + muscovite + ilmenite. Conspicuous in some samples is a white selvage surrounding the sillimanite porphyroblasts. This selvage is composed of late muscovite  $\pm$  staurolite which was produced during retrogression of the upper plate.

- 13.7 (8.0) Continue south on N.H. Route 12.
- 14.1 (8.4) Historic marker turnoff.
- 14.65(8.95) Turn left (east) onto Route 123 south.
- 15.1 (9.4) Turn left following Route 123 towards Alstead.
- 15.4 (9.7) Road forks. Take right fork off of Route 123. (Warning to users of Bellows Falls 15' quadrangle: Route 123 now goes by St. Peter's cemetery and the Route 123 shown on the maps is now the "old road.")
- 16.1 (10.4) Intersection at Old Drewsville Road. Go straight.
- 16.3 (10.6) Pull off on right side of road. Outcrop is on left side of road.

**Stop 6. Staurolite-kyanite (sillimanite?) zone, Skitchewaugh nappe (lower plate to the Fall Mountain nappe)** (20 minutes). (Bellows Falls 15' quadrangle; Walpole 7 1/2 minute quadrangle. Sample location BF-92). The purpose of this stop is to examine further P-T path constraints in rocks immediately below the Bellows Falls pluton in the Skitchewaugh nappe.

At this stop are schists of the Littleton Formation. We are only a few tens of meters below the contact of the Bellows Falls pluton, which crops out in the woods to the north. Metamorphic grade here is staurolite-kyanite to

- 13.6 (7.9) Stoplight at bridge to Vermont. Note excellent outcrops in river. Proceed straight (south) through stoplight.
- 13.7 (8.0) Pull vans off onto shoulder on right (west) side of road. We will exit vans here so leave enough room to get out, but exercise caution, because this is a busy road. Climb down bank to outcrops in river below.

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- 16.1 (10.4) Intersection at Old Drewsville Road. Go straight.
- 16.3 (10.6) Pull off on right side of road. Outcrop is on left side of road.

**Stop 6. Staurolite-kyanite (sillimanite?) zone, Skitchewaung nappe (lower plate to the Fall Mountain nappe)** (20 minutes). (Bellows Falls 15' quadrangle; Walpole 7 1/2 minute quadrangle. Sample location BF-92). The purpose of this stop is to examine further P-T path constraints in rocks immediately below the Bellows Falls pluton in the Skitchewaung nappe.

At this stop are schists of the Littleton Formation. We are only a few tens of meters below the contact of the Bellows Falls pluton, which crops out in the woods to the north. Metamorphic grade here is staurolite-kyanite to staurolite-sillimanite. Many samples contain the minerals staurolite + biotite + kyanite + garnet + chlorite + muscovite + quartz.

The most interesting rocks are to be found at the west (left) end of the outcrop. Here we see a schist that contains prismatic pseudomorphs. Photomicrographs of this rock are featured in Figure 9c and 9d of Spear et al. (1990). In cross section, these pseudomorphs have the appearance of chiastolitic crosses and it is suggested that they may be replacements of original andalusite. Within the pseudomorphs are randomly oriented flakes of muscovite and in the muscovite are growing garnet, staurolite, kyanite, biotite and chlorite. Biotite is seen touching staurolite and kyanite within the pseudomorphs, so the peak grade is in the kyanite + biotite zone. Locally within muscovite flakes is a fibrolitic sillimanite. It is not clear whether the sillimanite is in equilibrium with biotite. Garnet is commonly intimately associated with chlorite and it appears that both have grown together.

#### CONCLUDING REMARKS AND "TECTOON"

A "tectoon" illustrating the model for the evolution of the inverted metamorphism along the Bronson Hill anticlinorium is shown in Figure 24. The pre-tectonic setting of eastern North America was one of a marginal basin (possibly the back-arc basin to the Taconian island arc). Early Devonian plutons such as the Kinsman quartz monzonite and Bethlehem gneiss intruded the sediments (ca 400-410 Ma) of this basin causing the early high T, low P metamorphism. Subsequent thrusting juxtaposed rocks of different grade and P-T path.

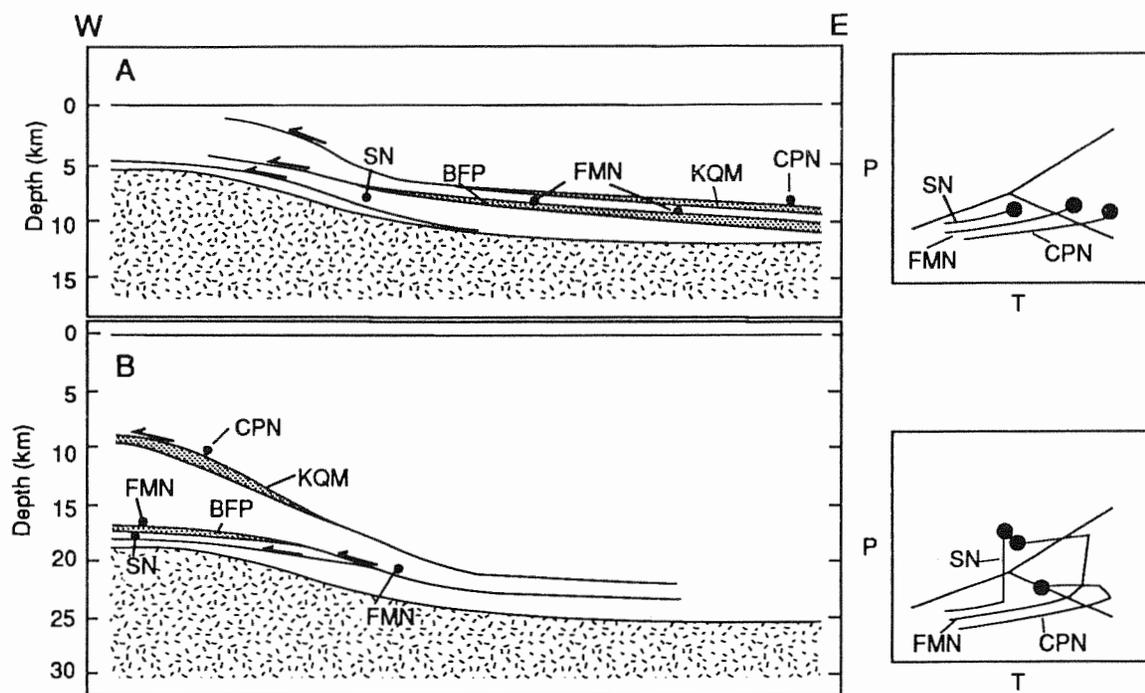


FIG. 24 "Tectoon" showing the evolution of the Acadian orogeny during the pre and early nappe stages. (a) shows the pre-thrusting environment and the intrusion of the Bethlehem gneiss and Kinsman quartz monzonite. (b) shows the relative juxtaposition of thrusts. Note that in this model the low pressures experienced by the Chesham Pond nappe rocks requires a bit of back sliding in order to bring these rocks to their present structural level.

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Chapter E

# Granitic Pegmatites in Northern New England

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## GRANITIC PEGMATITES IN NORTHERN NEW ENGLAND

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### INTRODUCTION

The New England region is famous for its granitic pegmatites which have been mined intermittently since 1803 when the Ruggles mine in Grafton, N. H. was opened for mica. Other commodities produced include beryl, feldspar, gemstones, and quartz. Mineral specimens are a perennial by-product of pegmatite mining and specimens from New England pegmatites are abundantly represented in public and private collections throughout North America and Europe. Study of the pegmatites and their minerals has led to a large scientific literature. The primary reference on New England pegmatites is *Pegmatite Investigations 1942-45 New England* by Cameron et al. (1954). It reviews the previous literature and summarizes the intensive study of New England pegmatites made by members of the U. S. Geological Survey to support development of domestic sources of strategic pegmatite minerals during World War II. The mineralogy, structure and zoning of many pegmatites are described in Cameron et al. and in most instances a detailed map is included.

Pegmatite mining has almost ceased in New England, but a few individuals continue to prospect and mine for gem rough and mineral specimens. This activity keeps some of the mines accessible and their exposures fresh. The four recently worked mines to be visited provide superior opportunities for the study of pegmatites in northern New England. The purpose of this trip is to examine the mineralogy, structure and zoning of these pegmatites. Their paragenesis will be discussed in light of mineralogical, textural and geochemical evidence and their relationships to nearby plutons will also be considered.

Most of the pegmatites mined in New England are concentrated in regions designated by the economic geology term *district* (Cameron et al., Figure 1). Cerny (1982, 1991b) has advocated a new vocabulary which describes pegmatite populations in petrogenetic terms. The fundamental unit is the pegmatite *group* which comprises several or many spatially-related and cogenetic pegmatite bodies. Wise and Francis (1992) used *series* to refer to spatially-related pegmatites in Maine that have not yet been demonstrated to be cogenetic or conclusively associated with specific granitic intrusives. A pegmatite *field* includes one or more pegmatite groups (or series) that occur within a common geological environment and were generated during a single tectonomagmatic event. The pegmatite groups in a single field have the same type of granitoid sources and approximately the same ages. The several pegmatite fields (formerly districts) in New England form a discrete *subprovince* of the Appalachian pegmatite *province*.

Granitic pegmatites are divided into *miarolitic*, *rare element*, *muscovite* and *abyssal* classes according to their mineralogy and geochemistry and their depth of emplacement and hence their metamorphic environment, (Cerny, 1982, 1991a). The economically important pegmatite fields in New England occur along the periphery of the Acadian metamorphic high in the sillimanite zone (Thompson and Norton, 1968). On the basis of their mineralogy (beryl, spodumene) they can be assigned to the rare element class and the LCT (lithium, cesium, tantalum) petrochemical family (Cerny, 1991c). This implies a low pressure (2-4 kb), upper greenschist to amphibolite facies (andalusite-sillimanite, 650-500°C) metamorphic environment which is appropriate for the Oxford field (Guidotti, 1989), but the Grafton field lies just to the high pressure side of the triple point isobar (Thompson and Norton, 1968). It, and to a greater extent the Keene field to the south, may represent the transitional rare-element to muscovite class mentioned by Cerny (1991a).

## PALERMO NO.1 MINE

## Overview

The Palermo No.1 mine is located in west-central New Hampshire in the Rumney 15' quadrangle a mile southwest of the former village of North Groton in the Town of Groton, Grafton County. It is one of the most famous pegmatites in the Grafton field for its production of mica, beryl, and other commodities as well as for its giant crystals of triphylite and many secondary phosphate species.

E. M. Simpson began the mining operations on the Palermo property about 1877 with an open pit on the east side of the pegmatite. The Hartford Mining Company acquired the property in February, 1878 and started an open trench on the south end of the pegmatite, working east and underground. The mineral rights were sold to George F. Bread in 1886 and passed to the Palermo Mining Company of New York one month later. Steam power was introduced, a crew of 85 of the local farmers was hired to work underground, and a trimming shop employing 25 women was set up on site. Near the end of the nineteenth century, the Palermo Mining Company went bankrupt, and the mine was sold at auction to Eugene Munsell & Company for \$50,000. The mine lay dormant for the next few years, but in 1914 the General Electric Company bought the mineral rights. During World War I General Electric mined mica underground. The upper cut was made sometime after 1920 to exploit feldspar. In 1942-43 G. E. resumed mining for mica and beryl. In May of 1944 the Ashley Mining Company was established in Rumney, New Hampshire with H. A. Ashley as president and E. Maltby Shipp as mining engineer. It took over the mining operation at Palermo and produced 4,222 tons of feldspar, 495 tons of scrap mica and 49 tons of beryl in 1947. The following year 1,800 tons of feldspar and 59 tons of beryl were mined. Poor health forced the closure of Ashley Mining Company in 1954. In 1958 John Maderic bought the mineral rights from General Electric and resold them to Milton Burlison of Mountain Mining Company the same year. The Mountain Mining Company operated underground for mica for three years. In 1963 Mineral Materials Incorporated leased the mine for five years and produced crushed quartz from the core. The Prudential Insurance Company building in Boston is faced with Palermo quartz. Some 40 tons of beryl were stockpiled as well. In 1973 Peter Samuelson leased Palermo solely for the production of gem beryl and mineral specimens. The following year R. W. Whitmore and Forrest F. Fogg (Palermo Mine Enterprises) purchased the mineral rights to preserve access to the locality for mineralogy. In the ensuing years they worked closely with Paul B. Moore and Anthony R. Kampf in their studies of pegmatite phosphates. Whitmore, now sole owner of the mineral rights, continues to investigate the pegmatite and collaborate with mineralogists.

## Geologic Setting

The Palermo No.1 mine lies in the Grafton pegmatite field (Olson, 1950, Plate I; Cameron et al. 1954, Plate I) on the west flank of the Kearsarge-Central Maine (formerly Merrimack) synclinorium (Billings, 1956, Lyons et al., 1991). It intrudes the light grey metaturbidite upper unit of the Lower Devonian Littleton formation, which was metamorphosed to sillimanite-muscovite grade during the Acadian orogeny. Representatives of all three members of the Devonian New Hampshire plutonic suite outcrop nearby. The northern end of the Cardigan pluton lies two miles southeast and the southern end of the Rumney pluton lies seven miles to the northeast. These are syntectonic gneissic S-type peraluminous granitoids of the Kinsman intrusive suite (Clark and Lyons, 1986). A mile and a half to the west is the Mt. Clough pluton, a strongly foliated biotite-muscovite granodiorite of the Bethlehem intrusive suite. A small mass of gray equigranular two-mica ("binary") granite belonging to the post-tectonic Concord granite intrusive suite outcrops at the terminus of the Cardigan pluton. No field and geochemical studies such as those undertaken in Maine have been made for western New Hampshire pegmatites. However, such observations as the lack of deformation and the lack of cross-cutting relationships (Cameron et al., 1954) make it probable that the pegmatites of the Grafton field represent a single intrusive episode and are members of the Concord suite.

The primary mineralogies of several pegmatites near the Town of Groton are summarized in Table 1. Although abundances are not indicated except for Palermo No.1, most of the pegmatites can be readily assigned to the *beryl type* of rare element pegmatites because beryl is common (Olson, 1950, Plate 7), whereas rare earth minerals and lithium-aluminum silicates are completely lacking in the Grafton field. Cerny (1991a, 1991c) has subdivided the beryl type into *beryl-columbite* and *beryl-columbite-phosphate subtypes*, but columbite is so sparse in this field that its use to denote a subtype seems inappropriate here. Accordingly, the Palermo No.1 is classed as a *beryl-phosphate subtype*.

Table 1. Primary mineralogy of the mica mines and prospects in Groton, Grafton Co., New Hampshire.

	1	2	3	4	5	6	7	8	9	10	11	12	13
beryl	X		X	X		X	C	X	X	X	X		X
biotite	X	X	X	X	X		C	X		X			
columbite							R				X		X
fluorapatite	X		X			X	U			X			X
garnet	X	X	X		X	X	R			X	X		X
grafftonite			X			X	U	?	X		X	X	
microcline	X	X	X	X	X	X	A	X	X	X	X	X	X
muscovite	X	X	X	X	X	X	A	X	X	X	X	X	X
plagioclase	X	X	X	X	X	X	A	X	X	X	X	X	X
quartz	X	X	X	X	X	X	A	X	X	X	X	X	X
tourmaline	X		X	X	X	X	C	X	X	X	X	X	X
triphylite	?		X			X	A	X	X		X	X	X
uraninite							U						

Mine names with sequence number from Cameron et al. (1954) Plate 1: 1. Charles Davis (24), 2. Craylip (26), 3. Fletcher (25), 4. Hackett (20), 5. Nancy No. 1 (23), 6. Nancy No. 2 (24), 7. Palermo No. 1 (27), 8. Palermo No. 2. (28), 9. Palermo No. 3 (29), 10. Pike's Ledge (31), 11. Rice (30), 12. Union (19), 13. Valencia (21).

X = present, A = abundant, C = common, U = uncommon, R = rare, ? = uncertain

Data from Cameron et al. (1954), Morrill (1960), and personal observations of R. W. Whitmore.

## Zoning

The Palermo No.1 (henceforth 'Palermo') is the most important of about sixty pegmatites on the Rice-Palermo property (Cameron et al., Plate 29). It was studied initially by Sterrett (1914, 1923) who provided the only description of the underground workings, which R. W. Whitmore recently reopened. Cameron et al. (1954) mapped Palermo in detail (Plate 30) and described its structure and zoning. It was restudied by Page and Larabee (1962, Plate 1) whose colored map at a smaller scale is easier to use. These authors revised the earlier description of the zoning. The two versions are correlated below:

Cameron et al. (1954)	Page and Larabee (1962)
Border Wall	Border Wall
1st and 2nd Intermediate	1st Intermediate
Core-Margin	2nd Intermediate
Core	Core-Margin and Core

Table 2 details the zoning following Page and Larabee (1962). The most recent geologic mapping (see Figure 1) was conducted in 1992 by R. W. Whitmore and is based upon an underground study of the mine.

## Mineralogy

The mineralogy of the Palermo mine is very well known. Segeler et al. (1981) listed 98 species. The phosphates have been the object of intensive study, but apart from these and the secondary uranium species, the minerals have not been investigated in much detail, especially from a petrogenetic point of view.

Muscovite, the mineral for which the mine was originally opened, occurs in at least small amounts throughout the pegmatite but was best developed in the wall and first intermediate zones. On the authority of L. R. Brown,

Table 2. Zoning of the Palermo No. 1 pegmatite (After Page &amp; Larrabee 1962).

Zone	Width	Development	Grain Size	Lithology	Comments
Border	<4"	north and east (hanging wall) contacts	very fine (<<1")	40-50% quartz 30-40% muscovite 10-20% plag., An <sub>12</sub>	Biotite, probably incorporated from wallrock, is a minor accessory.
Wall	<5'	north and east contact	medium (1-4")	50% plag., An <sub>4-7</sub> 35% quartz 15% muscovite	Biotite and schorl are minor accessories. This zone was an important source of sheet mica, mostly mined out.
First Intermediate	<25'	south and southwest (footwall)	medium to coarse (4-12")	45% plag., An <sub>4-7</sub> 25% quartz 25% perthite 5% muscovite	Biotite (in strips intergrown with muscovite) and schorl are common accessories. This zone produced some sheet mica.
Second Intermediate	<40'	north wall and east side (thought to have formed a hood over core)	medium to coarse	35% quartz 35% plag., An <sub>4-7</sub> 30% perthite	Any of the three major constituents may locally comprise 75%. This zone was mined for perthite before 1942.
Third Intermediate (Core-Margin)	<30'	footwall side of core	fine to coarse	30% plag., An <sub>4-7</sub> 25% quartz 25% muscovite 20% perthite	Albite occurs as oval aggregates of cleavelandite to 10' x 5' bordered by greenish-yellow wedge muscovite. Equally large masses of medium-grained muscovite (scrap mica) and of quartz, similar to core, separate cleavelandite aggregates. Beryl occurs as giant crystals most abundantly along interior margin; smaller golden crystals in muscovite masses. Giant triphylite crystals also found along inner margin. Uraninite occurs in muscovite masses.
Core	100'	central	very coarse (<12")	60% quartz 40% perthite	Subhedral perthite crystals reach 30' x 10'. Albite (An <sub>4</sub> ) occurs as thin, sugary-textured rims on perthite. Hydrothermal veins contain crystals of quartz and phosphates.

then superintendent for the Palermo Mining Company, Sterrett (1914) reported exceptional crystals as large as four feet across. Large masses of scrap muscovite can be seen in the core-margin zone.

Perthite was mined in the upper pit sometime after 1920 and again beginning in the mid 1940s by H. A. Ashley. The second intermediate zone is thought to have been a hood over the core up to 8 meters thick composed of essentially pure microcline perthite with crystal faces where it abutted the quartz core. This unit is now mostly mined out but giant crystals can be observed in the core. The plagioclase compositions cited in Table 2 are optical determinations quoted from Cameron et al. (1954). They represent the only analytical work on the rock-forming minerals from Palermo known to us except for one analysis of a plagioclase containing 0.227 wt.% P<sub>2</sub>O<sub>5</sub> reported by London et al. (1990). In that study the phosphorus content of alkali feldspars from 59 mostly North American pegmatites of nine types and subtypes was determined. A detailed examination of the phosphate content of Palermo feldspars is currently underway.

Biotite and tourmaline are the only mafic minerals present. Biotite will be seen as strips intergrown with muscovite in an outcrop of the first intermediate zone. Otherwise it is inconspicuous. Black tourmaline, presumably schorl, is abundant along the hanging wall exposed on the southeast side of the pegmatite near the arch. This may represent partially digested xenoliths of the country rock. It can also be found with biotite in the first intermediate zone. A single tourmaline prism enclosed in triphylite is among the material studied by Frondel (1949). Garnet is very rare at Palermo.

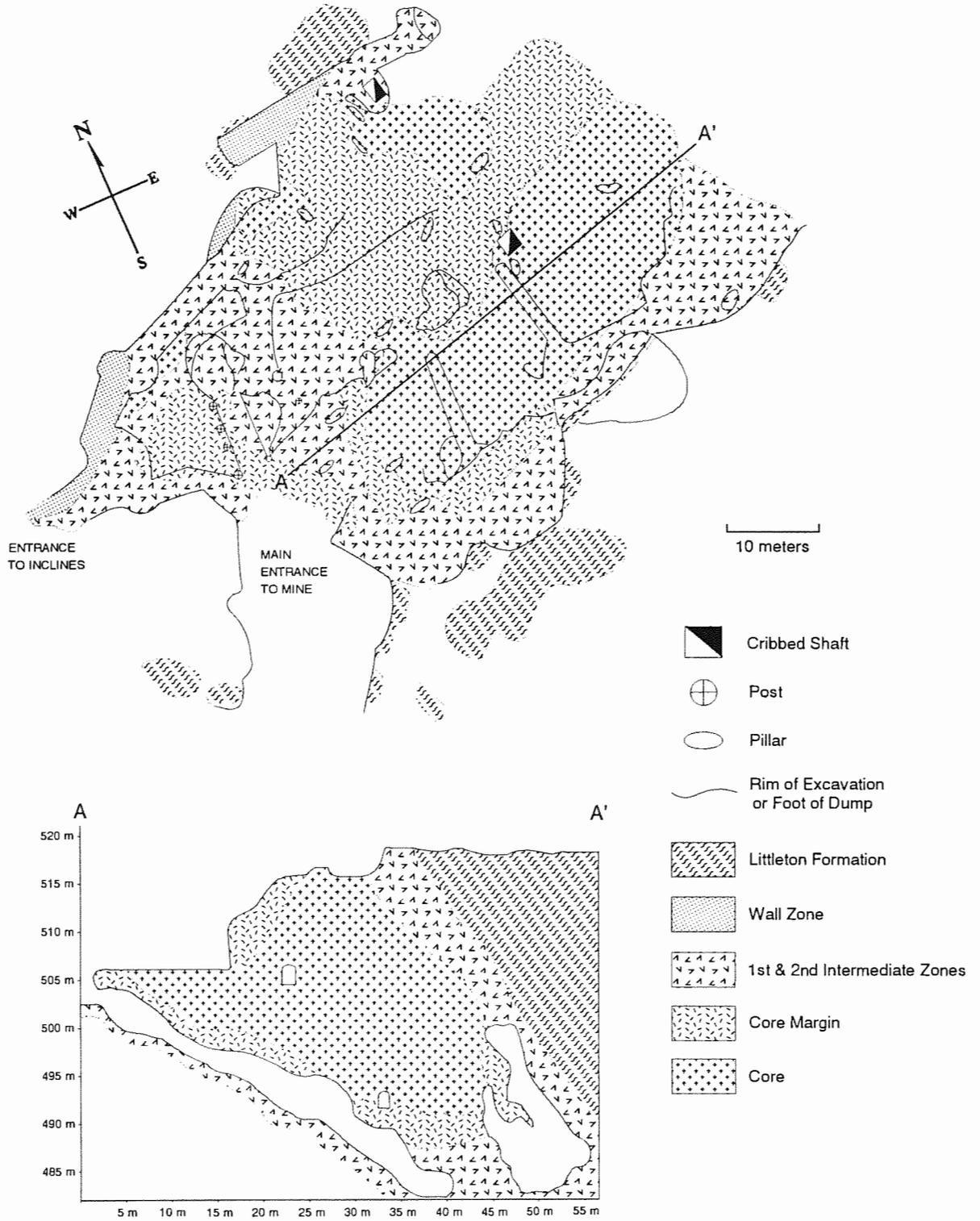


Figure 1. Geology of the Palermo No.1 pegmatite.

Table 3. Phosphate minerals of the Palermo No. 1 pegmatite (excluding fluorapatite and uranyl phosphates).

Phosphate	Formula	Stage			Occurrence		Oxidation	Other Cations
		P	M	H	Pods	Veins		
anapaite	$\text{Ca}_2(\text{Fe,Mn})(\text{PO}_4)_2 \cdot 4\text{H}_2\text{O}$			◆	U		■	Ca
arrojadite	$\text{KNa}_5(\text{Fe,Mn,Mg,Ca})_{14}\text{Al}(\text{PO}_4)_{12}(\text{OH})_2$		◆	◆	U		■	K,Na,Ca,Mg,Al
augelite	$\text{Al}_2\text{PO}_4(\text{OH})_3$	◆		◆	U	R		Al
barbosalite	$(\text{Fe,Mn})\text{Fe}_2(\text{PO}_4)_2(\text{OH})_2$			◆	U		■	
beraunite	$\text{Fe}_6(\text{PO}_4)_4(\text{OH})_5 \cdot 6\text{H}_2\text{O}$			◆	U		■	
bermanite	$\text{Mn}(\text{Fe,Mn})_2(\text{PO}_4)_2(\text{OH})_2 \cdot 4\text{H}_2\text{O}$			◆	R		■	
beryllonite	$\text{NaBePO}_4$			◆		R		Na,Be
bjarebyite*	$(\text{Ba,Sr})(\text{Mn,Fe,Mg})_2\text{Al}_2(\text{PO}_4)_3(\text{OH})_3$			◆	R		■	Ba,Sr,Mg,Al
brazilianite	$\text{NaAl}_3(\text{PO}_4)_2(\text{OH})_4$			◆		U		Na,Al
cacoxenite	$\text{Fe}_9(\text{PO}_4)_4(\text{OH})_{15} \cdot 18\text{H}_2\text{O}$			◆	R		■	
carbonate-hydroxylapatite	$\text{Ca}_{10}(\text{PO}_4)_3(\text{CO}_3)_3(\text{OH})_2$			◆	C	C		Ca
childrenite	$(\text{Fe,Mn})\text{AlPO}_4(\text{OH})_2 \cdot \text{H}_2\text{O}$			◆	C	U	■	Al
diadochite	$\text{Fe}_2(\text{PO}_4)(\text{SO}_4)(\text{OH}) \cdot 5\text{H}_2\text{O}$			◆	U		■	
dufrenite	$\text{Ca}_{0.5}\text{Fe}_6(\text{PO}_4)_4(\text{OH})_6 \cdot 2\text{H}_2\text{O}$			◆	R		■	Ca
eosphorite	$(\text{Mn,Fe})\text{AlPO}_4(\text{OH})_2 \cdot \text{H}_2\text{O}$			◆	C	U	■	Al
ernstite	$(\text{Mn,Fe})\text{AlPO}_4(\text{OH},\text{O})_2$			◆	R		■	Al
fairfieldite	$\text{Ca}_2(\text{Mn,Fe})(\text{PO}_4)_2 \cdot 2\text{H}_2\text{O}$			◆	R		■	Ca
ferrisicklerite	$\text{Li}_{1-x}(\text{Fe,Mn})\text{PO}_4$	◆			C		■	Li
foggite*	$\text{CaAlPO}_4(\text{OH})_2 \cdot \text{H}_2\text{O}$			◆	U			Ca,Al
frondelite	$(\text{Mn,Fe})\text{Fe}_4(\text{PO}_4)_3(\text{OH})_5$			◆	U		■	
goedkenite*	$(\text{Sr,Ca})_2\text{Al}(\text{PO}_4)_2(\text{OH})$			◆	U			Sr,Ca,Al
gordonite	$(\text{Mg,Fe})\text{Al}_2(\text{PO}_4)_2(\text{OH})_2 \cdot 8\text{H}_2\text{O}$			◆	R		■	Mg,Al
goyazite	$(\text{Sr,Ba})\text{Al}_3(\text{PO}_4)(\text{PO}_3\text{OH})(\text{OH})_6$			◆	U	U		Sr,Ba,Al
grafonite	$(\text{Fe,Mn,Ca})_3(\text{PO}_4)_2$	◆			U		■	Ca
hagendorffite	$\text{NaCaMn}(\text{Fe,Mn})_2(\text{PO}_4)_3$		◆	◆	U		■	Na,Ca
heterosite	$(\text{Fe,Mn})\text{PO}_4$		◆		C		■	
hinsdalite	$(\text{Pb,Sr})\text{Al}_3(\text{SO}_4)(\text{PO}_4)(\text{OH})_6$			◆	R			Pb,Sr,Al
hopeite	$\text{Zn}_3(\text{PO}_4)_2 \cdot 4\text{H}_2\text{O}$			◆	R			Zn
hureaultite	$(\text{Mn,Fe})_5(\text{PO}_3\text{OH})_2(\text{PO}_4)_2 \cdot 4\text{H}_2\text{O}$			◆	U		■	
hydroxyl-herderite	$\text{CaBePO}_4(\text{OH})$			◆		R		Ca,Be
jahnsite-(CaMnFe)	$\text{CaMn}(\text{Fe,Mg})_2\text{Fe}_2(\text{PO}_4)_4(\text{OH})_2 \cdot 8\text{H}_2\text{O}$			◆	U		■	Ca
kryzhanovskite	$(\text{Fe,Mn})_3(\text{PO}_4)_2(\text{OH},\text{H}_2\text{O})_3$			◆	R		■	
laucite	$(\text{Mn,Fe})\text{Fe}_2(\text{PO}_4)_2(\text{OH})_2 \cdot 8\text{H}_2\text{O}$			◆	C		■	
leucophosphate	$\text{K}(\text{Fe,Al})_2(\text{PO}_4)_2(\text{OH}) \cdot 2\text{H}_2\text{O}$			◆	R		■	K,Al

Columbite group minerals,  $(\text{Fe,Mn})(\text{Nb,Ta})_2\text{O}_6$ , are useful as exploration guides and for petrogenetic studies because they monitor Fe/Mn and Nb/Ta fractionation independently. These species are rare in the Grafton field. Only four specimens from Palermo, all ferrocolumbites, were available for analysis. They show a fraction trend toward slightly lower Mn and dramatically higher Ta concentrations. This is opposite to the trend of strong Mn enrichment followed by strong Ta enrichment seen in columbites from the Oxford field (Francis et al., 1991).

Beryl is the only primary beryllium mineral in the entire Grafton field. At Palermo, beryl crystallized only in the core-margin zone in sizes ranging from 1 centimeter to giant crystals 2 meters long and 1 meter in diameter. Page and Larabee (1962, Plate 1) indicate the location of concentrations or "nests" of beryl, but these have subsequently been removed. Giant crystals can, however, still be seen in the underground workings.

Phosphate	Formula	Stage			Occurrence		Oxidation	Other Cations
		P	M	H	Pods	Veins		
ludlamite	(Fe,Mn) <sub>3</sub> (PO <sub>4</sub> ) <sub>2</sub> • 4H <sub>2</sub> O			◆	C		■	
messelite	Ca <sub>2</sub> (Fe,Mn)(PO <sub>4</sub> ) <sub>2</sub> • 2H <sub>2</sub> O			◆	C		■	Ca
metavivianite	Fe <sub>3</sub> (PO <sub>4</sub> ) <sub>2</sub> (OH) <sub>2</sub> • 6H <sub>2</sub> O			◆	R		■■■	
mitridatite	Ca <sub>2</sub> Fe <sub>3</sub> O <sub>2</sub> (PO <sub>4</sub> ) <sub>3</sub> • 3H <sub>2</sub> O			◆	C		■■■■	Ca
montebrasite	LiAlPO <sub>4</sub> (OH,F)	◆		◆	U	R		Li,Al
moraesite	Be <sub>2</sub> PO <sub>4</sub> (OH) • 4H <sub>2</sub> O			◆		R		Be
palermoite*	(Li,Na) <sub>2</sub> (Sr,Ca)Al <sub>4</sub> (PO <sub>4</sub> ) <sub>4</sub> (OH) <sub>4</sub>			◆	U	R		Li,Na,Sr,Ca,Al
paravauxite	(Fe,Mg)Al <sub>2</sub> (PO <sub>4</sub> ) <sub>2</sub> (OH) <sub>2</sub> • 8H <sub>2</sub> O			◆	U		■	Mg,Al
phosphoferrite	(Fe,Mn) <sub>3</sub> (PO <sub>4</sub> ) <sub>2</sub> • 3H <sub>2</sub> O			◆	R		■	
phosphophyllite	Zn <sub>2</sub> (Fe,Mn)(PO <sub>4</sub> ) <sub>2</sub> • 4H <sub>2</sub> O			◆	R		■	Zn
phosphosiderite	FePO <sub>4</sub> • 2H <sub>2</sub> O			◆	R		■■■■	
pseudolaueite	(Mn,Fe)Fe <sub>2</sub> (PO <sub>4</sub> ) <sub>2</sub> (OH) <sub>2</sub> • 8H <sub>2</sub> O			◆	R		■■■	
pseudomalachite	Cu <sub>5</sub> (PO <sub>4</sub> ) <sub>2</sub> (OH) <sub>4</sub>			◆	R			Cu
rockbridgeite	(Fe,Mn)Fe <sub>4</sub> (PO <sub>4</sub> ) <sub>3</sub> (OH) <sub>5</sub>			◆	C		■■■	
roscherite (monoclinic)	Ca(Mn,Mg,Fe) <sub>2</sub> (Fe,□)Be <sub>2</sub> (PO <sub>4</sub> ) <sub>3</sub> (OH) <sub>3</sub> •2H <sub>2</sub> O			◆		U	■■■	Ca,Mg,Be
samuelsonite*	(Ca,Ba) <sub>9</sub> (Mn,Fe) <sub>4</sub> Al <sub>2</sub> (PO <sub>4</sub> ) <sub>10</sub> (OH) <sub>2</sub>			◆	R	R	■	Ca,Ba,Al
sarcopside	(Fe,Mn,Mg) <sub>3</sub> (PO <sub>4</sub> ) <sub>2</sub>	◆			R		■	Mg
schoonerite*	ZnMnFe <sub>3</sub> (PO <sub>4</sub> ) <sub>3</sub> (OH) <sub>2</sub> • 9H <sub>2</sub> O			◆	U		■■■	Zn
scorzalite	(Fe,Mg)Al <sub>2</sub> (PO <sub>4</sub> ) <sub>2</sub> (OH) <sub>2</sub>		◆		U		■	Mg,Al
sinkankasite	(Mn,Fe)Al(PO <sub>3</sub> OH) <sub>2</sub> (OH) • 6H <sub>2</sub> O			◆	R		■	Al
stewartite	(Mn,Fe)Fe <sub>2</sub> (PO <sub>4</sub> ) <sub>2</sub> (OH) <sub>2</sub> • 8H <sub>2</sub> O			◆	U		■■■	
strengite	FePO <sub>4</sub> • 2H <sub>2</sub> O			◆	R		■■■■	
strunzite	(Mn,Fe)Fe <sub>2</sub> (PO <sub>4</sub> ) <sub>2</sub> (OH) <sub>2</sub> • 6H <sub>2</sub> O			◆	C	U	■■■	
tavorite	LiFePO <sub>4</sub> (OH)			◆	R		■■■■	Li
triphylite	Li(Fe,Mn)PO <sub>4</sub>	◆			C		■	Li
vivianite	Fe <sub>3</sub> (PO <sub>4</sub> ) <sub>2</sub> • 8H <sub>2</sub> O			◆	C		■	
wardite	NaAl <sub>3</sub> (PO <sub>4</sub> ) <sub>2</sub> (OH) <sub>4</sub> • 2H <sub>2</sub> O			◆		R		Na,Al
whitlockite*	Ca <sub>18</sub> (Mg,Fe) <sub>2</sub> (H <sub>2</sub> ,Ca)(PO <sub>4</sub> ) <sub>14</sub>			◆	C	U	■	Ca
whitmoreite*	(Fe,Mn)Fe <sub>2</sub> (PO <sub>4</sub> ) <sub>2</sub> (OH) <sub>2</sub> • 4H <sub>2</sub> O			◆	U		■■■	
wolfeite*	(Fe,Mn) <sub>2</sub> PO <sub>4</sub> (OH)	◆	◆	◆	U		■	
xanthoxenite*	Ca <sub>4</sub> Fe <sub>2</sub> (PO <sub>4</sub> ) <sub>4</sub> (OH) <sub>2</sub> • 3H <sub>2</sub> O			◆	U		■■■■	Ca

\* Species first described from the Palermo No. 1 pegmatite  
Note: P = primary, M = metasomatic, H = hydrothermal, C = common, U = uncommon, R = rare.  
Oxidation: ■ = Fe<sup>2+</sup>; ■■■ = Fe<sup>2+/3+</sup>; ■■■■ = Fe<sup>3+</sup>(Mn<sup>2+</sup>); ■■■■■ = Mn<sup>2+/3+</sup>; ■■■■■■ = Mn<sup>3+</sup>

**Phosphate Mineralogy.** Sixty-nine phosphate minerals have been identified from this pegmatite. The uranyl phosphates, autunite/meta-autunite, torbernite/metatorbernite, and phosphuranylite, are discussed under Uranium Mineralogy below. Except for fluorapatite, a primary mineral found in the pegmatite's outer zones, the remaining phosphates at Palermo are essentially confined to phosphate pods and hydrothermal veins. It is the occurrence and paragenesis of this latter group that has garnered the greatest scientific interest and yielded ten new mineral species (see Table 3). For individual descriptions of the phosphate species the reader is referred to Segeler et al. (1981).

The large variety of phosphate minerals at Palermo is a direct result of the wide range of conditions over which they have formed. Temperatures have varied from about 700°C down to 50°C or lower. Oxidation conditions have varied from reduced (all iron in the divalent state) to oxidized (all iron trivalent) and extremely oxidized (all manganese trivalent). Solution chemistries have varied considerably, both spatially and temporally.

The phosphate paragenesis can be divided into three successive stages or processes which are largely temperature dependent. Superimposed upon these are the effects of oxidation and chemistry. The stages and their characteristics are as follows:

*Primary crystallization (700°C-600°C).* During this stage, several phosphates, most notably triphylite, crystallized directly from the pegmatite melt. Their placement in the core margin zone indicates that they formed prior to the consolidation of the bulk of the core. Reducing conditions prevailed throughout primary crystallization. The minerals formed are water-free, although a few contain hydroxyl. Large, crude crystals are typical of primary crystallization. The placement of chemically different primary phases determines, in large part, the spatial distributions of secondary phases. The predominance of iron over manganese in the triphylite,  $\text{Li}(\text{Fe}_{0.76}\text{Mn}_{0.19}\text{Mg}_{0.05})\text{PO}_4$ , indicates a similar predominance in the pegmatite magma; furthermore, a similar predominance is reflected in most secondary phosphates containing both elements.

*Metasomatic alteration (600°C-350°C).* This stage was characterized by the leeching of lithium from triphylite and the addition of sodium, calcium, and aluminum from adjacent areas. The process took place with very little volume change and relatively small movement away from the original source minerals. Temperatures were still too high to allow incorporation of water into phosphate structures. Conditions were locally reducing to very oxidizing as indicated by the valence state of the iron and manganese in the phases produced. Granular masses exhibiting little or no crystal form (i.e., scorzalite) are typical of metasomatic minerals.

*Hydrothermal attack and remobilization (350°C-50°C).* During this stage, core-derived aqueous solutions attacked the primary and metasomatic phosphates, remobilizing their metal and phosphate ions. Triphylite was most susceptible to attack, and large portions of these giant crystals were replaced. Early high-temperature hydrothermal solutions rich in  $\text{CO}_2$  resulted in large masses of interlocking siderite crystals, while those rich in  $\text{Ca}^{+2}$  resulted in dense intergrowths of whitlockite and carbonate-hydroxylapatite. At lower temperatures, massive ludlamite and vivianite resulted. Within these major replacement units, a vast variety of phosphate species crystallized in lesser quantities. Earlier minerals are often found embedded in the massive replacement units, while later ones grew freely in cavities in these units. Metal and phosphate ions were also transported away from the phosphate pods, forming crystals in hydrothermal veins and sometimes incorporating additional ions, such as  $\text{Be}^{+2}$  from beryl. Quartz, which is a common constituent throughout the hydrothermal assemblages, is especially prevalent in these veins. Fe and Mn are not readily transported in the hydrothermal solutions and are, for the most part, confined to the phosphate pods. The phosphates formed below about 250°C often contain free water, with later, lower temperature phases usually containing greater proportions. Some phosphates probably formed by near surface alteration resulting from circulating ground water and weathering, but they cannot always be distinguished from those of strictly hydrothermal origin. Oxidation conditions can be found covering the entire range from reduced to very oxidized, with assemblages often showing an evolution from reduced to oxidized. Associated species often bear striking similarities in chemistry, especially with respect to certain metals: Fe, Al, Ca (Sr, Ba), Zn, and Be.

The secondary phosphate micro-assemblages are extremely numerous and varied at the Palermo No.1 pegmatite and a more detailed treatment of paragenesis and associations is beyond the scope of this field guide. The discussion of paragenesis above, coupled with the information compiled in Table 3, should provide important insights to anyone attempting to make sense of the phosphate mineral assemblages encountered.

**Sulfides.** Sulfide minerals occur only sparsely and have not been carefully investigated. They occur particularly with the triphylite in the core-margin zone. Magnetic pyrrhotite was abundant in veinlets penetrating cleavage planes of a cluster of giant triphylite crystals exposed in the mid 1970s. Pyrite may be found in triphylite that shows signs of incipient alteration and in the alteration rinds as lustrous euhedral crystals up to one centimeter in diameter. Pyrrhotite is completely absent in the alteration assemblages which is consistent with known phase relations (Craig & Scott, 1974). Frondel (1949) and Wolfe (1949) noted the presence of arsenopyrite and sphalerite in addition to pyrite. A replacement of triphylite by massive whitlockite and carbonate-apatite uncovered in the mid 1970s contained concentrations of both sphalerite and galena. Bornite and chalcopyrite have been found associated with the triphylite as well. In general, all but the iron sulfides are considered rare. Cameron et al. (1954) reported pyrite and chalcopyrite as well as lazulite (actually scorzalite) as accessories in their first intermediate zone. Having never observed sulfides or scorzalite in that zone, we suspect that these actually originated in a phosphate pod in the core-margin zone. Sulfides also occur in the core zone where isolated masses of bornite, chalcopyrite, galena, and loellingite in quartz have been collected by R. W. Whitmore.

**Uranium Mineralogy.** Uraninite,  $UO_2$ , is an accessory mineral in many pegmatites, but never occurs in sufficient quantity to constitute an ore. Its alteration, like the alteration of triphylite, gives rise to numerous secondary species. The Ruggles and Palermo mines are particularly noted for their abundance and variety of uranium minerals. At Palermo, uraninite was reported to occur in the "muscovite masses" of the core-margin zone (Page & Larabee, 1962). In fact, the matrix of the many uraninite specimens in the Harvard collection is a medium-grained plagioclase-muscovite-quartz lithology which will be seen in outcrop in the centers of nodular masses rimmed by cleavelandite. In detail, uraninite occurs as dendritic concentrations (up to 10 cm in diameter) of octahedral crystals (typically 1mm but up to 1 cm in diameter) enclosed in plagioclase. Muscovite is the common associate. Smoky quartz, golden beryl and brown zircon crystals may be present. The feldspar or quartz immediately surrounding uraninite crystals may be stained dark red by hematite? along cracks. This has been erroneously labelled clarkeite by local collectors, but clarkeite has not yet been confirmed at Palermo.

The secondary uranium minerals from Palermo were studied by Frondel (1956) as part of his larger study of "gummite," the ill-defined term referring to the colorful alteration rims on and pseudomorphs after crystals of uraninite. Alteration of uraninite involves the oxidation of  $U^{4+}$  to form the divalent uranyl cation  $(U^{6+}O_2)^{2+}$ . All of the secondary uranium minerals are uranyl compounds. The paragenesis follows a regular sequence which can be divided into three stages. First is the *in situ* oxidation of  $U^{4+}$  to  $U^{6+}$  by the incorporation of O atoms into the uraninite structure. E. Berman (1955) determined the unit cell edge of Palermo uraninite to be  $a = 5.439 \text{ \AA}$ , which is in the lower part of the range for pegmatitic uraninites reported by R. Berman (1957). No chemical analyses (or age determinations) of Palermo uraninite are known to the writers.

The second or gummite stage involves the formation of the brightly colored alteration rinds and pseudomorphs. The red-orange to orange inner zone is composed of hydrated lead uranyl oxides, the lead being radiogenic. Vandendriesscheite, fourmarierite?, and the incompletely described Mineral A were reported from Palermo by Frondel. Although these minerals are conspicuous because of their color, they are not nearly as abundant as the yellow minerals of the last stage.

The third stage of alteration accounts for the outer zone of yellow minerals of the gummite "eyes" and the yellow or yellow-green films and crusts deposited in cracks that form halos about the original site of the uraninite. Lead has been leached, and calcium, silica or phosphate, and water are added. The hydrated calcium uranyl silicates, uranophane and beta-uranophane, and the hydrated calcium uranyl phosphates, autunite/meta-autunite and phosphuranylite, were identified at Palermo. Autunite is particularly common and easy to identify by its green fluorescence under both short wave and long wave ultraviolet light. Because of the mobility of the uranyl ion in solution, autunite may potentially be found as tiny specks in specimens from any zone. Where copper is available from the breakdown of sulfides, torbernite, the copper analogue of autunite crystallizes. Both dehydrate from the decahydrate to the hexahydrate readily, so collection specimens are likely to be meta-autunite or metatorbernite, at least in part.

Frondel concluded that the alteration of uraninite is caused by interaction with meteoric waters rather than late stage pegmatitic fluids. The uranium mineral assemblage at Palermo is currently being restudied by Stanley Korzeb and Eugene Foord who have also identified schoepite. A formal description of Mineral A is underway. In this study rutherfordine was identified among Harvard specimens (L. Pittman, personal communication).

## MIAROLITIC PEGMATITES OF THE WHITE MOUNTAIN BATHOLITH

### Overview

The Conway granite, a prominent unit throughout the White Mountain igneous province, is noted for its miarolitic pegmatites. In North America, only the Pikes Peak field in Colorado (Foord and Martin, 1979) with its amazonite- and topaz-bearing pegmatites is more famous as an example of the miarolitic class of granitic pegmatites. Hurricane Mountain was added to the itinerary to provide an opportunity to examine briefly one of the occurrences of miarolitic pegmatites and contrast it with the rare-element pegmatites that are the main subject of this trip.

Miarolitic pegmatites are in general much smaller bodies than the rare-element pegmatites and they usually occur within or adjacent to epizonal plutons eliminating all ambiguity about their parentage. Cavities yielding

Table 4. Minerals from cavities in the miarolitic pegmatites of the Conway Field.

albite	C	biotite	C	loellingite	R	quartz	C
allanite-(Ce)	R	calcite	U	magnetite	R	riebeckite	U
anatase	R	cassiterite	U	microcline	C	schorl	R
arfvedsonite	U	danalite	U	microlite	R	siderite	C
arsenopyrite	R	epidote	R	milarite	R	stilpnomelane	R
astrophyllite	R	ferrocolumbite	R	molybdenite	U	thorite	R
bastnaesite-(Ce)	R	fluorapatite	R	monazite-(Ce)	R	titanite	R
bavenite	R	fluorite	C	muscovite	C	topaz	C
bazzite	R	goethite	U	phenacite	U	zircon	U
bertrandite	R	hematite	U	pyrochlore	R		
C = common, U = uncommon, R = rare							

fine crystals of smoky quartz, microcline, topaz and other species are characteristic of, but not unique to, this pegmatite class.

### Geological Setting

The White Mountain batholith is the largest of sixteen A-type granitoid intrusions of Jurassic (200-155Ma) or Cretaceous (130-110Ma) age in eastern New Hampshire. It is a focus of Trip 15 'Ring Dikes and Plutons: The Deeper View of Calderas' by Creasy and Eby in this volume (q. v.). This synopsis is drawn from Eby et al., (1992). The batholith consists of three intrusive petrologic associations: (1) gabbro-diorite-monzonite, (2) alkali syenite-quartz syenite-alkali granite, and (3) metaluminous biotite granite. In addition, the Moat volcanics comprise a 3,600 m thick caldera-fill sequence of comendite, trachyte and tuff-breccia. The batholith is thought to be the result of fractional crystallization of mantle-derived melts that were somewhat contaminated by a crustal component.

The Mt. Washington Valley and surrounding region is a noted resort area located in the eastern portion of the White Mountain batholith. The oldest unit, dated at 187 my, is the Mount Osceola granite, a green, medium-grained amphibole granite. The Albany porphyritic quartz syenite (179 and 170 my) forms incomplete ring dikes surrounding the Moat volcanics. The prevailing rock type is the Conway granite, a pink, medium grained, two-feldspar biotite granite with accessory zircon, allanite, fluorapatite and titanite. Detailed mapping has revealed separate intrusions that date at 183, 180, 171 and 155 my.

The history of mineral collecting in the Conway pegmatite field and the macroscopic cavity minerals have been reviewed by Samuelson et al. (1990). Collectors prospecting for new digging sites have explored the granite near its contact with the overlying Moat volcanics on the basis that late, pegmatite-forming fluids rose to the top of the pluton but were trapped by its volcanic cap. The lower slopes of Moat Mountain have been a prolific source of specimens for decades. In addition, gravel pits in the floor of the valley, which exploit grus or "rotten rock" for local road construction, sometimes expose solid ledges containing pegmatites. The Government pit in the White Mountain National Forest in Albany was a popular collecting site until it was closed in 1986. The abundance of cavities at this site seems inconsistent with the foregoing explanation because it was deeper in the pluton than the collecting sites on Moat Mountain. However, Hoisington (1977), in the only geological study made of the Government pit, recognized that the cavities occur along a contact between two masses of the Conway granite. A tabular pegmatite about 0.3m thick exposed for at least 5 meters along strike and about 3 meters down dip can presently be seen there. Several notable finds have been made in this particular pegmatite. A 56 x 44 x 12 cm specimen illustrating the paragenesis of a cavity or "pocket" is displayed at the Harvard Mineralogical Museum. Fine-grained graphic granite with biotite laths is succeeded by coarse-grained microcline (cream and green) + albite + quartz. Euhedral crystals of microcline, quartz, muscovite and topaz project into the central space which was later filled with clay. The mineralogy of the cavities (Table 4) is well described by Bearss and Janules (1992), who emphasize the microscopic crystals so frequently overlooked by collectors of

cabinet specimens. The petrology and geochemistry of the miarolitic pegmatites have not been studied by modern methods.

### GRANITIC PEGMATITES OF THE OXFORD PEGMATITE FIELD

The granitic pegmatites of southwestern Maine occur within a 50 km wide, northwest trending band extending from the coastal region near Rockland to Rangeley Lake, approximately 30 km west of the Maine - New Hampshire border. The pegmatite population is divided into the Brunswick and Oxford pegmatite fields based on differences in gross mineralogy, chemistry and textures (Table 5). The Brunswick field pegmatites vary from moderately fractionated rare-earth type to beryl type (beryl-columbite subtype) to highly fractionated complex (spodumene subtype) rare-element pegmatites. The bulk of the pegmatites in the Oxford field are moderately to highly fractionated beryl-columbite, beryl-columbite-phosphate, spodumene and petalite subtypes.

The Oxford pegmatite field is dominated by the extensive Sebago batholith in the southern portion of the field. Several smaller granodioritic (e.g. Songo pluton) and granitic plutons (e.g. Rumford, Mooselookmeguntic, Phillips plutons) lie to its north and east. Pegmatites are concentrated interior and marginal to the granitic plutons, both of which intrude middle- to upper-amphibolite grade metasedimentary rocks (Fig. 2).

The pegmatites are primarily known for their production of gem tourmaline, but purple apatite, aquamarine and morganite have also attracted the attention of miners and collectors. The interest in pegmatites in the Oxford field dates back to 1820 when the first discovery of tourmaline in North America was recorded at the Mt. Mica pegmatite. The commercial mining of feldspar, muscovite, beryl, pollucite and gem tourmaline has continued sporadically for the last 173 years. Although feldspar mining in Maine began in Topsham in 1852, feldspar quarries in the Oxford field opened in the late 1800s at Mt. Apatite. Sheet and scrap mica mining probably began soon after the discovery of Mt. Mica, although most of the mica mining in the area was largely a by-product of feldspar mining. Between 1942 and 1945 the mines were examined particularly for the strategic metals Be, Nb, Ta and Li (Cameron et al., 1954). Current studies are concentrating on the petrogenesis of the pegmatites and related granites using field relationships and geochemical data (Brown and Wise, 1991; Francis et al., 1991; Wise and Francis, 1992).

#### The Mount Apatite Pegmatites (Western Quarries)

**Background.** Gem-bearing pegmatites exposed on Mount Apatite, Androscoggin County, Maine, have been known since the first tourmaline was found there in 1868 (Bastin, 1911). An account of the mining history of these quarries is summarized in Table 6 and in Perham (1987). The quarries include the Maine Feldspar and Greenlaw quarries on the east flank of Mt. Apatite and the well-known Pulsifer, Wade, Keith and a more recent excavation, The-Hole-In-The-Ground (hereafter referred to as Hole) on the western flank from which gem tourmaline and purple apatite have been found.

Table 5. General characteristics of the Brunswick and Oxford pegmatite districts, Maine.

Pegmatite Field	Internal Structure	Texture	Geochemistry	Mineralogy
Brunswick	Poorly zoned Non pocket-bearing	Graphic common	REE, Be, Nb>Ta, ((Li))	biotite > muscovite monazite, magnetite, beryl, columbite
Oxford	Moderately to well zoned Pocket-bearing	Graphic uncommon, replacement, aplitic and metasomatic units common	Li, Rb, Cs, Be, Nb $\approx$ Ta, P, B, F	muscovite > biotite beryl, columbite-tantalite, cassiterite spodumene, lepidolite, petalite, tourmaline

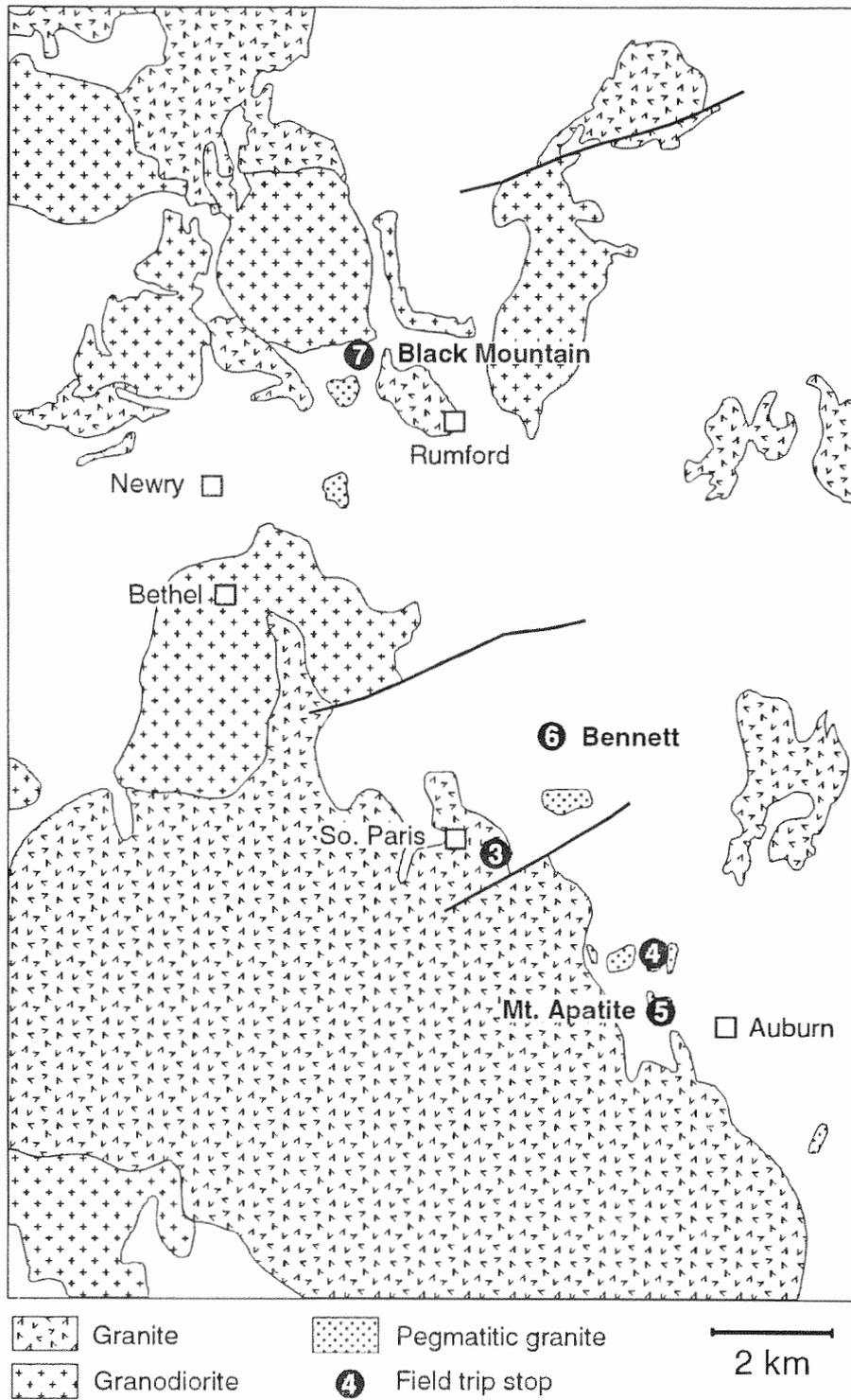


Figure 2. Map of the Oxford pegmatite field.

Table 6. Mining histories of the Mt. Apatite, Bennett and Black Mountain pegmatites.

Date	Mining Operator	Commodity
<b>Mt. Apatite</b>		
1883 (MA)	Nathaniel Perry	tourmaline
1884 - 1890 (MA)	Loren Merrill / Thomas Lamb	tourmaline
1891 (MA)	Mt. Apatite Company	tourmaline?
1896 - 1906? (Tr)	E. Y. Turner	feldspar?
1901 - 1904 (P,W)	Pitt Pulsifer	tourmaline, apatite
1902 - 1930's (MF)	Maine Feldspar Co.	feldspar
1904 - 1906 (W)	Maine Tourmaline Co.	tourmaline, aquamarine
1907 - 1909? (Tw)	Maine Feldspar Co. / J. S. Towne	feldspar, tourmaline
1916 - 1948 (K)	Martin L. Keith	tourmaline, feldspar
Mid 1940's (P)	Stanley Perham / Hillard Nevin	tourmaline
1963 (P)	Ken Grover	dump material
1964 - 1965 (P)	Irving Groves / Frank Perham	tourmaline, apatite
1966-1967 (P)	Terrance Szenics / Frank Perham	tourmaline, apatite
1966 - 1967 (W)	Irving Groves	tourmaline
Early 1970's - present (P)	Irving Groves	tourmaline, apatite
1971 - present (K)	Irving Groves	tourmaline
<b>Bennett</b>		
1920 - 1924	Paul Bennett	feldspar
1924 - 1926	Maine Feldspar Co.	feldspar
1926 - 1931	Harold C. Perham	quartz crystals
1931 - 1933	Whitehall Feldspar Co.	feldspar
1944	United Feldspar & Minerals Co.	feldspar
1970		
1977		
1989 - present	Ron and Dennis Holden	fourmaline, morganite
<b>Black Mountain</b>		
1901 - 1903	Stamford Mica Company	mica
1904 - 1905	Gildersleeve & Poor Co.	scrap mica, feldspar
1931 - 1938	Stanley Perham	scrap mica
1938 - 1942	Cesare Trusiani	scrap mica, feldspar, spodumene, beryl, lepidolite
MA = Mt. Apatite, Tr = Turner, P = Pulsifer, W = Wade, MF = Maine Feldspar, Tw = Towne, K = Keith		

**Description of the pegmatite.** The local geology of the Mt. Apatite area has been mapped by Creasy (1979) and is dominated by exposures of the Sebago batholith which intrude calc-silicate rocks interbedded with biotite granofels, pelitic gneiss and schists. The pegmatite is exposed in a discontinuous series of excavations over a distance of approximately 750 m. Biotite schist-pegmatite contacts are exposed in three of the quarries. The contacts strike approximately N75°E and dip 20-24° NW. These contacts are interpreted as the top (hanging wall) surface of the pegmatite in each case, and nowhere is the lower (footwall) contact exposed. A number of nearly vertical basaltic dikes, all less than one meter in width and striking N45°E and N60°E, cut the pegmatite.

Five distinct zones are observed within the pegmatite:

*Border zone.* The border zone consists of fine- to medium-grained equigranular quartz, plagioclase, minor biotite and garnet. The extent of this zone is difficult to ascertain due to its poor exposure.

Table 7. Mineralogy of the Mt. Apatite, Bennett and Black Mountain pegmatites.

	MAP	BEN	BMT		MAP	BEN	BMT
albite	C	A	A	mccrillsite	-	-	R
amblygonite / montebrasite	R	R	R	microcline	A	A	C
apatite	R	C	R	microlite	-	-	R
arsenopyrite	-	R	-	mitridatite	-	-	R
beryl	R	C	R	montmorillonite	R	R	R
beryllonite	-	-	R	muscovite	A	A	A
biotite	R	R	R	pollucite	R	R	R
cassiterite	R	R	R	pyrite	R	-	R
cookeite	R	R	R	quartz	A	A	A
elbaite	C	C	C	reddingite	-	R	R
eosphorite	-	R	R	rhodochrosite	R	R	R
fairfieldite	-	R	R	rockbridgite	-	-	R
ferrotapiolite	-	R	-	romanechite	-	R	R
gahnite	R	-	-	roscherite	-	R	R
garnet	C	R	R	schorl	R	C	C
goyazite	-	R	R	siderite	-	R	R
heterosite	-	-	R	sphalerite	-	-	R
hureaulite	-	R	-	spodumene	-	R	C
hurlbutite	-	-	R	strunzite	-	-	R
hydroxyl-herderite	R	R	R	topaz	R	R	-
kaolinite	R	R	R	triphylite / lithiophilite	R	R	R
kosnarite	-	-	R	uraninite	R	R	R
landesite	-	R	-	uranophane	-	-	R
lepidolite	R	R	C	vivianite	-	-	R
manganocolumbite	R	C	C	whitlockite	-	-	R
manganotantalite	R	R	R	zircon	-	R	R

MAP = Mt. Apatite, BEN = Bennett, BMT = Black Mountain  
A = abundant, C = common, R = rare

*Wall zone.* The wall zone consists of slightly graphic K-feldspar, quartz, biotite and garnet.

*First intermediate zone.* The first intermediate zone is characterized by coarse-grained graphic feldspar and plumose muscovite-quartz aggregates. Garnet and schorl are found as accessory minerals and rare beryl was observed in the Wade quarry. This zone varies in thickness from approximately 1.5 meters in the Wade quarry to 6 meters in the Keith quarry.

*Second intermediate zone.* The first intermediate zone grades irregularly into a coarse-grained plagioclase + quartz + muscovite second intermediate zone.

*Pocket zone.* The pocket zone assemblage, which lies below the second intermediate zone and immediately above the "garnet seam", consists primarily of cleavelandite and quartz, although locally blocky K-feldspar and muscovite are also abundant. Gem tourmaline, apatite, hydroxyl-herderite, manganocolumbite and cookeite are among the minerals which are found within the pocket zone. The "garnet seam" occurs as a 1 to 5 cm thick layer composed of 2 mm to 3 cm sized euhedral garnet + anhedral smoky quartz. The "garnet seam" is reportedly continuous throughout all the quarries (I. Groves, oral communications, 1991), and it serves as a guide to the location of gem pockets. Mining below the "garnet seam" was considered to be unpromising in the search for gems as no pockets or "mineralization" occur below the seam. Thus mining was confined to just above the "garnet seam" and as a result, the zonation below the seam and the thickness of the pegmatite are unknown. An

assemblage immediately adjacent to the "garnet seam" (perhaps below the seam) consists of blocky feldspar, smoky quartz, gahnite and minor garnet.

**Mineralogy.** The mineralogy of the Mt. Apatite pegmatite (see Table 7) is highlighted by the occurrence of purple fluorapatite and green tourmaline. Purple fluorapatite is typically embedded in quartz and black mud found in small pockets within cleavelandite-rich areas immediately above the "garnet seam". Gem tourmaline at Mt. Apatite includes green, pink, lilac, blue and colorless elbaite which occurs most frequently in pockets. Manganocolumbite is the most common Nb-Ta oxide mineral present, but rare. Garnet is most prevalent in the "garnet seam" and green spinel (gahnite) is found in the unit immediately above the "garnet seam".

### The Bennett pegmatite

**Background.** The Bennett pegmatite is one of several complex, rare-element granitic pegmatites originally mined for feldspar and gem minerals. The pegmatite is located approximately 5 km west of Buckfield and 7 km northeast of Paris just off Paris Hill Road. The world-famous Mt. Mica quarry is approximately 5 km west of the Bennett pegmatite along the same road. Recently, the pegmatite has become well-known due to its production of perhaps the best gem-grade morganite from the eastern U.S.

**Description of the pegmatite.** The geology in the vicinity of the Bennett pegmatite has been mapped by Warner (1967), Pankiwskyj et al. (1976) and Osberg et al. (1985) as a sequence of interbedded metasandstone, metapelite and limestone of Silurian age. The metamorphic grade of the rocks of the area is upper amphibolite facies (Guidotti, 1989). These rocks have been intruded by the Carboniferous age (325 Ma) Sebago batholith, the Streaked Mountain pegmatitic granite and numerous pegmatite bodies (Osberg et al., 1985). The pegmatite crops out irregularly over a distance of approximately 380 m and is roughly 200 m wide at its maximum (Landes, 1925; Cameron et al., 1954). It strikes east-west, dips 65-75° to the south, and is hosted by interbedded quartz-biotite schist, amphibolite and gneiss. The pegmatite is well-zoned and mineralogically diverse.

Detailed study of the pegmatite shows that the internal zonation (see Figure 3) consists of a wall zone, two intermediate zones, a core-margin, a core and an aplitic unit (Wise et al., in press). Much of the primary zonation in the western part of the quarry has been obscured by a cleavelandite-rich replacement unit which hosts gem-bearing pockets.

**Wall zone.** The wall zone consists of fine- to medium-grained plagioclase, with subordinate quartz, biotite and K-feldspar. Accessory red-brown garnet occurs locally with biotite. Numerous large xenoliths of quartz-biotite schist and gneiss occur throughout this zone.

**First intermediate zone.** The wall zone grades into a medium- to coarse-grained first intermediate zone dominated by plagioclase, sub-graphic K-feldspar and quartz. This zone is also marked by the occurrence of abundant black tourmaline and the general absence of biotite. Red-brown garnet, biotite, muscovite and green apatite occur sporadically throughout the zone.

**Second intermediate zone.** The second intermediate zone consists largely of very coarse-grained blocky perthite, quartz, muscovite and accessory black tourmaline and green apatite. Green to blue-green beryl is locally abundant. This zone has been partly replaced by the cleavelandite-rich replacement unit in the west end of the quarry.

**Core Margin.** The core margin is comprised of blocky microcline, quartz, plagioclase and spherical green apatite. Apatite reaches a maximum dimension of roughly 10 cm, but is typically only 5 mm to 2 cm in diameter. Minor anhedral schorl, columnar yellow-green beryl, medium-grained muscovite plates and manganocolumbite are scattered throughout.

**Core.** The lens-shaped core is approximately 30 m long, and consists of milky quartz and very, coarse-grained, blocky, gray to tan perthitic microcline. Cleavage faces of some microcline vary in size from 20 cm to 2 meters in length. Coarse-grained books of muscovite occur locally. Masses of arsenopyrite, up to 10 cm across, and blue-green beryl are rare.

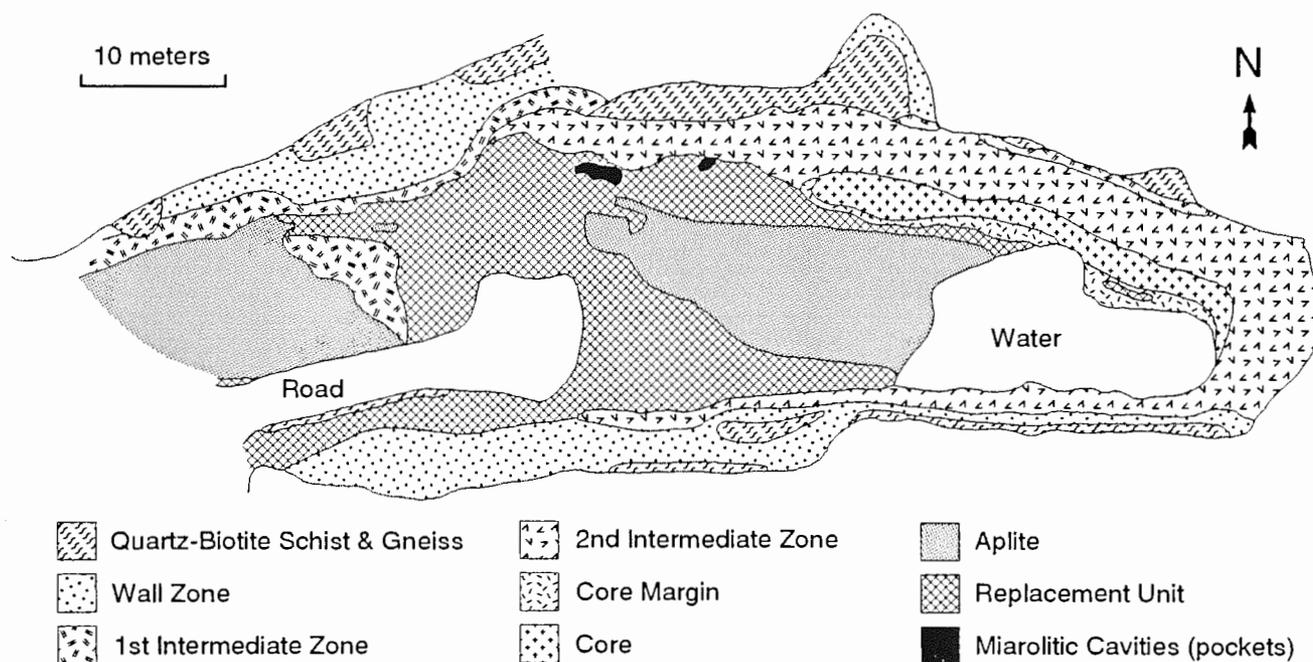


Figure 3. Geology of the Bennett pegmatite.

**Aplite.** The central portion of the pegmatite is occupied by a gray aplitic unit elongated along the long axis of the pegmatite body. Smaller lenses and blocks of aplite are also found scattered elsewhere in the pegmatite. The aplite consists of fine-grained equigranular quartz, albite, garnet and mica with abundant pea-sized poikilitic black tourmaline intergrown with quartz. Locally, the aplite varies from light gray in color to dark gray due to changes in the color and amount of muscovite. The garnets are generally less than 1 mm in diameter and are red-brown to pink in color. Rare pale green apatite is sparsely distributed throughout the unit.

**Replacement unit.** The replacement unit consists of medium- to coarse-grained cleavelandite with accessory schorl, yellow to green beryl, quartz, green apatite, muscovite and lepidolite. Radiating blades of manganocolumbite and masses of cassiterite are locally abundant. This unit cuts both intermediate zones and the aplite. A discontinuous band of large subhedral schorl crystals, some reaching 13 cm in diameter, occur along the contacts of the replacement unit and the aplite and second intermediate zone. In all cases, however, the tourmaline masses are found only in the cleavelandite.

**Pockets.** Pocket development is restricted to the replacement unit. Small pockets, averaging roughly 30 cm in diameter and 15 to 20 cm deep, are formed in the replacement unit, typically near the aplite contacts and in cleavelandite veins cutting the aplite. These pockets commonly contain cleavelandite, lepidolite, cookeite, "pencil" elbaite, hydroxyl-herderite, rare manganocolumbite-manganotantalite and pink clay (montmorillonite?). Cookeite commonly coats everything and contains fragmented tourmaline crystals. Large muscovite books adjacent to the pocket become curved and some are clearly altered to lepidolite. Larger pockets varying in size from 2 to 4 meters long, such as those which contained abundant morganite crystals, occur in a "zone" near the contact of the replacement unit and the second intermediate zone in the west-central portion of the pegmatite. The ceilings of these pockets are made up of large clear to smoky quartz crystals, many of which point downward into the pocket. The walls and floor consist of transparent to white cleavelandite with occasional pink or colorless beryl, green to blue apatite up to 2 cm in length and 1 cm across, manganotantalite and cassiterite. Tourmaline is usually uncommon or absent.

**Mineralogy.** The mineralogy of the Bennett pegmatite is quite diversified (Table 7) and similar to a number of pegmatites in western Maine, particularly Mt. Mica. A thorough description of the Bennett pegmatite mineralogy has been given by Landes (1925) and Wise et al. (in press). Quartz and feldspars are ubiquitous throughout the entire pegmatite, some reaching several meters in length. The principal mica species which

dominates the Bennett pegmatite is muscovite, although biotite, lepidolite and cookeite are locally abundant in some zones. Almandine-spessartine is locally abundant in the wall and first intermediate zone, the aplite, the replacement unit and the xenoliths of quartz-biotite gneiss. Tourmaline occurs in most zones of the pegmatite, as well as in the aplite and replacement unit. Schorl predominates in the first and second intermediate zones, aplite and replacement unit. Zoned tourmaline, (black cores and green rims, "watermelon" and "cucumber") are sometimes observed in the second intermediate zone and pocket assemblages of the replacement unit. Beryl varies in color from yellow to green in the second intermediate zone, to rare aquamarine in the core and near the pocket zone to pale pink to orange-pink columnar and stubby crystals of the pocket zone. Colorless and milky-white beryl occurs infrequently in the pocket zone as small tabular crystals with etched bipyramid and prism faces. Nb-Ta-Sn oxides occur as crystals of manganocolumbite-manganotantalite, cassiterite and rare ferrotapiolite. Green to blue fluorapatite is the dominant phosphate mineral found in the Bennett pegmatite. Other phosphates occurring in minor quantities include amblygonite-montebrasite, triphylite-lithiophilite, and secondary phosphates.

### The Black Mountain pegmatite

**Background.** The Black Mountain pegmatite is located in the northern portion of the Oxford pegmatite field. Best known for its radial aggregates of pink tourmaline (rubellite) in massive fine-grained lepidolite, the Black Mountain pegmatite is located 11.5 km NW of the town of Rumford. Within close proximity to the pegmatite are the Rumford and Whitecap Mountain granitic plutons.

**Description of the pegmatite.** Outcrops of pegmatite occur as six roughly lenticular, subparallel bodies which strike N25°W and dip 20-25° NE (Maillot et al, 1949). The pegmatites intrude interbedded biotite granofels, two-mica sulfidic schists and quartzite.

Five zones and two units were identified in the main exposure of the pegmatite (see Figure 4). The quartz-muscovite-plagioclase border zone consists of yellow-green muscovite intergrown with gray anhedral quartz. Dark green tourmaline averaging 1 cm in length are also present as are small amounts of white anhedral plagioclase. Small, less than 1 mm in diameter, flakes of sulfides are disseminated throughout the unit and are locally weathered, imparting a rusty coating to the zone.

**Wall zone.** The albite-quartz-muscovite-tourmaline wall zone consists primarily of coarse-grained light yellow-green subhedral muscovite ranging in size from 2 to 6 cm in diameter. The diameter of the muscovite increases progressively away from the border zone towards the first intermediate zone. The muscovite together with subhedral gray-white plagioclase and gray anhedral quartz form the main constituents of the wall zone. Subordinate blue-black tourmaline and red garnet are also present. The tourmaline ranges in size from 1 to 12 cm in diameter. Red anhedral garnets are scattered throughout the zone.

**First intermediate zone.** The muscovite-cleavelandite first intermediate zone consists of roughly equal amounts of muscovite and cleavelandite with subordinate green tourmaline. The tan muscovite occurs as large books approximately 45 x 75 cm. Green tourmaline crystals are often flattened or splayed and are intergrown between the sheets of the muscovite books. The tourmaline blades are typically 1 to 2 cm long, but some up to 5 cm in length have been found. Occurring with the large muscovite books is coarse-grained white cleavelandite which locally contains yellow beryl up to 7 cm in diameter.

**Second intermediate zone.** The spodumene-quartz-K-feldspar-lepidolite second intermediate zone contains predominantly anhedral gray quartz associated with white to tan elongated blades of spodumene, microcline, schorl and lepidolite. The microcline is sparse and is approximately 2 x 3 cm. Bi-colored pink to green tourmaline occurs locally, but is rare. Fine-grained lepidolite pods, typically less than 30 cm long, may contain 1 to 2 cm long, pink, light blue-gray, or colorless tourmaline. A large lepidolite pod approximately 3 meters in length which grades in color from violet to apple green to silver occurs on the left side of the pegmatite. These lepidolite pods are small in comparison to the exposure of a 4 x 6 meter mass of lepidolite reported by Marble and Morill (1945).

**Lepidolite unit.** Locally the second intermediate zone grades into a unit that is a medium- to fine-grained mixture of lepidolite, albite and rubellite with occasional spodumene blades. The texture of this unit varies from

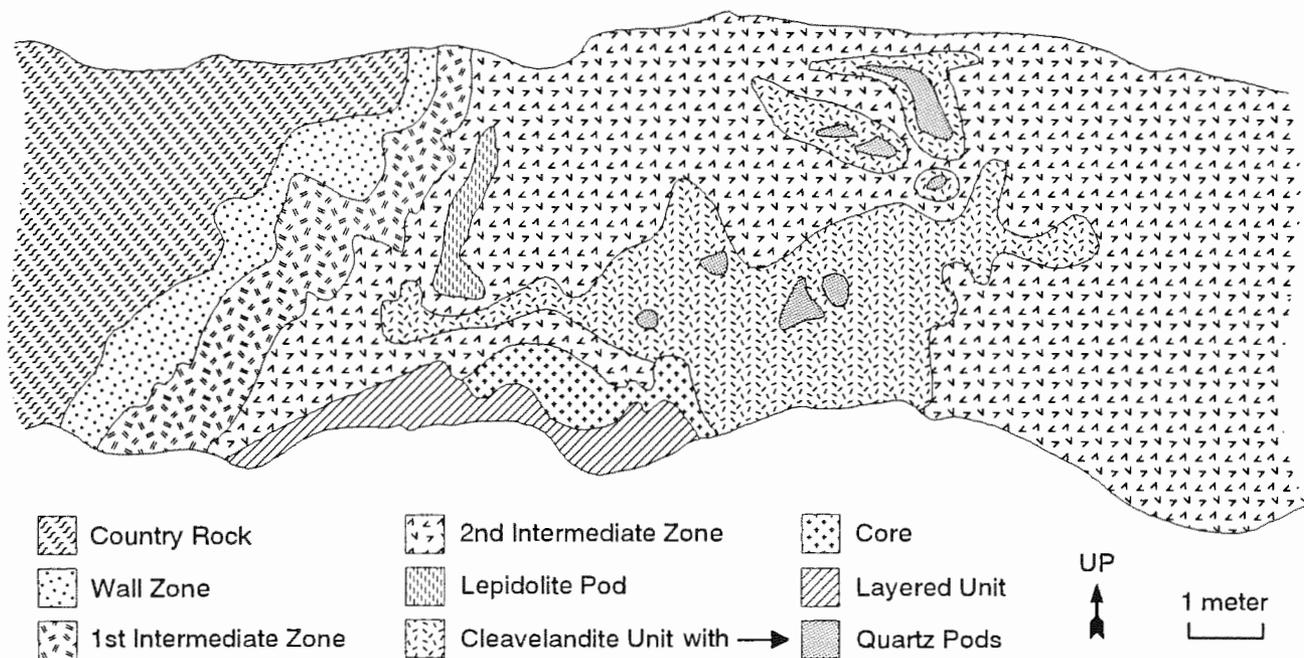


Figure 4. Vertical section of the main exposure of the Black Mountain pegmatite.

medium- to fine-grained, equigranular to segregated layers of lepidolite versus pink elbaite and albite. This unit has been referred to as a tourmaline granite (Bailey, 1929) or "rubellite granite" (Brown and Wise, 1991).

**Core.** The core consists of anhedral quartz which is typically massive, but also occurs as blebs suspended in albite. Large, blocky (8 x 15 cm to 18 x 25 cm) perthitic microcline is locally surrounded by a 1 cm thick rim of amblygonite. In addition to amblygonite, spodumene, fine-grained lepidolite and pink tourmaline occur locally. The core is discontinuous, occurring as a massive lens near the bottom of the pegmatite. It segregates into quartz pods towards the top of the pegmatite.

**Cleavelandite unit.** Coarse white-gray cleavelandite, which makes up the bulk of the cleavelandite unit, hosts numerous large dark blue-green tourmalines, bladed black-brown columbite and brown-black massive cassiterite. The tourmaline is euhedral and is approximately 7 to 8 cm long and 2 to 4 cm in diameter. Gray quartz occasionally occurs within the cleavelandite. Manganocolumbite generally occurs as 5 cm long blades and cassiterite occurs locally as 3 cm subhedral crystals. Muscovite is rare, but when found, it occurs as silver books. Opaque, light blue-gray amblygonite was identified in this unit by X-ray diffraction analysis.

**Aplite.** Samples collected from the dumps indicate the presence of an aplite. The aplite was not observed in place, but from dump pieces it has been ascertained that it occurs next to the wall zone. The aplite is a fine-grained mixture of quartz, plagioclase, muscovite, schorl and apatite. The tourmaline in the aplite is schorl and is locally separated into bands.

**Layered unit.** A unit of alternating 1 cm thick layers of cleavelandite and silver lepidolite occurs beneath the massive quartz core. The unit ranges from a very distinct layered unit to a fine-grained granitic unit. Also occurring in this unit are minor amounts of amblygonite, columbite and quartz. Rafts of up to 25 cm of layered unit material are found suspended in the overlying quartz core. Lenses of bluish white saccharoidal albite containing disseminated cassiterite occur locally within the second intermediate zone.

**Mineralogy.** The Black Mountain pegmatite is composed primarily of quartz, albite, muscovite and perthitic microcline. Tourmaline from the pegmatite includes schorl, which is most common in the wall zone, dark green-black elbaite of the cleavelandite-tourmaline unit and green elbaite of the first intermediate zone. Black Mountain is best known for its spectacular fans of pink elbaite (rubellite) which occur in lepidolite pods. Black

manganocolumbite and massive dark brown cassiterite are abundant in the cleavelandite unit. Lithium mineralization is best characterized by the occurrence of spodumene, lepidolite and amblygonite-montebrazite. Spodumene occurs as white crystals up to 1.5 meters long in the second intermediate zone along with pods and lenses of purple, yellow-green and silver lepidolite. Other minerals present in minor quantities include yellow and white beryl, almandine, triphylite, zircon and pyrite (see Table 7).

## Granites

Granitoids within the Oxford pegmatite field are the predominant igneous rocks populating the region. They are typically heterogeneous, varying in composition from tonalite to granite. Pegmatites are generally marginal to the granites and as a result, previous workers have used this field relationship as the sole criteria for identifying granites parental to the pegmatites. Recognition of a separate pegmatitic facies within the plutons and detailed geochemical data on the granitoids is poor. Establishing parent granite-pegmatite relationships through detailed field examination and geochemical analysis is the focus of work currently in progress (unpublished data of M. A. Wise).

**The Sebago batholith.** The Sebago batholith is a large Carboniferous age ( $325 \pm 3$  Ma; Aleinikoff et al., 1985) heterogeneous body, varying from typical granite textures to strongly pegmatitic textures. Near the center of the pluton, a porphyritic to gneissic biotite granite facies with small pegmatitic veins of simple mineralogy (K-feldspar, quartz, biotite) are observed. Towards the margins of the batholith, the biotite granite grades to a fine- to medium-grained muscovite-biotite or muscovite granite. Along the eastern margin, but less so on the west, a pegmatitic facies, commonly hosting xenoliths of metasedimentary rocks, is developed. Immediately to the east of the batholith, are several small granitic stocks which have been determined from field examination to be pegmatitic granites. Pegmatitic granite has been identified at Streaked Mountain, Singepole Mountain, Hebron and Center Minot and it is believed that these plutons are part of the main Sebago batholith. The pegmatitic granite consists of coarse, blocky K-feldspar megacrysts surrounded by a medium-grained matrix of quartz, plagioclase and muscovite with accessory tourmaline and minor garnet. Locally, this phase grades into potassic pegmatite pods of simple mineralogy, although rare beryl and columbite have been observed. A sodic aplite facies consisting of fine-grained albite, quartz and accessory garnet and tourmaline is locally developed and commonly is interlayered with pegmatitic veinlets.

**The Rumford area granitoids.** The granitic rocks of the Rumford area consist primarily of sphene-bearing granodiorite, two-mica granite and pegmatitic granite. The Mooselookmeguntic pluton ( $371 \pm 6$  Ma; Moench and Zartman, 1976) is described as a mixture of medium-grained sphene-bearing granodiorite, tonalite and quartz diorite which occur as separate phases or as large, blocky inclusions in two-mica granite (Moench and Hildreth, 1976). Accessory minerals include allanite, apatite, epidote, zircon, ilmenite and pyrite. The fine- to medium-grained two-mica granite contains accessory garnet, apatite and zircon.

The Rumford pluton is in many respects identical to the Mooselookmeguntic pluton with the exception that a pegmatitic granite facies is strongly developed. This pegmatitic granite is best exposed at Whitecap Mountain where it consists of coarse, graphic, megacrystic K-feldspar and medium-grained quartz, albite and muscovite (Wise and Francis 1992). The coarse-grained pegmatitic granite grades into very coarse-grained potassic pegmatite pods and layered sodic aplite. Minor beryl has been found in the pegmatitic pods and unlike the pegmatitic portions of the Sebago batholith, tourmaline appears to be uncommon.

## Geochemistry

The K/Rb ratios of microcline and micas attest to the variable, yet highly fractionated nature of the Mt. Apatite, Bennett and Black Mountain pegmatites. K/Rb for blocky K-feldspar from central zones ranges from 141.8 to 22.9, 105.5 to 20.7 and 44.1 to 14.5 for the Mt. Apatite, Bennett and Black Mountain pegmatites, respectively. A plot of K/Rb vs Cs for K-feldspar illustrates the chemical diversity and progressive fractionation within the individual pegmatites (see Figure 5). The subparallel but nonsequential trends may suggest different styles of fractionation, different sources or both. The trace element fractionation in muscovite-lithian muscovite-lepidolite from the Mt. Apatite, Bennett and Black Mountain pegmatites shows subparallel, but overlapping trends for K/Rb vs Tl and K/Rb vs Sn (see Figure 5). The fractionation trends for columbite-tantalite are characterized by progressive Mn enrichment from ferrocolumbite through manganocolumbite to manganotantalite. The trends are also representative of the entire Oxford pegmatite field as illustrated by Francis and Wise (1991).



Nb/Ta fractionation within the pegmatites is generally moderate as shown by the predominance of columbite over tantalite in most pegmatites. The presence of cassiterite and wadginite is evidence of Sn enrichment which is not uncommon in highly fractionated pegmatites. The presence of lepidolite, elbaite, minor spodumene, montebrasite, triphylite-lithiophilite and pollucite strongly supports at least moderate levels of Li and Cs fractionation.

The pegmatites surrounding the Sebago batholith (e.g. Mt. Apatite, Bennett and Mt. Mica) are strongly enriched in B and F as indicated by the presence of abundant tourmaline-bearing pockets. Local enrichment in P is observed in the Mt. Apatite and Bennett (apatite) and Mt. Mica (amblygonite-montebrasite) pegmatites, but in general, the pegmatites surrounding the Sebago batholith appear to be P-poor.

In contrast, pegmatites of the Rumford area are notably P-rich as evidenced by the occurrence of abundant Li-Fe-Mn phosphates. However, the level of P enrichment is considerably less than that observed in the Groton, New Hampshire or Black Hills, South Dakota areas. With the exception of the Black Mountain and Newry pegmatites, most pegmatites of the Rumford area are B-poor and virtually devoid of tourmaline. Furthermore, with the exception of the Newry pegmatite, the Rumford area pegmatites generally lack gem pockets.

### Genetic Considerations

Preliminary field observations and chemical data strongly suggest that the peraluminous and heterogeneous Sebago, Rumford and Phillips plutons are parental to pegmatites distributed along their margins (unpublished data of M. A. Wise). Gradual textural and mineralogical changes from primitive biotite granite through two-mica granite to more fractionated pegmatitic leucogranite are observed within these plutons and are characteristic of fertile granites parental to rare-element pegmatite fields as described by Cerny and Meintzer (1988).

The pegmatites, which are distributed in a crude regional zonation about the granites, crystallized late in the magmatic history of the pluton from a geochemically evolved, hydrous silicate melt that was granitic in composition, yet enriched in Be, Nb, Ta, Li, Rb, Cs and locally in B, P, and F. Pressure estimates based on the stability relations of lithium aluminosilicates (London, 1984), suggest that pegmatites containing primary spodumene + quartz or petalite-bearing assemblages (e.g. Mt. Mica, Bennett, Tamminen, Black Mountain, Newry) probably crystallized between 3 and 4 kbars. Other cogenetic pegmatites lacking lithium aluminosilicates probably crystallized in the same pressure range.

The development of pockets within the Maine pegmatites may be linked to the timing of the crystallization of tourmaline which removes the solidus depressing boron from the pegmatite melt (London, 1986). Pegmatites which crystallize tourmaline early and continuously in their evolution (e.g. Black Mountain) tend to be devoid of pockets, whereas pegmatites containing pockets tend to have abundant tourmaline crystallizing in the late stages of pegmatite consolidation (e.g. Bennett).

In conclusion, it must be added that little petrochemical data exists for the Maine pegmatites despite their popularity among mineral collectors and scientists. Therefore, any models of pegmatite evolution presented at this time are no more than generalizations or speculations. Solving the basic questions concerning the petrogenesis of the Maine pegmatites must await the completion of detail mineralogical and geochemical studies currently in progress.

### ACKNOWLEDGEMENTS

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## ROAD LOG

**Mileage**

- 0.0 Road log begins at I-93 exit 26 in Plymouth, N. H. Proceed westward on Route 25 to Rumney Depot.
- 4.7 Turn left on Halls Brook road, which ends at the site of the former village of North Groton.
- 8.4 Turn left across the bridge and a tenth of a mile further on turn right onto Edgar Albert Road (not signed).
- 9.5 The gravel road becomes an unimproved wood road. From the gate, which will be unlocked, continue through the woods, over the bridge, and right up the hill.
- 10.3 Park at the mine.

**STOP 1. PALERMO NO.1 PEGMATITE.** The purpose of this stop is to examine a well-zoned pegmatite, both on the surface and underground, with particular attention to the variations in mineralogy and texture of the several zones.

- 0.0 Road log begins in Conway on Route 16 at the intersection of Route 153. Proceed north on Route 16.
- 0.1 Bear left at light. Drive through North Conway almost to the Bartlett line.
- 1.8 Turn right onto Hurricane Mountain road.
- 3.5 Note Bill Ross' rock shop on right. Beyond here the grade gets quite steep. At the sign indicating the steep hill, shift down and use the lower gear to the top.
- 5.4 Park at the top of the mountain on the right hand side of the road. Walk back to the sign for the trail to Black Cap. Take the unimproved access road on the opposite (north side) of the road up the hill to where a slight clearing due to ledges crosses the road. This is about a ten minute walk. From there walk up to the right to the ledges where excavated pegmatites may be inspected.

**STOP 2. HURRICANE MOUNTAIN.** This locality was selected for its relatively easy access but it is a petrologically interesting choice as well. Much of the mountain is underlain by Conway granite with topaz-bearing pegmatites, but on the summit a late-formed riebeckite granite is exposed with pegmatites that are pod-shaped unlike the tabular body exposed in the Government Pit. Several pegmatites excavated by mineral collectors (Gallant, 1960) display mineralogical and textural zoning. Black prismatic arfvedsonite crystals embedded in feldspar and core quartz will be obvious. Cavities have yielded fine crystals of smoky quartz, microcline, and arfvedsonite with accessory zircon and rare astrophyllite.

- 0.0 Road log begins in South Paris at the junction of Routes 26, 117 and 119. Take Rt. 119 (Hebron Rd.)
- 3.4 Park and examine outcrop on right.

**STOP 3. SEBAGO BATHOLITH.**

- 3.4 Continue south on Rt. 119.
- 5.7 Hebron Post Office on right.
- 8.7 West Minot turn right continuing on Rt. 119.
- 12.9 Turn left onto Shaw Hill Road.
- 14.2 Turn left onto Center Minot Road.
- 14.8 Pass church on left and turn right onto Jackson Hill Road.
- 15.4 Park in access road on left by water-filled pit. Walk about 200 feet ahead to pavement outcrop in woods on right.

**STOP 4. SEBAGO BATHOLITH.**

- 15.4 Return to Rt. 119.
- 17.6 Turn left (south) onto Rt. 119 from Shaw Hill Road.
- 19.6 Turn left onto Rts. 11 & 121 at Hackett Mills.
- 21.8 Turn left onto Hatch Road.
- 22.6 Turn left into mine road and park.

**STOP 5. MT. APATITE PEGMATITES.**

- 22.6 Return to Rts. 11 & 121 at the foot of the hill.

**Mileage**

- 23.4 Turn right onto Rts. 11 & 121.
- 29.1 Proceed straight through intersection at Mechanic Falls on Rt. 121.
- 39.3 Turn right onto Rt. 26 at Welchville and return to South Paris.
- 46.9 Bear left at the traffic circle, remaining on Rt. 26.
- 47.3 Turn right across bridge onto Paris Hill Road.
- 48.5 Bear right at fork and continue to the golf club.
- 49.7 Turn right onto Buckfield Road just beyond golf club.
- 54.4 Bennett farm is on the left. Follow the farm road through the woods to the mine.

**STOP 6. BENNETT MINE.**

- 0.0 Road log begins in Rumford at the junction of Rts. 2 & 120. Proceed north on Rt. 120.
- 7.8 Turn left onto secondary road.
- 9.9 Turn left onto gravel road and proceed uphill.
- 10.6 At utility pole #J-17 the principal road continues to the left and a logging road branches to the right. Continue straight ahead up the hill to the mine.

**STOP 7. BLACK MOUNTAIN PEGMATITE.****REFERENCES CITED**

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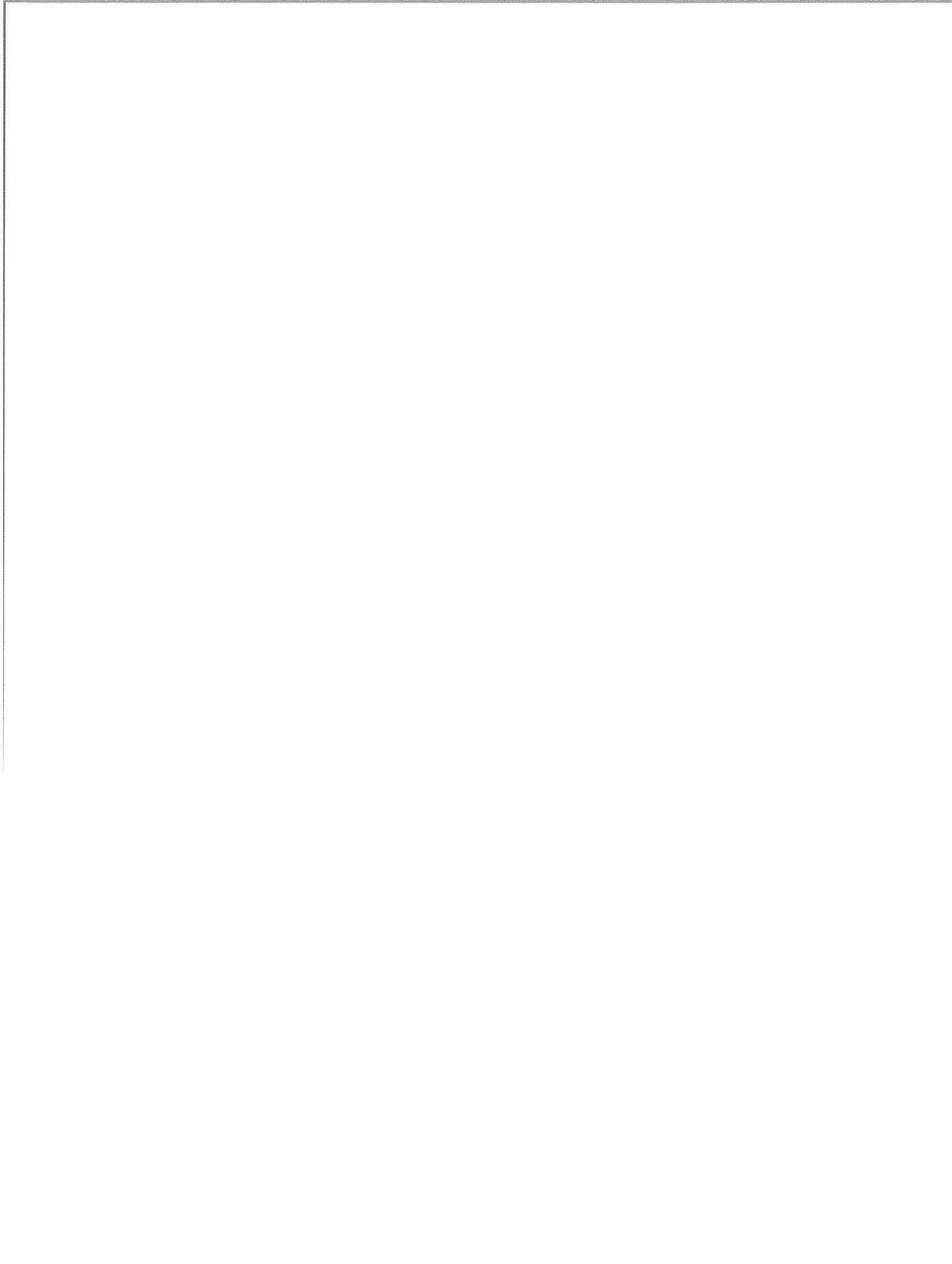
Chapter F

# A Transect Through the Taconide Zone of Central Vermont

By R.S. Stanley, D.C. Martin, and R.A. Coish

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## A TRANSECT THROUGH THE TACONIDE ZONE OF CENTRAL VERMONT

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### INTRODUCTION

Central Vermont provides an excellent field laboratory for studying the tectonic transition between the foreland and the pre-Silurian hinterland of western New England because many of the major lithotectonic units are exposed in a 40 km wide belt (fig. 1) and the underlying Middle Proterozoic granitic basement is present in the small, but tectonically significant, Lincoln massif (fig. 2). The rocks of the foreland consist of a basal rift-clastic sequence overlain by carbonate and siliciclastic rocks of the platform whereas the rocks of the hinterland consist of metamorphosed pelitic, psammitic, and mafic volcanic rocks of the slope-rise and outer rise sequence (fig. 3). Major west-directed thrust faults cut the section and become more numerous to the east. The metamorphic grade ranges from essentially unmetamorphosed, although well lithified and cleaved, sedimentary platform rocks of the foreland to kyanite-chloritoid grade in the hinterland. Temperature estimates range from 200<sup>o</sup> C or lower along the Champlain thrust to nearly 500<sup>o</sup> C in the central part of the hinterland (fig. 2). Using these temperatures and the preserved fabrics along many of the faults, deformation is estimated to have occurred at near-surface conditions in the western foreland (2.5 km or less) and at much greater depths in the hinterland (17-20 km) as a result of stacked and highly deformed thrust sheets (fig. 4; Strehle and Stanley, 1986). Based on regional considerations (Stanley and Ratcliffe, 1985) and available isotopic-age analysis, deformation and metamorphism throughout the western (foreland) and central (accretionary complex) part of the belt largely occurred during the Taconian orogeny (Middle Ordovician). Younger Acadian deformation (Middle Devonian) has severely deformed the Ordovician Moretown and Cram Hill Formations (fore-arc sequence) as well as the Silurian and Devonian section. The Acadian deformation decreases in intensity westward into the Taconian accretionary complex where its influence is thought to be minimal. During the Early Cretaceous alkalic dikes and normal faults cut the existing geology and were probably associated with abortive rifting during the opening of the North Atlantic (McHone, 1987; Stanley, 1980).

**WESTERN VERMONT** - The structure of the Vermont foreland is dominated by major, north-trending folds and imbricate thrust faults (Doll and others, 1961 and the western part of fig. 2). The largest of these, the Champlain thrust (stop 1), extends from southern Quebec to Albany, New York and places the older platform sequence over the younger foreland basin. Estimates of westward displacement range from 15 km. to 100 km (Stanley, 1987). Recent seismic traverses across western Vermont show that the Champlain thrust dips eastward at about 15 degrees beneath the hinterland of the Green Mountains and that many of the major folds of western Vermont are formed as duplexes and related structures on this and other faults in the foreland. Along the Green Mountain front several of the larger anticlines appear to be fault-propagation folds (Suppe, 1985; fig. 2). Seismic information suggests that some of the high-faults in the western foreland are older than the Champlain thrust whereas others are younger.

An eastern, but smaller Hinesburg thrust (stop 2), places transitional and rift clastic rocks over the platform sequence and as such forms a boundary between the foreland rocks of western Vermont and the hinterland. Dorsey and others (1983) have demonstrated that the Hinesburg thrust developed as a breakthrough thrust on the overturned limb of the Hinesburg nappe and therefore is quite different from the geometry of the Champlain thrust. Westward displacement is estimated to be 6 km.

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HINTERLAND - In central Vermont the earlier mapping by Cady and others (1962) and Osberg (1952) have been very helpful because it has defined many of the formations of the pre-Silurian hinterland and, as such, has highlighted the areas that must be reexamined in light of plate tectonic theory and recent findings in other hinterlands of the world. Furthermore, the earlier petrologic and isotope work of Albee (1965, 1968), Garlick and Epstein (1967), Cady (1969), Laird and Albee (1981), Lamphere and others (1983), and Laird and others (1984) is particularly important in defining metamorphic facies and establishing the physical conditions and timing of metamorphism and deformation. Much isotopic-age analyses, however, must be done in order to clarify the relative importance of the Acadian orogeny and understanding the thermal history of this part of the orogen (Sutter and others, 1985).

Prior to the current work of Stanley and his graduate students, the formations in the Vermont hinterland were considered to be a coherent, depositional sequence, although considerable internal flattening and shearing were recognized (Doll and others, 1961). As a result of recent mapping in western Massachusetts (Zen and others, 1983) and detailed mapping in north central Vermont (Stanley and others, 1984) ductile faults were recognized as an important, if not, a fundamental feature of the hinterland. The work in central Vermont has further shown that the faults occurred at different times during the evolution and metamorphism of the orogen (Table 1). Perhaps the most important of these are the early or premetamorphic faults because they control the distribution of major rock types and have been transported the farthest. The older high-pressure metamorphic series recognized by Laird (1984, 1987; fig. 4) in the mafic schists of the hinterland (fig. 2) formed either during or after these early thrust faults. As a result, the only reliable criterion for these faults is the truncation of units on either the footwall or the hangingwall. Later synmetamorphic and late metamorphic faults offset these older thrusts. Displacement criteria for many of these faults indicate east-over-west motion.

Although each of the thrust slices have distinctive lithic assemblages, there are sufficient similarities between them to establish their relative position (fig. 5). Thus the thrust slices directly east of the Lincoln massif are dominated by metawacke, metasandstone, and garnet muscovite schist whereas the thrust slices farther east are characterized either by noncarbonaceous, albitic-quartz schists, aluminous schists, or carbonaceous schist. Mafic volcanic rocks are only abundant in the western metawacke sequence or in the eastern albitic sequence. These lithologies are thought to represent the rift-drift sequence that formed along the western margin of Iapetus during the Late Proterozoic and Cambrian (fig. 5).

Comparison of the pre-Ordovician sequence in the Taconic allochthons (fig. 3) with the hinterland sequence in central Vermont suggests that the northern extension of the Taconic root zone is located to the east of the metawacke-rich sequence and to the west of the albitic-rich sequence (fig. 2). This position is further supported by the geochemistry of the mafic rocks as described by Coish (1987, 1988) and Ratcliffe (1987). As shown in figure 6 the major and REE of the Taconic metabasalts are more similar to the mafic rocks of Zone 2 than they are to either Zone 3 or 4. The relative positions of these zones are shown in figures 2 and 5. Because the Taconic allochthons were emplaced prior to any substantial metamorphism, they would correlate with either the early thrust faults recognized in the Vermont hinterland or perhaps still earlier faults whose record is basically destroyed. Thus the fault zone marked as the "Root Zone for the Taconic Allochthons" on figure 2 has had a very long and complex history.

Because of the required brevity of the trip text, handouts for each stop will be available to the participants free of charge. The 2.5 day field conference will attempt to cover the full transect from the western foreland to the Taconian unconformity along the base of the Silurian-Devonian section to the east. Day 1 (Friday) will be in the foreland and transitional zone. Day 2 (Saturday) will start along the Taconian unconformity and progress westward toward the heel of the Taconian accretionary complex. Day 3 will focus on the accretionary complex before our return to Boston.

*Stanley, Martin, and Coish*

## THE FORELAND AND TRANSITIONAL ZONE

### ITINERARY

The first two stops are located on South Hero Island, the southern-most island of the larger islands in northwestern Vermont.

**Stop 1 LESSOR'S QUARRY** (Directly south off of Sunset View Road) - This quarry is located in the fossiliferous Glens Falls Limestone of Middle Ordovician age. The Glens Falls contains excellent example of graded beds of fragmented fossil material representing distinct pulses of carbonate sedimentation possibly caused by storms. Study the stratigraphy of the larger blocks.

The quarry contains a number of imbricate bedding plane thrusts which can be studied by mapping all the walls. The most conspicuous thrust, called the Lessor's Quarry thrust, is well exposed on the north wall. Conspicuous slickenlines, east-dipping fault-zone cleavage, and bent S1 cleavage indicate that the upper plate moved to the west-northwest over the lower plate. The amount of displacement is unknown. A conspicuous syncline forms the lower plate on the east side of the north wall. This syncline folds an older bedding-plane thrust (blind thrust) which dies out in small faults and pressure solution features on the western limb of the syncline. Multiple cleavages along the blind thrust indicate that it moved westward before it was folded by the syncline. The syncline and its eastern anticline probably developed over a deeper ramp. You will note that the Lessor's Quarry thrust is not folded by the syncline or its eastern anticline, but instead truncates the blind thrust. We believe that the Lessor's Quarry thrust is an excellent example of an out-of-sequence thrust in the foreland.

The S1 cleavage is a superb example of pressure solution. You will note that it is discontinuous, filled with foliated, black clay-like material, and off sets bedding where the bedding-cleavage angle is less than 90 degrees. Furthermore, fossils are truncated by the cleavage. The calcite from the pressure solution was largely deposited in fractures and along fault surfaces.

**Stop 2** (Outcrops on Rt 2 1 mile east of Lessor's Quarry). "THE BEAM" This is a superb outcrop that serves as a field laboratory for research and teaching of foreland deformation. Use cameras but not hammers. Read **Stanley, R.S., 1990, The evolution of mesoscopic imbricate thrust faults - an example from the Vermont foreland: Journal of Structural Geology, v. 12, p. 227-242.**

The outcrop is located in the Cumberland Head Formation (Middle Ordovician) 5 miles west of the exposed front of the Champlain thrust fault or approximately 4600 feet (1400 meters) below the restored westward projection of the thrust surface. The major questions that will be discussed are: 1. How do ramp faults form?, 2. Are there criteria to determine if imbricate thrust faults develop toward the foreland (west) or the hinterland (east)?, 3. What is the relation between faulting and cleavage development?, 4. What processes are involved in the formation of the fault zones?, 5. Are there criteria that indicate the relative importance and duration of motion along the fault zone?, 6. Is there evidence to suggest that abnormal pore pressure existed during faulting?, and finally 7. What is the structural evolution of the imbricate faults and cleavage formation? The first six questions will largely be addressed by evidence at the outcrop. The seventh question will be answered by palinspastically restoring the imbricated and cleaved sequence to its undeformed state.

Continue southward on Rt 2 through South Hero Village to Interstate 89 for a distance of approximately 10 miles (16 kilometers). Travel south approximately 10 miles (16 km) on Interstate 89 to exit 14 where you turn west on Williston Road (Rt 2). Travel west several miles until you reach the Burlington waterfront. Turn north on Rt 127 and travel 3 miles (4.8 km) to Burlington High School. Turn left onto the road at the light and proceed to the Episcopal Diocese of Vermont. Read **Stanley, R. S., 1987, The Champlain thrust, Lone Rock Point, Burlington, Vermont: Geological Society of America, Centennial Field Guide for the Northeast Section, p. 67-72** for further instructions and a discussion of the outcrop.

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**Stop 3 CHAMPLAIN THRUST FAULT AT LONE ROCK POINT.** This is by far the best exposure of the Champlain thrust fault in the northern Appalachians. Furthermore, it is one of the best exposures of a foreland thrust fault in the United States. Wave erosion of the weaker shale of the lower plate has exposed the fault zone for a distance of over 1 mile (1.6 km). Copies of the aforementioned paper will be distributed to participants.

Return south along Rt 127 to the junction with Rt 2 in Burlington. Travel east several miles to South Burlington where you will find the junction of Rt 2 and Rt 116. Travel south on Rt 116 approximately 10 miles (16 km) to the junction of Rt 116 and the Hinesburg-Shelburne Road. A 4-way stop light marks the junction. There is a small shopping center on the west side of the road. Turn east (left) onto Champlain Valley High School Road. Go past the high school and take the first left (north) at the 4-way junction onto Richmond Pond Road. Travel a mile up the road and park near the barn on the west side. Walk west down the dirt road about 0.25 miles (0.4 km) and take the dirt road heading north to the outcrop of the Hinesburg thrust fault which is located at the base of the wooded hill.

**Stop 4 HINESBURG THRUST FAULT.** This is the type locality of the Hinesburg thrust fault which, for central and north Vermont, is the major structural boundary between the carbonate-siliciclastic platforms of the foreland to the west and the highly deformed, and metamorphosed rift - drift sequence of the Taconian hinterland to the east. Westward displacement is in the order of 4 miles (6.4 km). Although the Lower Ordovician carbonate rocks of the lower plate are poorly exposed here, the upper plate of argillaceous quartzite of the Lower Cambrian Cheshire Formation and stratigraphic lower Fairfield Pond Phyllite form the cliffs along the western front of the hill. These rocks contain many such fault related structures as deformed extension fractures, isoclinal folds, shear bands, mylonitic textures, slickenlines, and pressure fringes around pyrite. Many of these have been analyzed by Stanley and his students (Gillespie and others, 1972; Strehle, 1985; and Stehle and Stanley, 1986). Currently, Stanley, Martin and Smith are completing a study on the "z" shaped extension fractures. The results of this and earlier work will be discussed.

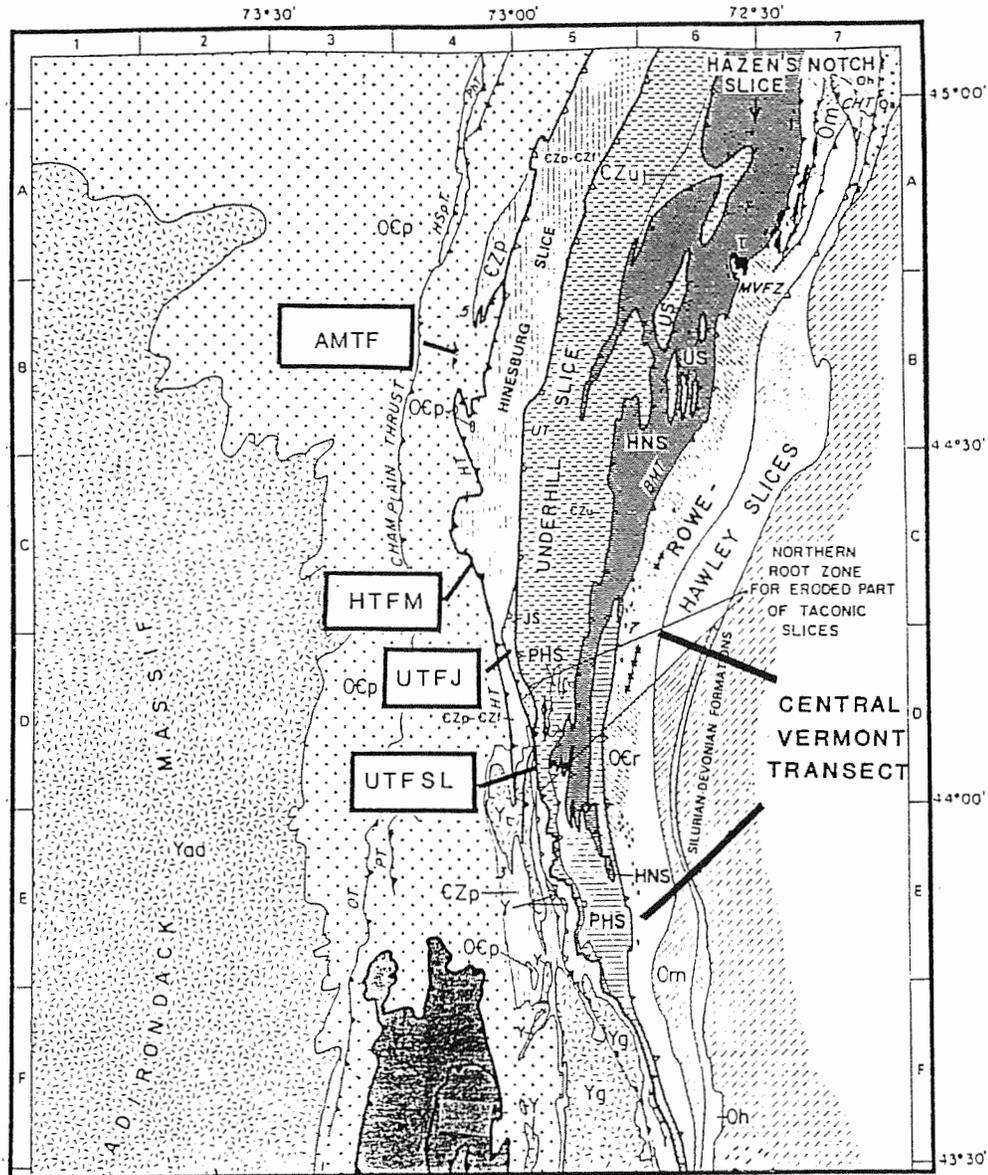
Return to the junction of Rt 116 and Champlain Valley High School Road and turn south. Travel through the village of Hinesburg 16 miles (26 km) to the junction of the New Haven River and Rt 116 just north of the village of Bristol. Turn east onto the Lincoln Gap Road.

Continue east onto the Lincoln Gap Road and travel 3 miles (4.8 km) to the village of Lincoln. Continue east through Lincoln to the second bridge over the New Haven River. Do not go up the Lincoln Gap Road, but turn south on Main Road which follows the New Haven River. Continue south to where the road changes to dirt paving. Keep to the left at the junction and travel several miles south to South Lincoln where 3 roads join near several houses. The bridge to the east is the Bridge at South Lincoln where our next stop is located.

**Stop 5 EASTERN CONTACT OF THE LINCOLN MASSIF -** The South Lincoln thrust fault at the bridge at South Lincoln. This outcrop along the New Haven River contains a mylonitic sliver of Middle Proterozoic gneiss in thrust contact with biotite-chlorite schist of the Hoosac (Pinnacle) Formation. A thick conglomerate containing quartzite and granitic gneiss pebbles, cobbles, and boulders in a matrix of biotite-chlorite schist and metawacke overlies the gneiss. The clasts are identical to the Middle Proterozoic rocks in the Lincoln Massif. Late northeast-plunging folds deform the fault zone. These folds are in turn overprinted by late biotite that is randomly oriented across the dominant schistosity. Based on the fact that the feldspar in the mylonitic gneiss has been dynamically recrystallized and the surrounding rocks have been metamorphosed to the garnet grade the temperature is estimated to have been in the order of 435<sup>o</sup> C to 450<sup>o</sup> C or slightly higher (fig. 2). These temperatures would correspond to pressures in the order of 5 to 6 kbars (17 to 20 km) (Strehle and Stanley, 1986). This synmetamorphic fault zone with its sliver of basement is typical of the eastern margin of the Lincoln massif. In fact, about 10 km. to the south fault zones like this become so numerous that they essentially make up the eastern half of the Lincoln massif as it is shown on the Geological Map of Vermont (Doll and others, 1961).

Return on the dirt road directly west of the New Haven River to the junction of the Lincoln Gap Road. Stop by the bridge that leads back to Lincoln. Do not go up the Lincoln Gap Road.

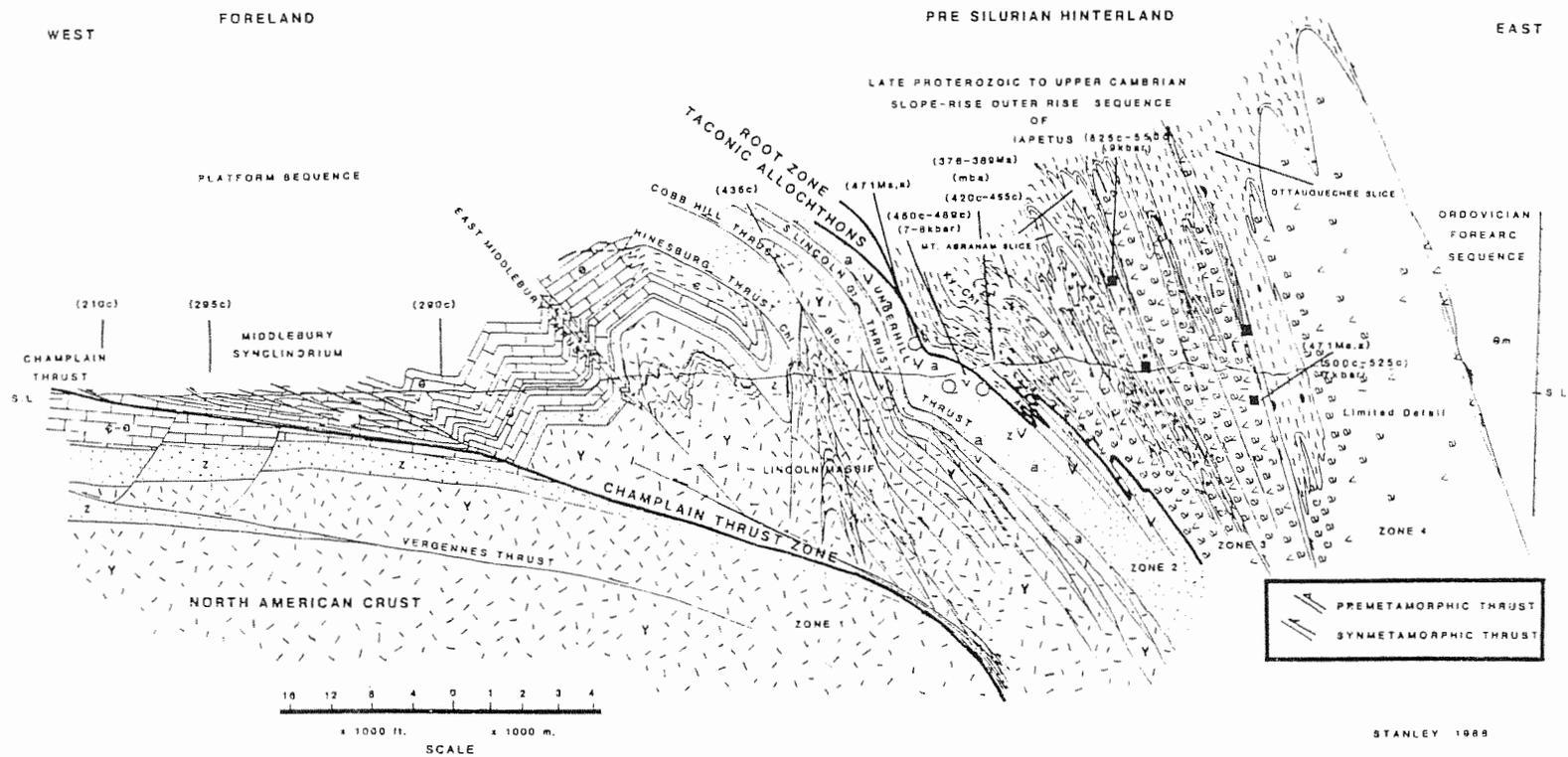
## Stanley, Martin, and Coish



**Figure 1** Interpretative Tectonic Map of Vermont and eastern New York showing the general location of the Arrowhead Mountain thrust fault (AMTF), the Hinesburg thrust fault at Mechanicsville (HTFM), and the Underhill thrust fault at Jerusalem (UTFJ), and South Lincoln (UTFSL). The geological map is taken from Stanley and Ratcliffe (1985, Pl. 1, figure 2a). Symbol T in A6 is a glaucophane locality at Tilliston Peak. Short line with x's (Worcester Mountains) and line with rhombos (Mount Grant) in C6 and D5 mark the Ordovician kyanite-chloritoid zones of Albee (1968). Widely-spaced diagonal lines in northcentral Vermont outline the region that contains medium-high pressure amphibolites described by Laird and Albee (1981b). Irregular black marks are ultramafic bodies. Open teeth of thrust fault symbols mark speculative thrust zones. The following symbols are generally listed from west to east. *Yad*, Middle Proterozoic of the Adirondack massif; *Yg*, Middle Proterozoic of the Green Mountain massif; *YL*, Middle Proterozoic of the Lincoln massif; *Y*, Middle Proterozoic between the Green Mountain massif and the Taconic slices, Vermont; *Ocp*, Cambrian and Ordovician rocks of the carbonate-siliciclastic platform; rift-clastic sequence of the Pinnacle (*Czp*) and Fairfield Pond Formations (*Czf*) and their equivalents on the east side of the Lincoln and Green Mountain massifs, *PHt*, Philipsburg thrust; *HSPt*, Highgate Springs thrust; *PT*, Pinnacle thrust; *OT*, Orwell thrust; *UT*, Underhill thrust; *HT*, Hinesburg thrust; *U*, ultramafic rocks; *CZu*, Underhill Formation; *CZuj*, Jay Peak Member of Underhill Formation; *OCr*, Rowe Schist; *Om*, Moretown Formation; *Oh*, Hawley Formation and its equivalents in Vermont; *JS*, Jerusalem slice; *US*, Underhill slice; *HNS*, Hazens Notch slice; *MVFZ*, Missisquoi Valley fault zone; *PHS*, Pinney Hollow slice; *BMT*, Belvidere Mountain thrust; *CHT*, Coburn Hill thrust; *Oa*, Ascot-Weedon sequence in grid location 7A.

FIGURE 2. Geological cross section through the pre-Silurian foreland and hinterland along a latitude of 44° N. The section is based on mapping by Cady (1945), Washington (1987), and Harding and Hartz (1987) in the foreland, by Tauvers (1982) and DelloRusso and Stanley (1986) in the Lincoln massif, and by Lapp and Stanley (1986), Prewitt (1986), Stanley (1986, 1987), and Kraus (1987) in the hinterland. North American Middle Proterozoic (Y) crust is shown by the random dash pattern. The overlying foreland rocks consist of Late Proterozoic (Z) rift clastic rocks (stippled pattern) and Cambrian and Ordovician platform rocks (carbonate and siliciclastic rocks overlain by shales) shown by the stacked rectangles. Symbols for the deformed and metamorphosed slope-rise and outer rise sequence correspond to those in figure 7. The sequence designations are those used in Stanley and Ratcliffe (1985) and figure 7. The black lenses in the hinterland sequence represent serpentinites. Zones 1 through 4 are based on mafic rock geochemistry (Coish, 1987, 1988). Temperatures of coexisting mineral assemblages are given as (i.e. 290°C) and are based on carbon and oxygen isotope analyses by Sheppard and Schwarcz (1970) for the foreland, oxygen isotope analyses by Garlick and Epstein (1967) and calcite-dolomite and amphibole-plagioclase temperatures (Laird and others, 1984) for the western part of the hinterland. The 471 Ma ages are based on  $^{40}\text{Ar}/^{39}\text{Ar}$  analyses of hornblende by Laird and others (1984). The 376-389 Ma ages are based on  $^{40}\text{Ar}/^{39}\text{Ar}$  total fusion ages of muscovite and biotite (Laird and others, 1984; Lanphere and others, 1983). The symbol a refers to amphibole, b refers to biotite, and m refers to muscovite. The metamorphic series assignment is based on amphibole analyses by Laird (1987). The solid squares represent hornblendes from the high-pressure metamorphic series and the open circles represent hornblendes from the medium-pressure metamorphic series (see fig. 7). The numbers 1 through 7 refer to field stops for the conference. The hinterland rocks, have been displaced westward along multiple generations of thrust faults and deformed by numerous generations of folds. The earliest recognized faults probably developed before or during the early high-pressure metamorphism (solid squares; fig. 7). These faults are deformed by reclined or sheath-like folds with

axes oriented parallel or nearly parallel to the pervasive southeast-plunging grain lineation. These folds and the older faults are in turn cut by synmetamorphic thrust faults with well-developed mylonitic fabrics. Based on stratigraphic and sedimentologic information, it is suggested that the root zone for the northern extension of the Taconic allochthons is located along a complex zone of pre- to late metamorphic faults that forms the western boundary of the albite-rich and aluminous rocks of the Hazens Notch, eastern Underhill and Mt. Abraham formations. This location is supported by the fact that the metabasalts in the Taconic allochthons are chemically more similar to the mafic rocks of the western Underhill and Pinnacle formations (Zone 2) than they are to the mafic rocks to the east in Zones 3 and 4. The exposed Middle Proterozoic rocks of the Lincoln massif consist of two major anticlines in which the eastern one has been severely flattened by numerous thrust faults which contain Paleozoic mylonites with east-over-west fabrics. A third anticline is buried beneath the platform sequence to the west of the East Middlebury thrust (Harding and Hartz, 1987). The breakthrough faults (East Middlebury thrust) and geometry of the western anticline suggests that the three anticlines of Middle Proterozoic rocks began their development as fault-propagation folds (Suppe, 1985) and evolved to flattened massifs and isolated slivers by the development of axial-surface cleavage and extensive high-angle faults. The available PT information suggests that this deformation occurs over temperatures that range from 290°C to the west to 435°C to the east. The temperatures would correspond to pressures in the range of 3 kbars (10 km) to the west and 4.5 to 5.0 kbars (16 to 18 km) to the east assuming a standard geothermal gradient of 20°C to 30°C per kilometer (Strehle and Stanley, 1986, fig. 5). This information again suggests that there was a significant tectonic load over the Lincoln massif when these structures formed. The minor imbricate faults of the Middlebury synclinorium are based on work by Washington (1987). The slice of North American crust floored by the Vergennes thrust is an interpretation based on some unpublished seismic reflection work. Shortening across this section is estimated to be between 400-500 km. of which approximately 70 km involved the folding and faulting of North American crust.



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Figure 2

EXPLANATION FOR LITHOTECTONIC CORRELATION CHART FOR CENTRAL VERMONT- The lithic symbols shown in each column are explained by geographic locality starting with the oldest unit.

West Central Vermont - Ymh- granitic gneiss of the Mt. Holly Complex, Y - mylonitic gneiss at South Lincoln, CZpbc - basal conglomerate in the Pinnacle Formation, CZpbg - biotite wacke, CZpfd - Forestdale Marble, CZpm - muscovite wacke, CZpcl - chlorite wacke and chlorite schist, CZhbc - basal conglomerate in the Hoosac Formation, CZhbg - biotite wacke, CZhg - schistose wacke, CZugl - quartzite laminated schist of the Underhill Formation, CZufg, well foliated wacke, CZu - mixed unit consisting of garnet, chlorite, muscovite schists, thin quartzite, CZugs - garnetiferous schist, CZfp - gray, finely laminated phyllite of the Fairfield Pond Formation, Cca - argillaceous quartzite of the Cheshire Formation, Ccm - massive, bedding quartzite of the Cheshire Formation, Cd - Dunham Dolomite, Cm - Monkton Quartzite, Cw - Winocski Dolomite, Cda - quartzite and dolostone of the Danby Formation, Cca - Clarendon Springs Dolomite, Lower Ordovician Beekmantown Group consists of limestone, dolostone and minor quartzite and shale of the Shelburne Formation (Da), Cutting Dolomite (Oc), Bascom Formation (Ob), and Chipman Formation (Ocb), Omm - Middlebury Limestone, Oo - Orwell Limestone, Ogf - Glens Falls Formation, Owl - fossiliferous limestone lenses near the base of the Walloomsac Formation, Oh - black carbonaceous slate and phyllite of the Hortonville Formation, Ow - dark gray, graphitic schist and phyllite of the Walloomsac Formation, Oww - Whipstock Breccia Member of the Walloomsac Formation.

Taconic Allochthone - Group 1 and 2 - CZnr, Rensselaer Graywacke Member of the Nassau Formation, CZnv - metabasalt and basaltic tuff, CZnm - lustrous, yellowish-green, purple laminated chloritoid - chlorite phyllite of the Mettawee Member, CZnt - Truthville Slates, CZnb - Bomoseen Graywacke Member, CZnzh - Zion Hill Quartzite, CZnmp Mud Pond Quartzite, Cwc - dark gray to black, pyritiferous and calcareous slate with thin sandy laminae of the West Castleton Formation, Chh - bluish-gray weathering black sulphidic slate and chert, Opo - white weathering, well laminated gray slate and chert, On - Normanskill Formation, Onnm - black slate and thin bedded cherts of the Mount Merino Member, Onr - red and green slate of the Indian River Member, Onag - Austin Glen Graywacke Member

Group 3 - CZgo, Light green to gray, white albite schist with some magnetite, chlorite granofels of the Greylock Schist, CZne - Netop Formation in the Dorset Mt. slice, CZgb - black of dark gray chloritoid or stilpnomelane albite, quartz knotted schist, lenses of feldspathic quartzite, conglomerate, and pink dolostone, CZg - light green, lustrous chloritoid phyllite and minor beds of white albite schist, CZac - St. Catherine Formation in the Dorset Mt. slice.

East Limb of the Berkshire massif - CZhab (CZh) - gray to white albite spotted schist of the Hoosac Formation, CZhgb (CZhga) - green albite magnetite schist, CZhg - green aluminous chloritoid schist. The symbols in parenthesis refer to designations of equivalent units on the Geologic Map of Massachusetts (Zen and others, 1983).

Green Mountain anticlinorium - Northfield Mountains (Many of the symbols are repeated from column to column. These symbols are explained in the first column that they appear in as the columns are read from left to right)

Lincoln Gap-Mt. Abraham-Hazens Notch slices - CZhn silvery white to dark green to black schist spotted with white albite, minor white to gray laminated quartzite of the Hazens Notch Formation. Mafic schist abundant in the eastern part, CZhnc - rusty weathering, black albite schist with widespread graphite and minor thin black or gray quartzite. CZhnc has been called the Battell Member of the Underhill (Doil and others, 1961) and the Granville Formation (Osberg, 1952). CZa - Mt. Abraham Schist. Silvery colored paragonite - muscovite - chloritoid - chlorite (garnet) schist with a distinctive pearly sheen on the schistosity, CZamg - similar to the main body of Mt. Abraham Schist but with abundant magnetite and chloritized garnet, CZage - Muscovite - chlorite schist of the Mt. Abraham Schist with large porphyroblasts of garnet and minor chloritoid,

Lincoln Gap-Pinney Hollow slices - Additional symbols are: CZhnca - rusty weathering, dark gray to black albite schist with discontinuous patches of graphite. Traceable into the Granville Formation. CZph - silvery green muscovite - chlorite - quartz schist of the Pinney Hollow Formation, CZpha light gray muscovite - chlorite - quartz schist with albite porphyroblasts, CZphg - mafic schists of the Pinney Hollow, CZphw - gray wacke with minor blue quartz, Co - black, pyritiferous and graphitic schist of the Ottauquechee Formation, Cobq - black and gray quartzite of the Ottauquechee Formation, Coqs - sandy quartzose schist of the Ottauquechee Formation, s - serpentinite and talc-carbonate rock.

Ottawaquechee - Stove slices - Additional symbols are: CZs - Stove Formation - silvery green, muscovite - chlorite - quartz schist identical to CZph, some schists are richer in chlorite, CZag - mafic schist in the Stove Formation.

Stowe-Moretown slices - Om, pinstriped schist and mafic schist of the Moretown Formation, Och - black, graphitic schist and thin quartzite of the Cram Hill Formation.



*Stanley, Martin, and Coish***Connecticut Valley -- Gaspé Trough**

**bcm** mafic igneous rocks of the Braintree complex: dark grey to black, salt and peppery, fine-grained hornblende diorite to gabbro confined to the western margin of the Braintree complex

**bcf** felsic igneous rocks of the Braintree complex: light cream color to white, biotite-muscovite granite that comprises approximately 90% of the Braintree complex

**DSn** Northfield Formation: very fine-grained, steel grey slate to phyllite with locally abundant pyrite cubes up to 1 cm. across; locally interbedded with thin brown to grey calcareous and non-calcareous metasilstone

**Sw** Waits River Formation: very fine-grained, steel grey slate and phyllite interbedded with brown or grey, calcareous or non-calcareous metasandstone and metasilstone

**Sspgc** Shaw Mountain Formation: phyllitic granule conglomerate with feldspar and blue-quartz porphyroclasts

**Ssqpc** white, polycrystalline quartz pebble conglomerate with clasts up to 2.5 cm. in diameter in a white to tan fine-grained quartzite matrix

**Ssqzt** very fine-grained, brown quartzite, locally calcareous

**Ssbl** fine-grained, brown, quartz-sand bearing limestone

**Ssqb** quartz-clast breccia in a fine-grained, blue or brown limestone matrix; clasts up to 5 cm. across

**Ssm y** white, gritty quartz mylonite

**Omgm?** grey-green to green quartz-feldspar metasandstone lithically identical to Omgm and found only in fault zones which cut the Taconian unconformity

**Ocrp** Cram Hill Formation: fine-grained, olive-grey phyllite thinly interbedded with fine-grained, quartz metasilstone; graded bedding occurs locally as do thick beds of black phyllite

**Ocrw** grey, feldspathic metawacke

**Ocrtm** fine-grained, non-calcareous, brown to tan metasilstone

**Ocrt** greenish-white, pyritiferous metatuff

**Ocra** greyish-white to greenish-white, greasy phyllite

**Ohbtm** Harlow Bridge Formation: fine-grained, massive to bedded, buff-tan to greenish-brown, non-calcareous, feldspar rich metasilstone

**Omgm** Moretown Formation: light green, fine-grained, bedded to massive quartz-feldspar metasandstone, metasilstone and phyllite

**Omtm** tan to brown, fine-grained, massive metasilstone lithically similar to Ohbtm

**Oma** greyish-white to greenish-white, greasy phyllite lithically similar to Ocra

**Omy** greenish-white, quartz-albite-chlorite-sericite protomylonite to mylonite and schist; quartz-feldspar laminae are separated by chlorite-sericite partings

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**STOP 6- WESTERN CONTACT OF THE EASTERN LINCOLN MASSIF** (second bridge east of Lincoln on the Lincoln Gap Road). This outcrop shows the western contact between the Middle Proterozoic rocks of the Eastern Lincoln massif and the overlying basal conglomerate of the Pinnacle Formation. Generally this contact is an erosional unconformity which dips steeply west or is slightly overturned with gentle north-plunging parasitic folds (DelloRusso and Stanley, 1986). Here, however, the contact is offset across a west-directed thrust fault (Cobb Hill thrust fault) and the associated folds in the cover have been rotated toward the transport direction. The Grenvillian foliation of the Middle Proterozoic rocks is progressively overprinted by the Taconian schistosity (K/Ar age of 410 Ma on biotite, Cady 1969) of the cover as the contact is approached from the basement. The basal conglomerate of the Pinnacle is generally a quartz cobble conglomerate with minor pebbles and cobbles of granitic rocks. Here, however, large granitic cobbles and boulders form lensoid deposits separated by quartz-feldspar metawacke. The origin of this deposit is controversial, but Stanley, following Tauvers (1982), will argue that they represent ancient channels that have been subsequently deformed into nearly reclined folds.

(Optional Stop) **STOP 7 - FAULTS ZONE ALONG THE TACONIC ROOT ZONE** (The Fire Road Fault Zone of Lapp and Stanley, 1986) - This fault zone is typical of the faults that mark the boundary between the Hazens Notch-Mt. Abraham thrust slices and the Underhill slices to the west. Two or three generations of thrust faults are commonly present in many of the

outcrops throughout this belt. The older pre- or early metamorphic faults are isoclinally folded and cut by syn- and late metamorphic thrusts with sharp, well foliated and lineated contacts. For example, at this outcrop the aluminous Mt. Abraham Schist (chloritoid-paragonite) overlies the carbonaceous schist along a synmetamorphic (post peak) thrust. Isoclinal folds of the aluminous schist in the carbonaceous schist indicate an earlier fault. A late to post metamorphic fault is marked by the offset of the large quartz vein. Return via Rt 116 to Burlington Vermont.

*End of field trip for Day 1.*

# Criteria for Pre-, Syn-, and Post- Metamorphic Faults

**PRE-METAMORPHIC FAULTS** Regional metamorphism occurs after movement. Earlier mineral assemblages are absent.

Criteria

1. Mappable lithic units are truncated on the hanging wall and foot wall of the fault zone.
2. Silvers of rocks foreign to either the hanging wall or foot wall (i.e. ultramafic rocks, felsic plutonic rocks, Proterozoic Y gneiss, exotic metasedimentary rocks).
3. Preserved fault zone fabrics are rare or very difficult to recognize.

**SYN-METAMORPHIC FAULTS** Metamorphic recrystallization occurs during faulting.

Criteria

1. Criteria 1 and 2 listed for pre-metamorphic faults.
2. Ductile fabrics preserved along fault zones.
  - a. Sharp lithic contacts that truncate internal fabric (i.e. older foliations, compositional layering, quartz veins).
  - b. Prominent anastomosing foliation is parallel to or is at a low angle to the fault surface.
  - c. Prominent lineation formed by elongate minerals or mineral clusters.
  - d. C-S fabrics.
  - e. Asymmetric folds, sheath folds, and/or reclined folds with sheared off limbs.
  - f. Mylonitic fabrics marked by fine grained, recrystallized minerals compared to larger grains in the surrounding rocks. Porphyroclasts may be present. Evidence for ductile flow is best seen in quartz and feldspar where undulose extinction, deformation twins, deformation lamellae, deformation bands, subgrains, polygonization and sutured grain contacts suggest dislocation glide, dislocation creep and grain boundary migration as important deformation mechanisms. Continued strain during recrystallization is marked by elongate new grains with evidence of internal deformation.
  - g. Absence of cataclastic fabrics.

**POST-METAMORPHIC FAULTS**

Movement occurs after regional metamorphism, but retrograde metamorphism may occur along the fault zone.

Criteria

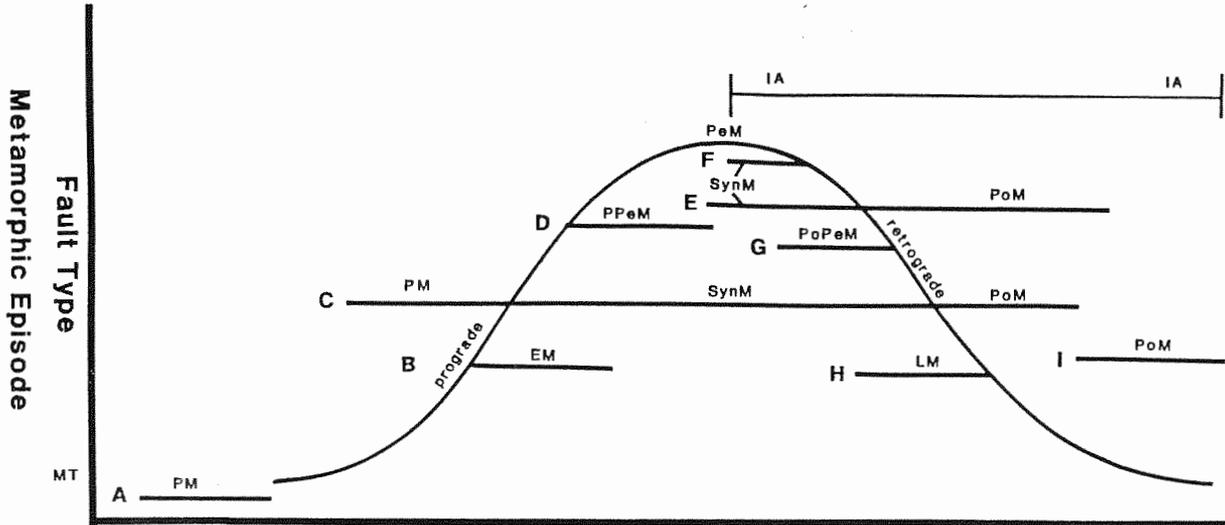
1. Criteria 1 and 2 listed for pre-metamorphic faults.
2. Strongly oriented, fault zone fabrics are well preserved and will depend upon the depth at which movement occurred. Brittle fabrics reflect movement within several kilometers of the surface whereas fabrics typical of the brittle-ductile transition reflect deeper levels.
  - a. Brittle-ductile fabrics consist of a mixture of minerals such as quartz and feldspar in which some of the minerals deform by ductile mechanisms (quartz) and others by fracture (feldspar). The overall mechanical behavior of the rock depends on the dominant mineral.
  - b. Retrograde minerals along fault zones. In pelitic rocks this may consist of fine grained mica (phyllosilicate). In gneiss the mica-rich layers are rich in white mica, chlorite, and/or biotite. Finely crushed feldspar may also be present.
  - c. Brittle fabrics consist of grains which are angular with random size and orientation. Microfractures are abundant.

**COMPOUND FAULTS**

Repeated movement occurs under different metamorphic conditions.

Criteria

1. Evidence for post-metamorphic faults superposed on older syn-metamorphic faults are readily preserved and easily recognized. All other examples of fault-superposition are unlikely to be recognized because of subsequent metamorphism.



Stanley 1987

**Time**

A simplified graph showing the schematic relations of eight representative fault types and a single metamorphic episode in the hinterland. PM represents pre-metamorphic faults, EM represents early metamorphic faults, PPeM represents pre-peak metamorphic faults, SynM represents syn-metamorphic faults, PoPeM represents post-peak metamorphic faults, LM represents late metamorphic faults, and PoM represents post-metamorphic faults. Faults C and E represent faults with a composite history. The distinction between "pre-metamorphic" and "early metamorphic" faults is commonly difficult to determine because metamorphism obliterates much, if not all, of the fabric information. The criteria that are listed under "pre-metamorphic faults" would apply therefore to "early metamorphic" faults although some fabric information might be preserved as inclusions in porphyroblasts. The distinction between "early metamorphic" and "pre-peak metamorphic" faults as well as "post-peak metamorphic" and "late metamorphic" faults is a matter of degree and subjective evaluation of the fault-zone fabric and mineralogy. All faults are distinguished by criteria 1 and 2 for pre-metamorphic faults. Oriented, fault-zone fabrics are preserved for those faults that were active during and after the peak of metamorphism and may well apply to "pre-peak metamorphic" faults. Ductile fabrics are formed during the metamorphic episode, while brittle fabrics develop before and after metamorphism. Transitional fabrics (brittle-ductile) are formed in the region where the fault crosses the metamorphic curve after the metamorphic peak has been achieved. PeM represents the peak of metamorphic intensity. MT represents the "metamorphic threshold" IA represents the range of possible isotopic ages associated with the metamorphic episode.

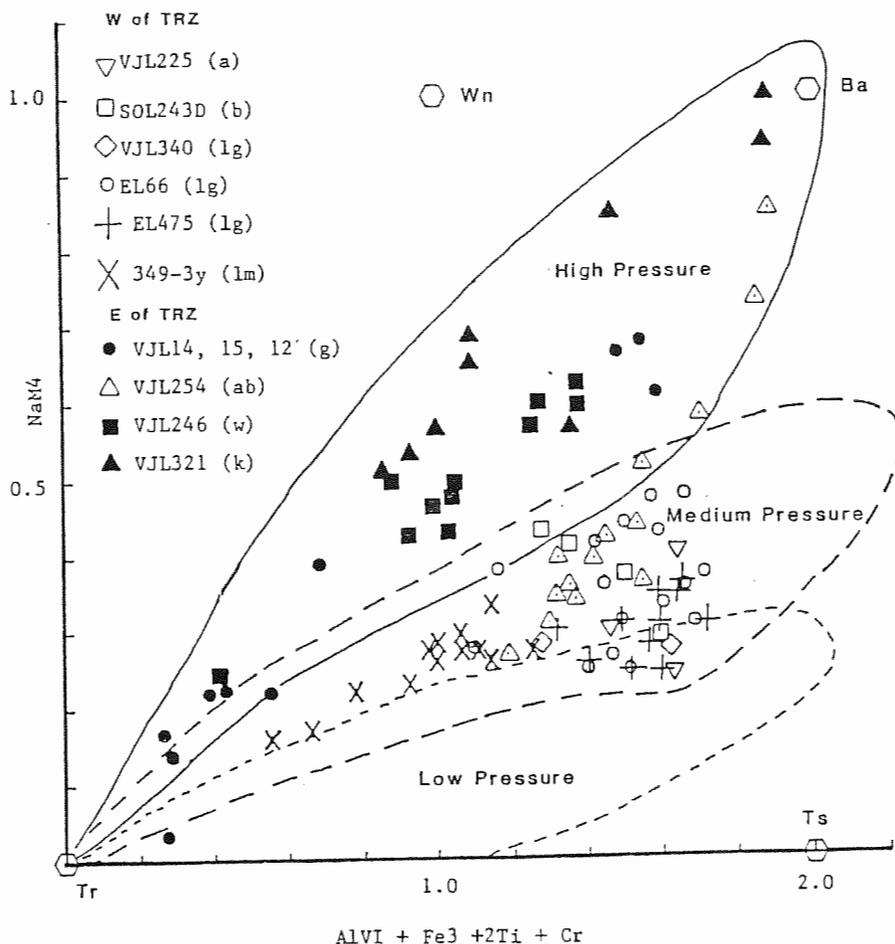


FIGURE 4.

Formula proportion  $\text{NaM}_4$  versus  $(\text{AlVI} + \text{Fe}^3 + 2\text{Ti} + \text{Cr})$  for amphibole - chlorite - epidote - plagioclase - quartz schist from central Vermont. Sample localities identified by VJL numbers are from Laird and others (1984). Localities in ( ) are from areas west (W) of in the inferred Taconic Root Zone (TRZ) or from areas east (E) of the Taconic Root Zone as shown in figure 2. Areas of high-, medium-, low-pressure facies series metamorphism are from Laird and others (1984). Increasing temperatures of metamorphism results in increasing advancement along the X and Y axes. Tr (tremolite), Ts (tschermakite), Wn (winchite), BA (barroisite). Analyses are electron microprobe data and normalized to total cations in which  $(\text{Na} + \text{K}) = 13$ . The mafic volcanic rocks from east of the Taconic Root Zone have core compositions indicating higher metamorphic grade than rim compositions. For example, at Granville Gulf (VJL 12, 14, 15), where the 471 Ma age has been obtained (fig. 2), actinolite overgrew barroisite. Because the actinolite rims add to the total amphibole in the rock, they record a second, but lower temperature and pressure metamorphism rather than an simple alteration of an earlier barroisite during cooling. In some of the samples west of the TRZ (VJL 225 and 340) the zoning is continuous and indicates progressive metamorphism whereas samples from the Lincoln massif (349-3y) farther to the west record decreasing temperature with time with the original temperature being less than the overlying cover to the east. The zoning in the other samples west of the TRZ is more difficult to interpret because it is irregular and may be related to miscibility. This figure is taken from Laird (1987).

**Figure 5** Simplified interpretative diagram of the ancient passive margin of North American with Iapetus during the Late Proterozoic through the Upper Cambrian. The diagram was modified from the model of a lower-plate margin as described by Lister and others (1986). This particular model is preferred over other models because it allows for a wide passive margin with relatively thin but somewhat different stratigraphic sequences deposited in various sedimentary depressions or basins which can be separated by relatively thin stratigraphic sequences. Restoration of the Taconic allochthons to their original depositional site east of the Green Mountain and Lincoln massifs and retrodeformation of the resulting Taconian hinterland suggests that the margin was at least 350 to 400 km wide, of which the carbonate-siliciclastic platform comprised at least 60 km of the total figure (see also Stanley and Ratcliffe, 1985, plate 2). This range is comparable to the Australian passive margin southeast of the Timor trough in the Indonesian convergent arc system. The present stratigraphic thickness of the Taconic allochthons has been estimated to be in the order of .6 km to 1.8 km. (Potter, 1972; Rowley and others, 1979). The original thickness of the slope-rise sequence was probably in the order of 4 km considering the fact that the preserved section is incomplete and has been dewatered and compacted. This value, although small, is comparable to the stratigraphic thickness along modern passive margins where the margin is wide and continental crust is being subducted beneath oceanic crust. For example, the slope-rise sediments along the Asian continental margin west of Taiwan are at least 7 km thick (Suppe, 1981) whereas those in the Browse Basin on the Australian passive margin southeast of Timor are at least 5 km thick (Hamilton, 1978). Although many of the rocks in the Taconic allochthons are lithically similar to rocks in the Taconian hinterland, the differences, particularly in the lower rift-clastic sequence, are significant and suggest deposition in somewhat isolated basins (Ratcliffe, 1987). For example, the rift clastic sequence in the Taconics allochthons is rich in metawacke, metasilstone, and metasandstone whereas many of the correlative rocks in the Taconian hinterland are rich in albitic schist and relative poor in metawacke and metasandstone. During subsequent deformation, which is considered to be largely Taconian (Middle Ordovician), the passive margin sequence was shortened by at least 80 percent to produce the geological section shown in the Central Vermont Transect (fig. 2). Lithic symbols for the sedimentary cover are the following: the platform sequence, shown by the rectangle and dot pattern, consists of carbonate and siliciclastic rocks of the Cheshire, Dunham, Monkton, Winooski, Danby, Clarendon Springs and equivalent formations. The slope-rise and outer rise sequence consists of four groups of rocks each of which is characterized by a dominant set of rocks which is considered distinctive for that group. Contacts between the groups are gradational. The basal sequence consists of two groups. The stippled pattern represents metawacke, metaconglomerate, metasandstone, metasilstone and quartz-rich pelitic rocks of the Pinnacle and Underhill (western part) Formations in central Vermont, the Rensselaer Graywacke Member and Bomoseen Graywacke Member of the Nassau Formation in the Taconic allochthons (group 1 and 2) and minor, but mappable, units in the Hoosac, Underhill (eastern part), Hazen Notch and Pinney Hollow Formations in central Vermont. The pattern with random "a" symbols are largely quartz-bearing pelitic rocks with distinctive porphyroblasts of albite typical of such formations as the Greylock Schist and the Netop Formation in the Taconic allochthons (group 3), the Hoosac Formation of western Massachusetts and southern Vermont, the Underhill (eastern part), Hazens Notch, Pinney Hollow, and Stowe in central Vermont.

Albitic schists are also present in the Hoosac and western part of the Underhill Formations in central Vermont. The basal two groups are characterized by mafic volcanics of basaltic composition (v symbol). These rocks are commonly absent from the upper two groups. The blank pattern of the central group consists of aluminous pelitic rocks which generally contain chloritoid at the appropriate metamorphic grade. Metasilstone and metasandstone are minor. Rocks of this group are found in the Nassau Formation (Mettawee Member), the Everett Schist, the Greylock Schist, and the St. Catherine Formation in the Taconic allochthons, the Hoosac Formation on the east limb of the Berkshire massif, and the Mt. Abraham Schist in central Vermont. The Fairfield Pond Formation of central and northern Vermont is included in this group although it does not contain chloritoid nor does it appear to be as aluminous as the others. The wavy dashed pattern of the upper group consists of dark colored, rusty or brown weathering, carbonaceous pelitic rocks that contain interlayered grey, black, or white metasandstone and metasilstone. Thin carbonate beds and limestone conglomerates are present in the Taconic allochthons (group 1) and the western part of the group. Rocks of this group are found in the Skeels Corner and Sweetsburg Formations of Vermont and Quebec, the West Castleton and Hatch Hill Formations of the Taconic allochthons (Group 1), the Hazen Notch, Battell, Granville, and Ottauquechee Formations of central Vermont. Mafic volcanic rocks of basaltic composition are rare in the upper two groups. The rocks in each of the aforementioned groups provide a basis for interpreting the rift-drift sequence for the Late Proterozoic to Upper Cambrian margin. The rocks of the basal groups and their associated basaltic volcanics represent the clastic rocks deposited in basins and depressions during the rifting of the ancient margin. The presence of granitic clasts, blue quartz, and abundant secondary albite in many of these rocks supports this interpretation. The minor and trace element geochemistry of the associated mafic volcanics systematically vary from west to east (Coish, 1987; Coish and Sinton, 1988) and also support the rift interpretation. All the rocks have basaltic compositions. Those to the west (Zone 1) are transitional to mildly alkaline basalts with high concentrations of Ti, P, Zr, Y and LREE-enriched patterns and are similar to continental rift basalts whereas the volcanics to the east (Zone 4) are tholeiitic basalts with low concentrations of Ti, Zr, P and LREE-depleted patterns and are similar to ocean ridge basalts. The intervening two zones are transitional between Zone 1 and 4. For example, Zone 2 mafic rocks in the Pinnacle and western Underhill formations are similar to Zone 1 except they have lower P and slightly higher Ti contents. The mafic volcanic rocks in the late Proterozoic to Early Cambrian Nassau Formation of the Taconic allochthons have many of the chemical characteristics of Zone 2 volcanics of central and northern Vermont (Ratcliffe, 1987). Support for this comparison is discussed in figure 8. These chemical similarities therefore support the relative palinspastic position of the Taconic allochthons within or slightly east of Zone 2. Zone 3 mafic volcanics have a mixed chemistry. Some of the volcanics are like those of Zone 1 and 2, whereas other bodies have lower concentrations of Ti, Zr, P, and LREE. The dark colored, carbonaceous rocks in the upper group are generally devoid of volcanic rocks and were deposited seaward of the carbonate-siliciclastic platform during thermal subsidence. The presence of graphite throughout many of these rocks indicates deposition under reducing conditions which were extensive when the polar icecaps were greatly reduced during the lower Paleozoic (Berry and Wilde, 1978). The aluminous rocks of the middle group are interpreted to be the transition from rift to drift conditions.

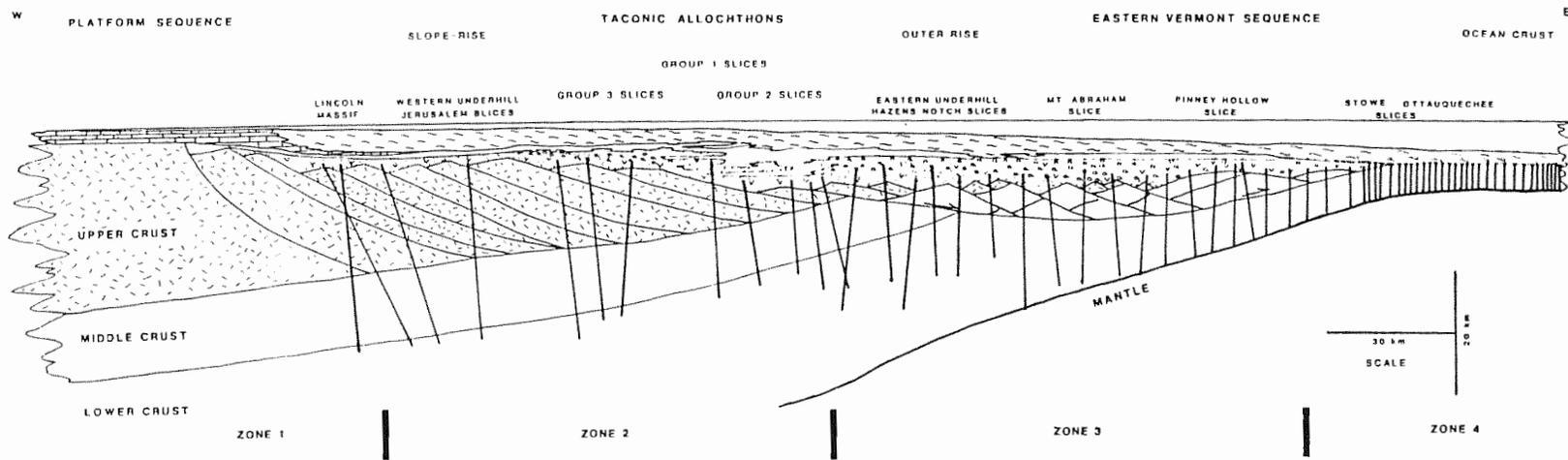


FIGURE 5

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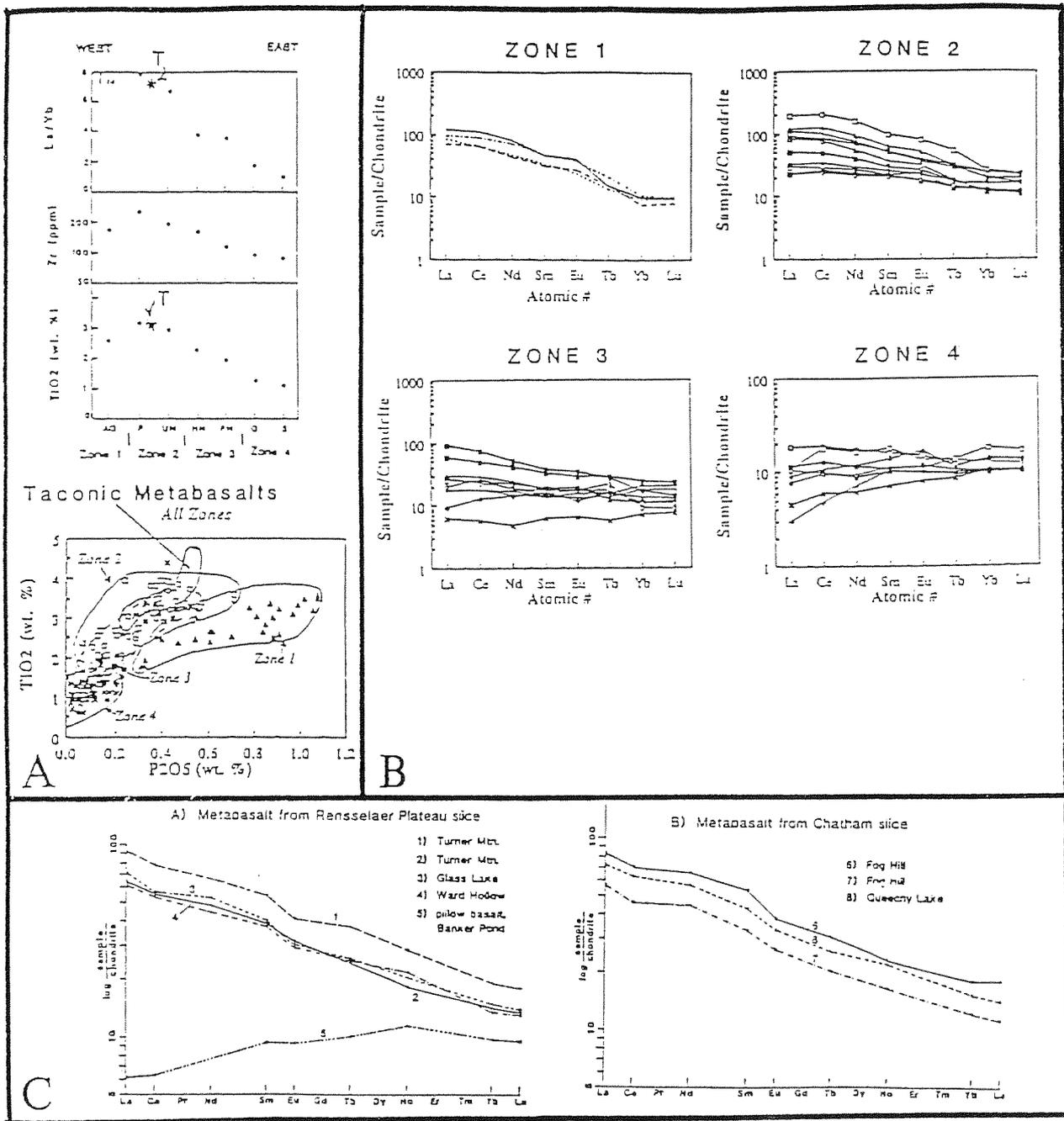


Figure 6 Geochemistry of mafic rocks in central Vermont (Coish, 1987 and 1988) and in the Rensselaer Plateau and Chatham slices of the Taconic allochthons (Ratcliffe, 1987). Diagrams in A show the average values of TiO<sub>2</sub>, Zr, and La/Yb (a measure of LREE enrichment) and the variation in TiO<sub>2</sub> vs P<sub>2</sub>O<sub>5</sub> (lower graph) for mafic rocks of basaltic composition from Zone 1 through Zone 4 from Vermont. The symbol "T" in the upper graph represents the average values for TiO<sub>2</sub> and La/Yb as given by Ratcliffe (1987, table 1 and fig. 6B). The values for TiO<sub>2</sub> and P<sub>2</sub>O<sub>5</sub> are also plotted for the Taconic metabasalts. They show a distribution that overlaps with Zone 2 and Zone 3 rocks. Graphs in B show the distribution of LREE from mafic rocks from Zone 1 through Zone 4 in Vermont. Zone 1 is characterized by enriched LREE. Samples from Zone 2 show a general pattern that is similar to Zone 1. Several samples, however, are less enriched in LREE compared to the HREE. Zone 3 samples are similar to Zone 1 and 2 in that they are high in REE but their patterns are generally much flatter than the other two. The group as a whole shows patterns that are similar to several samples in Zone 2 and 4. Zone 4 samples have lower concentrations of REE than the other zones and either show flat patterns or are slightly or moderately depleted in LREE compared to the HREE. They are therefore quite different from all the mafic rocks in Zone 2 and many of the mafic rocks in Zone 3. The graphs in C show the REE distribution for 8 metabasalts from the Nassau Formation in the Taconic allochthons (Ratcliffe, 1987). These patterns are similar to Zone 2 mafic rocks in central and northern Vermont. Graphs in A, B, and C indicate that the Taconic metabasalts are quite similar to Zone 2 mafic rocks although they do show similarities to some samples from Zone 3.

## THE FORE-ARC SEQUENCE

Detailed 1:12,000 scale mapping from the Silurian-Devonian sequence westward across the Taconian unconformity into the heel of the Taconian accretionary complex has divided the belt into 3 major lithotectonic belts: a) the heel of the Taconian accretionary complex (pre-Ordovician Stowe Formation), b) the Moretown through Cram Hill Formations (Ordovician fore-arc sequence), and c) the fossiliferous Silurian-Devonian sequence (fig. 1). There are many differences lithically and structurally between rocks of the Stowe Formation and the eastern sequences. This contact is a regionally important fault zone designated the Roxbury mylonite zone (fig. 1). The purpose of this part of the field trip is discuss the differences in the depositional and structural histories of these rocks. The field trip starts east of (stratigraphically above) the Taconian unconformity in rocks deformed during the Acadian Orogeny because the unconformity is the important reference surface that provides constraints on the age of structures to the west.

East of the Roxbury mylonite zone in both the fore-arc and Silurian-Devonian sequences, the penetrative dominant schistosity ( $S_n$ ) deforms bedding, is the oldest fabric of 3 recognized in these rocks and defines a well developed, upright schistosity fan which dips steeply east or west. The axis of the schistosity fan marks the eastern margin of the Moretown Formation (fig. 2). The dominant schistosity is axial planar to high amplitude ( $\approx$  1500 feet), isoclinal, reclined folds (Stops 1 & 4) of which the West Brookfield syncline is an example (figs. 1 & 2). The Moretown-Cram Hill sequence and overlying Silurian-Devonian sequences are dissected by numerous faults which cut across the Taconian unconformity. Distances of displacement on these faults are difficult to calculate because of numerous faults and insufficient exposure to trace key marker beds. Displacement on individual faults is probably less than 50 meters but taken together they have thickened the section significantly. The Roxbury mylonite zone is the only fault of major lithotectonic importance in this part of central Vermont (fig. correlation chart).

Three generations of structural elements observed in the Silurian-Devonian sequence have been traced westward, across the Taconian unconformity into the Moretown Formation without any change in orientation or style. Bedding is readily recognized in chlorite and feldspar rich meta-sandstones, meta-siltstones, and grey to black phyllites as the oldest planar element (Stop 1 & 4). Furthermore,  $L_n$  mineral stretching lineations on  $S_n$  surfaces plunge steeply to the northeast or northwest depending on their geographic position within the schistosity fan (Stop 4). These fabrics are Acadian in age. There is no evidence for an older (Taconian) fabric in the bedded Moretown-Cram Hill sequence. Very few greenstone dikes (2) and no serpentinites are observed. In addition, these rocks have not experienced greater than lower greenschist facies metamorphism.

The dominant Acadian fabric ( $S_n$ ) of the Ordovician Moretown-Cram Hill sequence has been traced across the Roxbury mylonite zone into the Stowe Formation without any change in style or orientation (Stops 5 & 6). Therefore, along the eastern margin of the Stowe Formation, the dominant fabric is also Acadian in age. To the west, in the Taconian accretionary complex there are fundamental characteristics which set these rocks apart both depositionally and structurally from those of the Ordovician fore-arc sequence. Bedding is not recognized in these middle to upper greenschist facies schists and greenstones. An older (Taconian) fabric ( $N-1$ ) is recognized as interfolial lithic hooks. Fault bounded greenstones and serpentinites are numerous. An even mixture of southeast and northeast plunging mineral stretching lineations is observed. Mineral stretching lineations become predominantly southeast plunging as they are traced over the axis of the Green Mountain Anticlinorium and into the Lincoln Massif. Thus, southeast plunging mineral stretching lineations are a characteristic of Taconian deformation while north plunging mineral stretching lineations are characteristic of Acadian deformation. Table 2 summarizes the lithic and structural differences across the Roxbury mylonite zone.

These differences suggest that rocks west of the Roxbury mylonite zone was subjected to an older (Taconian) period of middle to upper greenschist facies metamorphism while those to the east were only metamorphosed to the lower greenschist facies possibly during the Acadian orogeny. A still older medium to medium - high pressure metamorphism is recognized in the Pinney Hollow and Hazen Notch greenstones

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accretionary complex to the west (Laird and Albee, 1981; Laird and others, 1984; Laird, 1987). The bedded sequence of the Ordovician Moretown-Cram Hill belt is thus possibly a syn- to post-Taconian/pre-Acadian depositional sequence. It is hypothesized that the Moretown unconformably overlies the Stowe Formation although this interpretation has not been fully investigated yet. Subsequent penetrative Acadian deformation resulted in a well developed schistosity in rocks east of the Roxbury mylonite zone and transposed the older Taconian fabric in rocks directly west of the Roxbury mylonite zone. Because of the mechanical contrast between the older, recrystallized rocks of the Stowe Formation and the overlying sedimentary sequence of the Moretown-Cram Hill belt, the overlying cover was detached from the underlying basement along the Roxbury mylonite zone (fig. 2).

The heel of the Taconian accretionary complex acted as a rigid buttress against which the fore-arc and Silurian-Devonian sequences were compressed. It is likely that structures in the "Acadian foreland" (fore-arc sequence) region were originally shallowly east dipping (dipping toward the "Acadian hinterland"). Subsequent rotation into their current steeply dipping attitudes occurred during uplift resulting in development of the large Acadian schistosity fan (figs. 3). The majority of porphyroclasts along fault zones are symmetrical suggesting that flattening was final but significant phase of deformation.

Stanley and Ratcliffe (1985) have interpreted the Moretown and Cobble Mountain Formations of western Massachusetts to be fore-arc basin fill based on its mineralogy (feldspar rich), grain size and bedded nature. No evidence has been found in the Roxbury, Vermont area to negate this hypothesis although other hypotheses are still viable (i.e. back-arc basin). Preliminary point counts on these meta-sedimentary rocks indicate a quartz:feldspar ratio of 2:1 with all of the feldspar being untwinned, unstainable albite. These conclusions are consistent with conclusion reached in Quebec (Tremblay, 1991).

*Table 2*

*Comparison of lithic and structural characteristics across the Roxbury mylonite zone.*

*West of Roxbury mylonite zone*

Bedding not observed.

Schists

Middle to upper greenschist facies metamorphism (biotite grade).

Numerous greenstones and serpentinites observed in fault contact with schist.

Mineral stretching lineations (chlorite/biotite) of mixed orientations: both northeast and southeast plunging; become entirely southeast plunging farther west.

*East of Roxbury mylonite zone*

Bedding readily observed; locally graded.

Meta-sandstones, siltstones and phyllite-slate.

Lower greenschist facies metamorphism (chlorite grade).

Large, fault bounded greenstones and/or serpentinites never observed; 2 very small greenstone dikes observed in a small outcrop along eastern margin of Roxbury mylonite zone.

Mineral stretching lineations (chlorite) plunge to the northeast or northwest.

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**Road Log:** Distances are measured from the intersection of rt. 12 and rt. 64. Take route 89 and get off at exit #5 (Williamstown/Rt 64 exit). Follow route 64 west. Park at the intersection of rt. 64 and rt. 12. The first outcrop is on the north side of rt. 64 approximately 150 feet from rt. 64.

*Total miles Distance in miles from  
last stop.*  
**0.0 0.0**

**Stop 1: NORTHFIELD FORMATION; ACADIAN FABRICS**

This is an outcrop of the Northfield Formation. At this location it is interbedded gray-black phyllite and brown siltstone. Two fabrics are separable in this outcrop:  $S_n$  and  $S_{n+1}$ .  $S_n$  is the penetrative schistosity while  $S_{n+1}$  is the spaced cleavage intersecting  $S_n$  at a low angle. If you cross rt. 64 and walk a short distance up the road (walk south east) so that you can look down on the outcrop, you will see a large isoclinal reclined fold (behind light pole). This is an  $F_n$  fold to which  $S_n$  is axial planar. Folds of this magnitude and larger are found throughout the Roxbury quadrangle and are typical of the Acadian deformation.

The fabric in this rock is Acadian because it is situated east of (above) the Taconian unconformity. This is the fabric which must be tracked into the heel of the Taconian accretionary wedge in order to evaluate the extent of Acadian penetrative deformation.

**Stop 2: SHAW MOUNTAIN FORMATION; ACADIAN FAULTS**

From stop 1 cross route 12 (west) and follow the winding dirt road (Lovers' Lane) to a turn out on the left side (south) of the road; park here.

**0.4 0.4**

The polycrystalline quartz pebble conglomerate of the Shaw mountain formation is exposed just inside the tree line along the south side of Lovers' Lane. Pebbles are relatively hard to spot because this outcrop has been sheared out. The fault zone is parallel to the dominant schistosity (weakly developed at this location).

The thin, discontinuous Shaw Mountain conglomerate unconformably overlies the Cram Hill Formation. The unconformity however, has been sheared out by numerous Acadian faults. Thus, the unconformity is rarely seen. It is observed as an extremely low-angle unconformity in a window through the West Brookfield syncline. At that location, the overlying conglomerate truncates bedding and contains phyllite chips of the underlying Cram Hill Formation.

**Stop 3: CRAM HILL FORMATION**

From stop 2 proceed to the end of Lovers' Lane and take a left turn onto rt. 12A. Follow rt. 12A for 0.5 miles where you will bear to the left onto a dirt road (Bull Run Road). Follow Bull Run road for 2.5 miles to the Diemer farm on the left side of the road. The farmer has allowed us to park along his road frontage but it is good idea to ask.

**3.4 3.0**

Across the road (west) is a small farm lane leading to a bridge. Enter the woods on the right before the bridge and climb down into stream bed. This is Bull Run. Follow Bull Run north to its intersection with a large stream flowing from the west (Ellis Brook). Climb up Ellis Brook for approx. 100 feet. At this location, dismembered interbedded grey phyllite and brown metasilstone of the Cram Hill Formation is exposed. While not graded at this locality, graded bedding has been observed fairly regularly within both the Cram Hill Formation and the Moretown Formation.

These rocks suggest that they were deposited as a distal (fine-grained) turbidite sequence. The overlying (locally calcareous) Shaw Mountain conglomerate and crinoidal limestones make a depositional interpretation difficult. I would suggest however that the Cram Hill Formation was deposited in a long, narrow basin between the Taconian accretionary complex and the island arc. With the onset of Acadian compression, the sediments of this

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basin were uplifted and eroded, hence the Taconian unconformity. With the onset of Acadian compression, subsidence of the basin occurred. The thin, discontinuous nature of the Shaw Mountain Formation suggests that a short period of shallow water carbonate deposition ensued. This implies that the rate of subsidence was quite high as the time line of the carbonate compensation was quite short. During this time, alluvial fans dumped pebbles from the wedge and the island arc into the carbonate system. The overlying phyllites of the Northfield and Waits River Formations were then deposited in deeper water conditions.

**Stop 4: MORETOWN FORMATION**

From Stop 3, proceed north on Bull Run Road back to rt. 12A. Take a left turn and follow rt. 12A west for 1 mile. Just past the golf course, the road will bend to the right. Park at the turn out on the left (south) side of the road. The outcrop is just west of the turn out and on the north side of the road.

**6.4 3.5**

This is an excellent example of interbedded chlorite rich meta-siltstone and phyllite of the Moretown Formation. A fault running through the outcrop is identified by a decrease in foliation spacing and an increase in mineral stretching lineation abundance. The fault is interpreted as syn-Acadian in age even though it has a papery fabric. Ln+1 crinkle lineation plunging shallowly to the north attest to the fault's age. Looking up high, a large wedge of rock sticks out over the ditch. This is the nose of an Fn isoclinal reclined fold similar to that observed at stop 1.

**Stop 5: ROXBURY MYLONITE ZONE**

From Stop 4 follow rt. 12 west and south through the village of Roxbury. Continue south of Roxbury for 1.3 miles to where the road takes a broad right curve. On the left side of the road is the entrance to an abandoned and reforested gravel pit. **Do not pull in, there are big rocks and a steep drop blocking the entrance.** Park at the entrance of the gravel pit. The outcrops of this stop are on the west side of the gravel pit.

**10.7 4.3**

The rock at this outcrop is a quartz-chlorite fine-grained protomylonite. The dominant fabric in this rock is Sn, the dominant Acadian fabric observed at all of the previous stops. If you look very carefully, very tiny interfolial lithic hooks can be found. These are interpreted as an older (Taconian) foliation. At the south end of the outcrop is an enigmatic, red-orange rock found in many places in the Roxbury mylonite zone.

**STOP 6: STOWE FORMATION SCHIST AND GREENSTONE**

From stop 5 continue along rt. 12A south for 0.75 miles and park at the State Fish Hatchery on the right hand side of the road. The outcrops are on the State Fish Hatchery land north of the buildings.

**11.45 0.75**

The rock at this stop is schist and greenstone of the Stowe Formation. Once again, the fabric of this rock is the same as that at Stop 5 and across the road (Roxbury mylonite zone). The contact of the Stowe Formation is marked by a thin to very thick greenstone which can be traced along strike to the southern edge of the Roxbury quadrangle. The fabric of these rocks (previously interpreted as Taconian) is Acadian and has been traced approximately 1 mile into the eastern edge of the Warren 7.5' quadrangle to the west.

*End of field trip for Day 2.*

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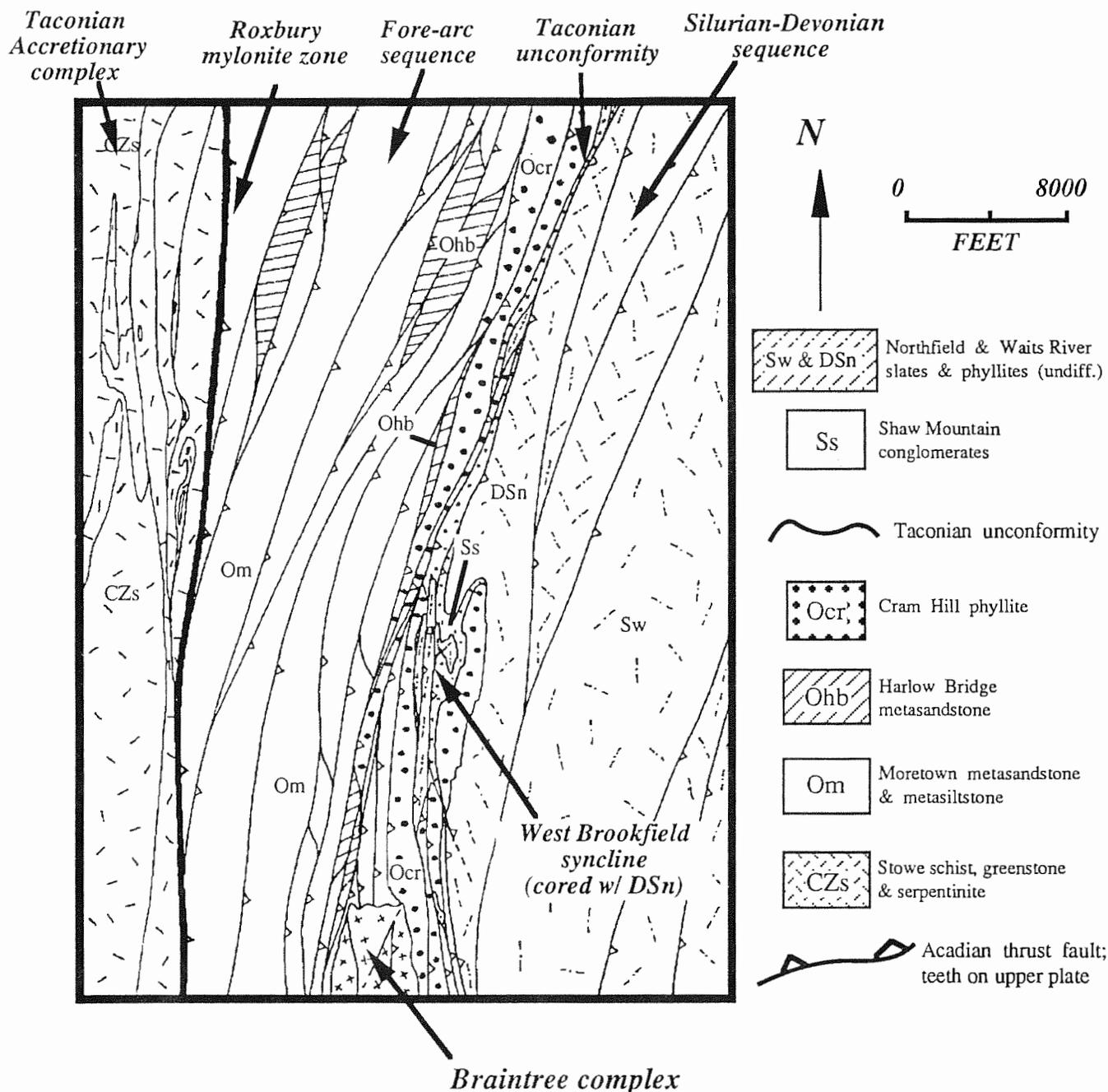


Figure 1. Simplified bedrock geology map of the Roxbury and eastern part of the Warren 7.5' quadrangles. The major structures and the 3 major lithotectonic belts and their contact relationships are described in the text. The Braintree complex is a suite of mafic and felsic intrusive igneous rocks. The mafic rocks are confined to the western margin and are clearly intruded by the felsic Braintree granite (eastern 90% of complex). Very weakly developed foliation in these rocks suggests that the intrusive events are syn- to post-Acadian in age. These rocks have been interpreted as part of the New Hampshire series (Doll, 1961). However, geochemical and radiometric age analysis have not yet been performed.

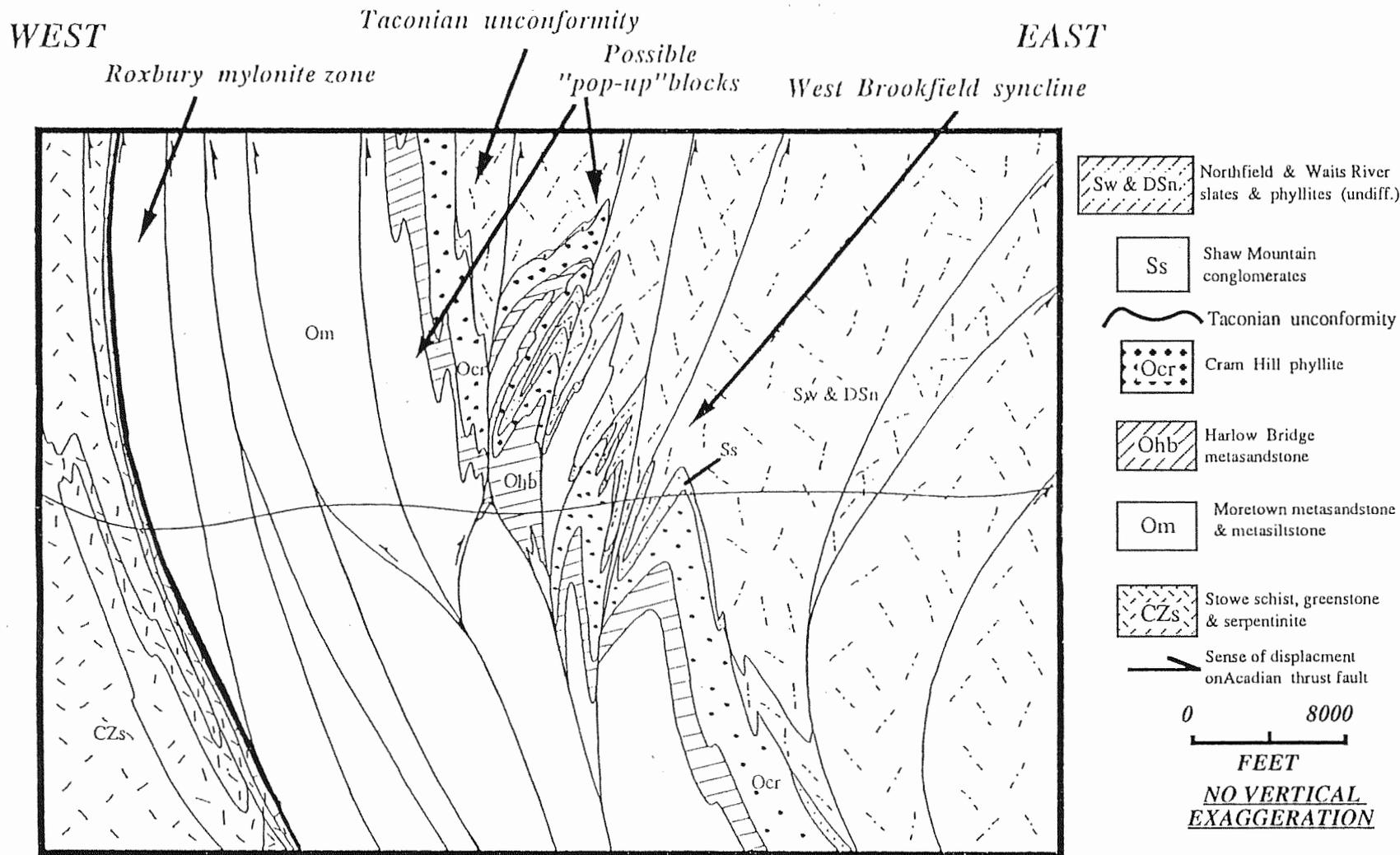
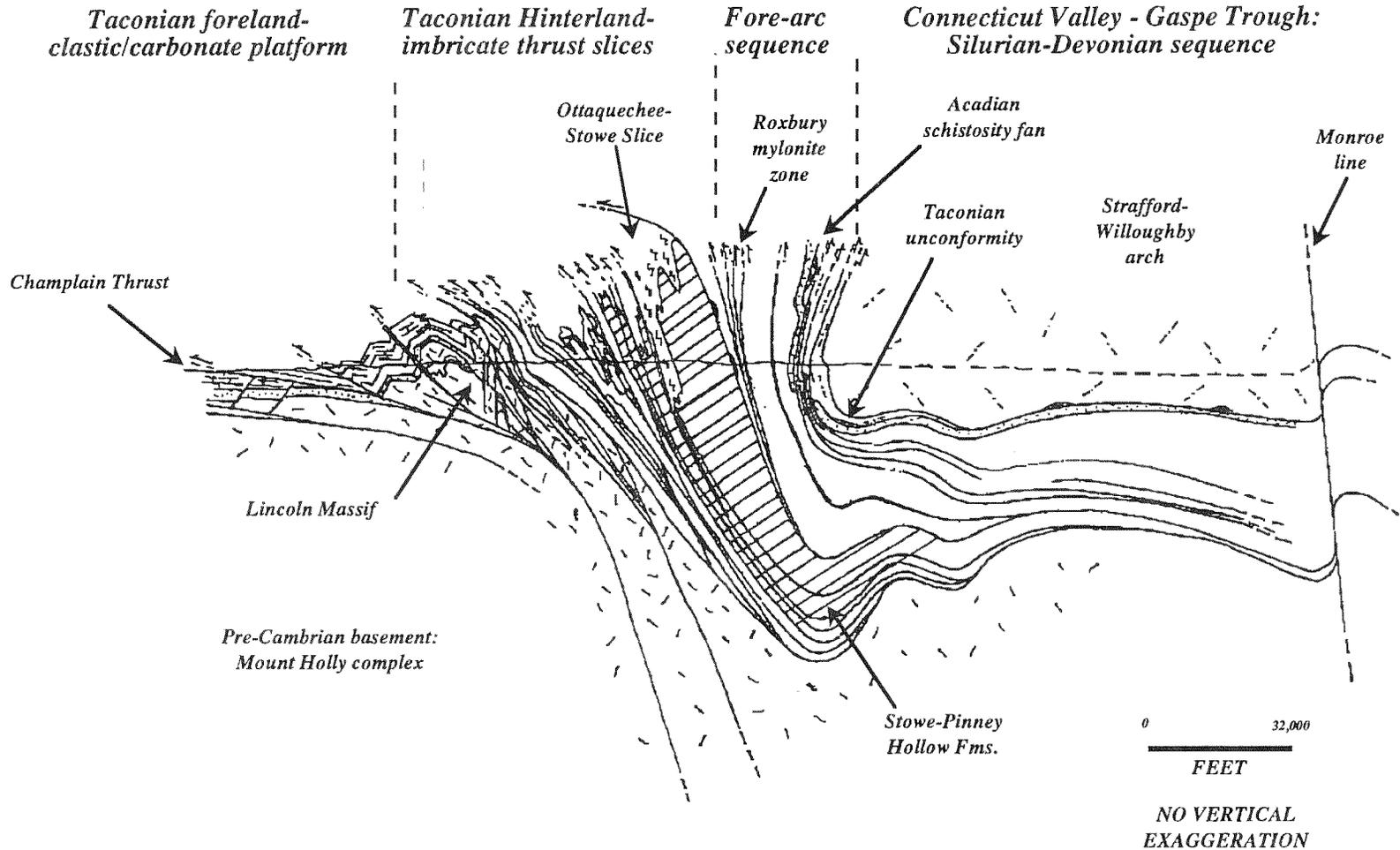


Figure 2. Simplified west to east cross-section across the Roxbury 7.5' quadrangle. It is hypothesized that all faults originally experienced east-over-west displacement. The sense of rotation of high amplitude folds (West Brookfield syncline) supports this idea. Evidence suggests that some of the faults have suffered some west-over-east adjustment during rotation and uplift of the schistosity fan. This led to development of possible "pop-up" blocks. The final phase of deformation in the Roxbury area was dominated by flattening.

West

East



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Figure 3. Simplified, schematic west to east cross section through the central Vermont Appalachians. Coeval ductile shortening of pre-fore-arc rocks and brittle shortening of overlying fore-arc and Silurian-Devonian rocks resulted in decoupling of the basement and the cover at the brittle-ductile transition. It is hypothesized that the Taconian accretionary complex acted as a rigid buttress thus limiting brittle Acadian east-over-west displacement. This, concurrent with ductile shortening of the underlying Taconian basement, resulted in uplift and rotation of syn-Acadian structures producing the Acadian schistosity fan.

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## TACONIAN ACCRETIONARY COMPLEX

### Itinerary

Because we must return to Boston today, the itinerary will be limited to convent outcrops along Rt 100 (the ski route) which is located along the Mad River and White River Valleys between the Green Mountains on the west and the Northfield-Braintree Range on the east. This route travels through the heart of the Taconian accretionary complex. Although we do not intend to exposure you to all the "gory details" we will show you outcrops which are representative of the tectonic fabric. Please refer to the short paper by Raymond Coish on the geochemistry of greenstones in the Appalachians which accompanies this trip description.

The rocks are largely Late Proterozoic to pre Ordovician rift - drift sediments that have been multiply deformed and metamorphosed during the Taconian Orogeny. Fragments of ocean crust (Iapetus) are represented by altered ultramafic rocks represented today by serpentinites and associated rocks which have been mined for dimension stone, asbestos, and talc throughout this belt from Quebec to western Connecticut. Compared to the relative simply fabrics which we saw in the Ordovician fore-arc sequence (Moretown and Cram Hill Formations) on Saturday, the fabrics here in the accretionary complex show evidence of repeated deformation associated with Taconian deformation. The papery schists associated with late fault zones and some of the younger open folds are probably Acadian in age.

Leave Burlington traveling eastward on Interstate 89 until Exit 10. Travel through Waterbury on Rt 2 eastward until it joins Rt 100. Turn south and travel along Rt 100 until you reach the village of Warren. The route passes through Waitsfield, which is approximately 5 miles north of Warren. Stop at the outcrop directly east of the junction of Lincoln Gap Road and Rt 100.

**Stop 1- COMPLEX SYNMETAMORPHIC FAULT FABRIC IN THE PINNEY HOLLOW FORMATION -** Warren, Vermont. The outcrop is located along the contact between the Pinney Hollow Formation, which underlies much of the Mad-White River Valleys, and the Hazens Notch Formation which underlies much of the Green Mountains north of its Middle Proterozoic core. Here the dominant mylonitic schistosity anastomoses and is well developed. Prominent mineral lineations and reclined folds of Fn age plunge down the dip. In one place an older Fn-1 fold is preserved in a lensoidal microlithon. Deformation of quartz- feldspar deposits suggests that synmetamorphic faulting occurred at temperatures in the order of 350<sup>0</sup> C to 400<sup>0</sup> C and pressures in the range of 3 to 4 Kbars (10 to 13 km). The assemblage quartz-muscovite-chlorite-biotite-albite with minor magnetite indicates greenschist facies metamorphism. In places along this belt of Pinney Hollow garnet is present with rims of chlorite which suggests an older an slightly higher grade of metamorphism which is consistent with the amphibole data from the Pinney Hollow mafic schists (greenstones). Because the pre Ordovician belt is dominated by several generations of mappable thrust zones, simple shear must have been a very important feature of the overall deformation history. Despite this scenario, remarkable little of this history is preserved in the fabric of the rocks. Asymmetrical porphyroclasts are relatively rare and inclusion patterns directly related to simple shear rotations are almost absent. Instead schistositities are well developed and are essentially parallel although they do anastomose. Porphyroclasts of quartz, albite, and garnet common have symmetrical tails indicating that flattening was a very important process in the late stages of deformation both of Taconian and Acadian age.

Continue south on Rt 100 approximately 8 mi. (13 km) to Granville Gulf. Park on the right side of the road just north of the long outcrop at the top of the hill leading into the Gulf.

**Stop 2 - GRANVILLE GULF - SYNMETAMORPHIC AND POST METAMORPHIC DUCTILE FAULT ZONE BETWEEN THE PINNEY HOLLOW SCHIST (east) AND THE OTTAUQUECHEE FORMATION (west) -** These outcrops are particularly important because they mark a complex, polydeformed fault zone along the western boundary of the Ottauquechee Formation and they contain mafic rocks. The faults zones are marked by well development foliation and down-dip lineations. As you can see these characteristic pervade the schists and are less obvious in the greenstones. The major criterion upon which these faults are based is the truncations of rock types

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along mutual boundaries. We interpret this geometry to represent footwall-hanging lithic cutoffs. The mappable rock types are silvery greenish gray schist, greenstone, rusty weathering, and slightly graphitic schist of the Pinney Hollow Formation, and black, graphitic, pyritiferous fine grained schist to phyllite of the Ottauquechee Formation. The postmetamorphic fault is marked by papery schists which represent a fault-zone foliation superposed on the older synmetamorphic foliation. The younger foliation envelops and cuts across albite porphyroclasts which grew across the synmetamorphic foliation. This older foliation is parallel to the synmetamorphic faults. The older pre/early metamorphic faults are offset by all the later faults. An altered serpentinite is present just to the east in the black fine grain schist of the Ottauquechee Formation.

The greenstones of the Pinney Hollow Formation contain barroisitic hornblende that has yielded  $40\text{Ar}/39\text{Ar}$  ages of 471 and 448 (Laird and Albee, 1981; Laird and others, 1984; Laird, 1987; see also V14, fig. 6). The hornblende is rimmed by actinolite which is also present throughout the matrix. The textures and compositions of the amphiboles indicate that these rocks have undergone two metamorphisms - a earlier medium-high pressure metamorphism and a later greenschist facies metamorphism.

Be careful as you walk along the road viewing the outcrops. Traffic can be plentiful. After you have seen enough of the road outcrops follow the trip leader into the woods to the east of the road where an altered serpentinite can be seen.

**Optional Stop - REFOLDED FOLDS OF Fn-1 AND Fn AGE WITH SYNMETAMORPHIC FAULTS - ALPINE VILLAGE.**

Travel south 4mi (6.4 km) to Granville. Turn west along Clark Brook Road. Stop at the first set of outcrops north of the road.

**Stop 3 PRE/EARLY METAMORPHIC AND SYNMETAMORPHIC DUCTILE FAULTS ALONG THE GRANVILLE AND PINNEY HOLLOW FORMATIONS.** Rock types include silvery greenish gray schist and greenstone of the Pinney Hollow Formation and rusty weathering, graphitic, albitic schist of the Granville Formation, a name given to these rocks by Philip Osberg (1952). Maps and other pertinent material will be distributed to the participants.

Travel 2 mi (3.2 km) south on Rt 100 to Lower Granville. Park on the west side of the road near the Church.

**Stop 4 SYMETAMORPHIC FAULTS AND GREENSTONE GEOCHEMISTRY AT ALLBEE BROOK -** This stop was originally mapped and described by Thomas Armstrong (Stanley and other, 1987). The geochemistry has since been Kafka, Elbert, and Coish (1993). These results are summarized in Raymond Coish's paper that accompanies this field trip description.

A synmetamorphic fault cuts an "S" shaped asymmetrical, reclined fold outlined by the contact between greenstone and schist. The fault is marked by a tan weathering, quartz-albite--mica mylonite. The mylonitic schistosity is parallel to the axial - surface schistosity coeval with the mappable folds. Their "S" shaped geometry suggests that displacement on the synmetamorphic fault was east-over-west. Maps and other published material will be handed out to participants.

**LAST STOP - ON TO BOSTON AND A FUN-FILLED WEEK!**

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## GEOCHEMISTRY OF GREENSTONES IN THE VERMONT APPALACHIANS<sup>1</sup>

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### Introduction

Metamorphosed mafic rocks (greenstones) from the Vermont Appalachians are found primarily in western to central Vermont. Samples for this study are from regions designated as zones 1, 2, 3 and 4 (Fig. 1). Farthest west, zone 1 contains late Proterozoic dikes that intrude the 1200 Ma Adirondack massif. The Proterozoic dikes are undeformed and exhibit low grade metamorphism, but preserve igneous features. Zone 2 includes dark, amphibole- or chlorite-rich greenstones mixed with metaconglomerates, metabasalts and phyllites. In places, the base of zone 2 (Pinnacle Formation and time-equivalent units) unconformably overlies Proterozoic gneissic basement. Felsic volcanic units in the Pinnacle Formation of Quebec are dated at 554 Ma (Kumarapeli and others, 1989). Zone 3 contains dark-green, chlorite-rich greenstones and light-green, epidote-rich greenstones intermixed with pelitic to psammitic schists. The age of zone 3 greenstones is not well-constrained, but probably lies between 554 Ma, the age of parts of the Pinnacle Formation, and 470 Ma, the age of metamorphism. Zone 4 includes light-green, epidote-rich greenstones intermixed with pelitic and psammitic schists. Zone 4 greenstones are older than 470 Ma, and perhaps contemporaneous with or younger than zone 3. Rocks in zones 2, 3 and 4 are deformed, and metamorphosed to greenschist and amphibolite facies. Greenstones in the northern part of zone 2 are identified as flows; elsewhere igneous form is obliterated by deformation and metamorphism. Contacts between zones 2, 3 and 4 are in many places west-directed thrust faults, indicating the rocks may have moved considerable distances eastward since their formation (Stanley and Ratcliffe, 1985).

### Geochemical Changes With Location

There are systematic changes in the chemistry of greenstones across western Vermont (Fig. 2). Zones 1 and 2 samples are alike in that they are mildly alkaline basalts with high contents of Ti, La/Yb, Zr, and Nb (Coish and Sinton, 1992). Their compositions are similar to oceanic island basalt (OIB) and continental rift basalt (Coish and others, 1985; Coish and Sinton, 1992). On the other hand, samples from zone 4 are tholeiitic basalts depleted in incompatible elements relative to zones 1 and 2, and similar to mid-ocean ridge basalts (Coish and others, 1986). Samples from zone 3 are mostly tholeiitic basalts, showing a wide range in Ti and incompatible element contents. The systematic nature of the chemical change with geographic location is illustrated by representative elements in Figure 2. Contents of TiO<sub>2</sub>, Zr, (La/Yb)<sub>n</sub>, and Zr/Y decrease whereas Sc and Al<sub>2</sub>O<sub>3</sub>/TiO<sub>2</sub> increase from west to east (Fig. 2). Whereas the average values of the elements show fairly smooth changes, there is a wide range of values within each zone leading to some overlap between zones. In particular, zone 3 shows considerable overlap with zones on either side. However, there is no chemical overlap between the western zones (1 and 2) and the eastern zone 4.

The variations between zones are also clearly seen when all data are plotted in a Ti-Zr-Y diagram (not shown here). This diagram has been used to assign ancient volcanic rocks to tectonic environments -- i.e., mid-ocean ridge, subduction zone, or within-plate (Pearce and Cann, 1973). Zones 1 and 2 overlap completely and plot in a within-plate environment in the Ti-Zr-Y diagram. Furthermore, because the greenstones in zone 2 are associated with coarse, continentally-derived metabasalts, it is argued that they formed within a continental plate. In contrast to zone 2, the chemistry of zone 4 greenstones falls in the undivided ocean ridge and island-arc field. On a Cr-Y diagram (not shown), zone 4 rocks clearly plot in the ocean-ridge field distinct from island-arc rocks (Coish and others, 1986). Greenstones from zone 3 span fields of within-plate and oceanic ridge environments. The Ti-Zr-Y variations and other chemical changes (Fig. 2) can be produced in basalt formed as a continental rift changes to an oceanic rift. In

<sup>1</sup> much of this report is taken from Coish et al. (1991)

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support of this contention, changes in basalt chemistry accompanying progressive stages in the development of the East African Rift - Red Sea area closely match the geochemical changes in metavolcanic rocks across the Appalachian orogen in Vermont (Coish and others, 1991).

Even though there is a general west to east systematic change in basalt composition, there is still a wide range of compositions within zone 3. In fact, within a single greenstone body at Albee Brook, there is a range in Ti and Zr values that is nearly as great as differences between zones 2 and 4. These variations are interpreted as igneous rather than metamorphic (Kafka, 1993). Furthermore, changes in basaltic compositions occurring over such a small region can be explained in a rifting stage transitional between continental and oceanic conditions. In this transitional stage, at least two mantle sources (enriched mantle plume, depleted asthenosphere) are probably available for melting to produce the different basaltic compositions.

Regional geochemical trends, ideas on mantle sources and implied tectonic environments of greenstones in western Vermont can be integrated into a model of splitting of an ancient North American continent and formation of the Iapetus ocean (Stanley and Ratcliffe, 1985). During the late Proterozoic (~ 600? Ma), the North American continent began to pull apart, perhaps because of a thermal mantle plume rising through the depleted asthenosphere. Magmas at this early stage may have been produced mainly from melting of the enriched mantle plume (OIB asthenosphere), giving rise to enriched basalt in zones 1 and 2. At a later proto-oceanic stage, the thinned continental lithosphere may have fragmented into continental blocks and narrow zones of incipient ocean crust. At this stage, in addition to the enriched mantle plume, the depleted asthenosphere may have been hot enough to melt partially, presumably by adiabatic decompression accompanying convective up welling. Thus, volcanism gave rise to both enriched and depleted basalt in zone 3. Still later (~525? Ma), the continental lithosphere was very thin or had completely split, and the depleted asthenosphere melted to produce the MORB-like basalts of zone 4. By this stage, the mantle plume had perhaps broken up and migrated far enough away from the main thermal anomaly that it no longer could contribute substantially to rift volcanism. The rift volcanics were then covered by early Ordovician, passive-margin sediments. Closure of the Iapetus ocean by eastward subduction in the mid-Ordovician deformed the continental margin, resulting in westward thrusting (Stanley and Ratcliffe, 1985) of zone 2, 3 and 4 rocks.

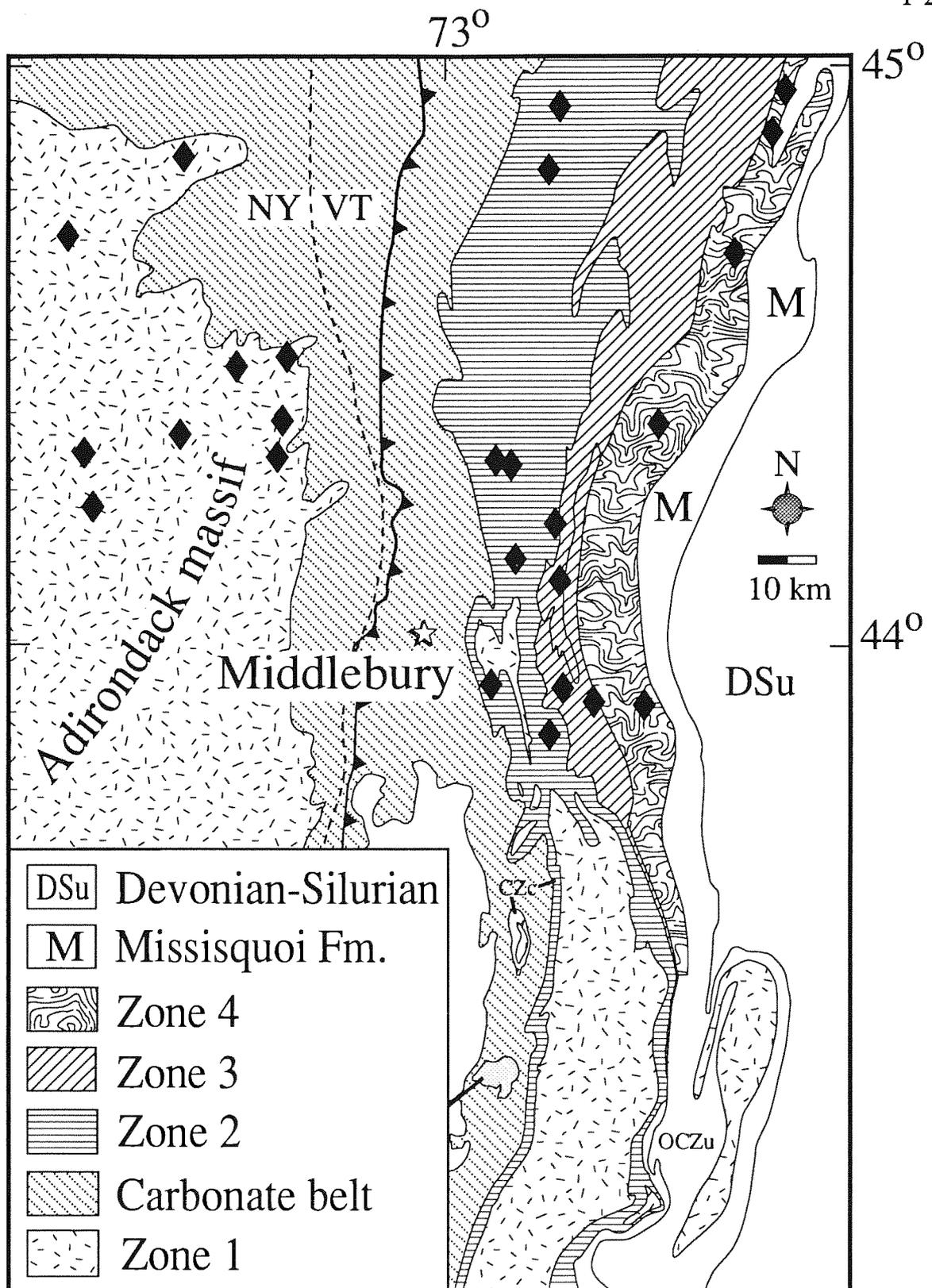


Figure 1. Simplified geologic map of northern Vermont and northeastern New York (after Stanley and Ratcliffe, 1985) showing sampled greenstones (diamonds). Zone 1 includes mafic dikes cutting Adirondack massif; zone 2 consists of Pinnacle and Underhill formations; zone 3 comprises Hazens Notch and Pinney Hollow formations; and zone 4 includes Ottauquechee and Stowe formations. Fault trending north across central part of map is Champlain thrust.

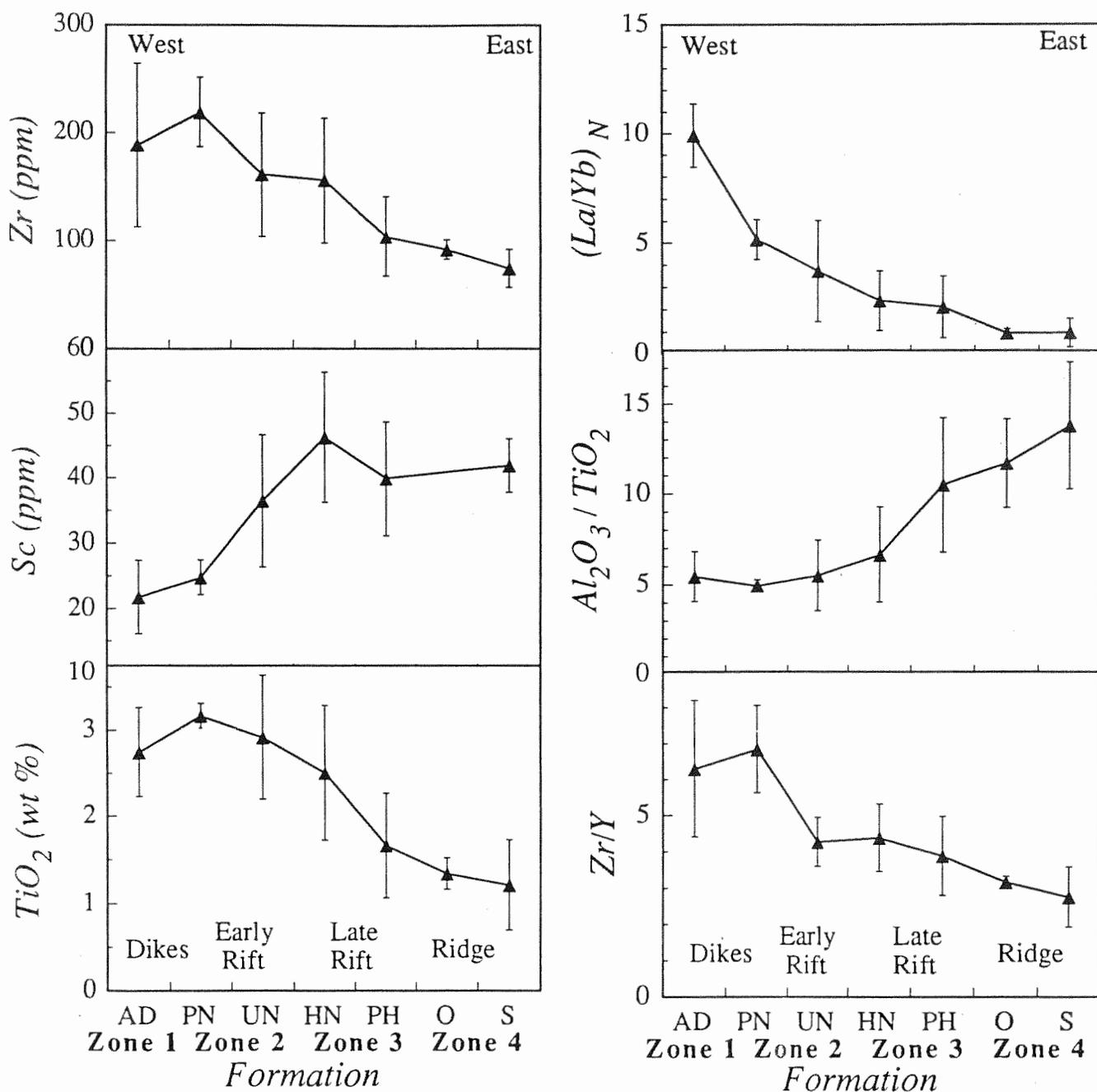


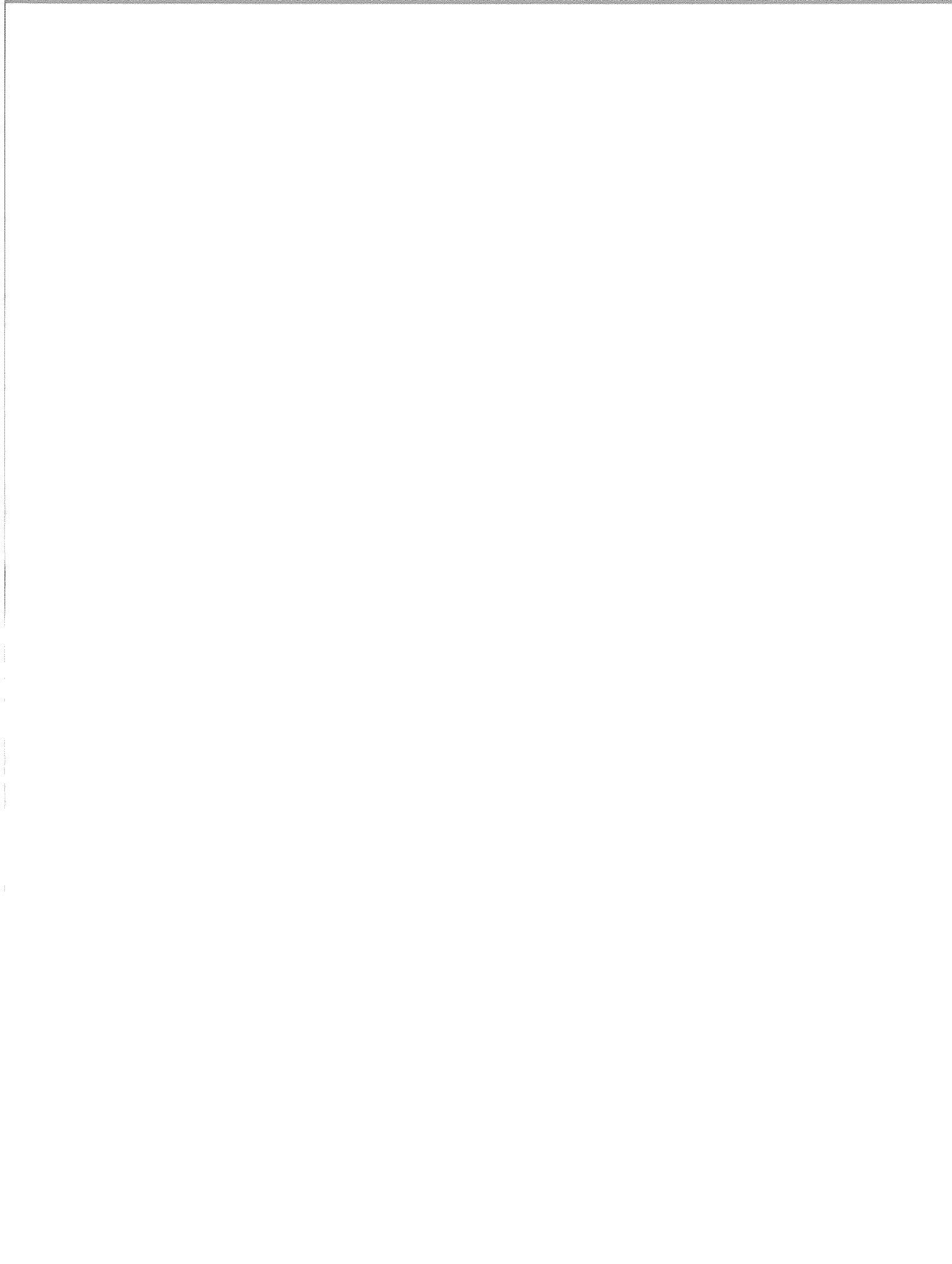
Figure 2. Variations in average concentrations (by formation and zone) of key elements and element ratios in mafic rocks across western Vermont Appalachians. Vertical line through each point represents one standard deviation. AD - Adirondack dikes, formations are PN - Pinnacle, UN - Underhill, HN - Hazens Notch, PH - Pinney Hollow, O - Ottauqueeche, S - Stowe.

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Chapter G

# Sea-level Change, Coastal Processes, and Shoreline Development in Northern New England

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## SEA-LEVEL CHANGE, COASTAL PROCESSES, AND SHORELINE DEVELOPMENT IN NORTHERN NEW ENGLAND

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### INTRODUCTION

The northern New England coast has long been recognized as a "drowned shoreline". Since the early work of Shaler (1874), Davis (1910) and Johnson (1925), many studies of relative sea-level change have focused on the coastal region from Boston to central Maine (Figure 1). On this trip we will review the history of study of sea-level change in New England and visit many of the most-studied coastal sites.

In recent decades there have been renewed studies of coastal processes in northern New England. Since coastal sediment is relatively scarce in this bedrock-framed region, we will focus on the processes responsible for sediment introduction to the coastal zone. In this context we will examine tidal inlets/river mouths and eroding glacial bluffs as possible sources of sediment for adjacent beaches and marshes, as well as consider the offshore region as a source/sink for sediment.

Even as the New England shoreline has responded to natural changes in sea level and variations in sediment supply, it has also been influenced by human activities for more than 300 years. As we visit the major beaches and marshes in the region we will review the case histories of their cultural development. Massachusetts and Maine have enacted strong legislation in the past decade to prevent unsound coastal development along their shores. We will visit some of the sites of past conflict between public and private rights, and between environmentalists and developers who have each used geological reasoning to make their respective cases.

### GEOLOGICAL SETTING AND HISTORY

Bedrock of Precambrian through Mesozoic age controls the overall shape of the New England coast. Boston Harbor and the surrounding environs are located in a Precambrian basin carved from relatively soft shales, conglomerates and volcanic rocks. Most of the natural environments of this coastal compartment have been filled, dredged, or otherwise altered for human purposes.

The Arcuate Embayments coastal compartment, extending north from Cape Ann, MA, is framed by capes or headlands of erosion-resistant plutonic rocks, with intervening bays sculpted from more easily eroded greenschist facies metamorphic rocks (Figure 1) (Kelley, 1987). The major coastal environments here involve the many curved beaches which protect extensive back-barrier marshes. Large sand beaches are especially well developed at the mouths of the Merrimack and Saco Rivers.

At Cape Elizabeth, ME, where metamorphic rocks of the Casco Bay terrane abruptly trend in a more northerly orientation, the Indented Shoreline Compartment begins (Figure 1). With few igneous rocks to interrupt them, the more resistant amphibolite grade gneisses of this region form elongate peninsulas, chains of islands, and shoals separated by narrow estuaries and bays. Tidal flats and salt marshes are the major environments along this coast, with large sand beaches confined to the mouth of the Kennebec River (Kelley, 1987).

The geomorphic setting of Maine's beaches differs from the classic coastal compartments of Fisher (1967): (Kelley and Belknap, 1989). Wells Embayment (Day 1) displays straight, semi-indented shorelines, and has local till and outwash as sources of sediment. Rivers in the Wells system have small drainages and do not transport large amounts of sand. The Kennebec River mouth shorelines (Day 3) protrude into the Gulf of Maine, and have a primary fluvial source of sediment, from the modern river and from the Kennebec Paleodelta (Belknap et al., 1986). Saco Bay (Day 2) is at the opposite end of a geomorphic spectrum, as it is deeply embayed and characterized by log-spiral crenulate shorelines, as well as straighter pocket beaches. Sources of sediments include fluvial input, primarily from the Saco River, direct erosion of pre-Holocene deposits at Western, Scarborough and Higgins beaches, and a probable reworking of shoreface Holocene littoral and Pleistocene glaciomarine deposits. Saco Bay also shows clear evidence of late Holocene progradation, probably due to the more protected embayment setting and

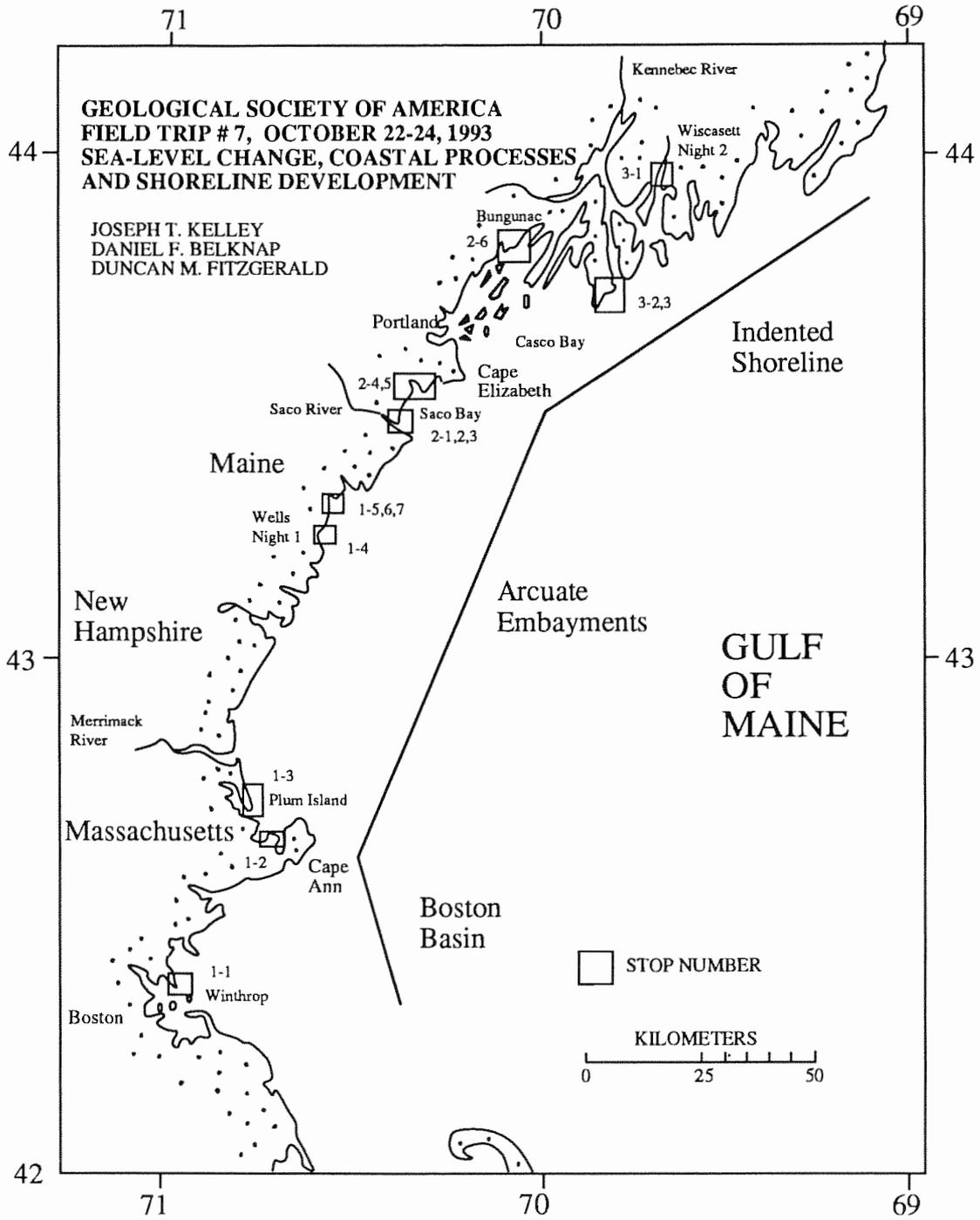


Figure 1. Location of the field trip stops in northern New England.

abundant sediment supply. Wells, in contrast, has transgressive barriers, probably due to the straighter, more exposed setting and lesser sediment supply. Kennebec River mouth beaches show a complex history of realignments, due to their variable orientations, complex shoreline morphology, and perhaps temporally varying sediment supply.

Although bedrock composition and structure determines the overall shape of the shoreline, glaciation has had the greatest influence on sediment availability in the coastal zone. The islands in Boston Harbor, and many of the unconsolidated bluffs to the north, are drumlins. Their erosion clearly supplies many of the local beaches with sand and gravel. Similarly, eroding bluffs of glaciomarine mud are the major source of fine-grained sediment to tidal flats and marshes in Maine (Belknap et al., 1986). Models have been developed to depict the behavior of beaches (Boyd et al., 1987; Hill and FitzGerald, 1992) and tidal marshes (Kelley et al., 1988) as available bluff sediment naturally declines. We will also visit sites where bluff-sediment supply has been terminated by human action, thereby accelerating shoreline change.

Another major influence of glaciation involves complex changes in relative sea level (Figure 2). In the late Pleistocene, the coastal lowlands of northern New England were flooded during and shortly after deglaciation because of isostatic loading of the crust. Glaciomarine sediment (the Presumpscot Formation, Bloom, 1963) covers lowland areas, while Gilbert-type deltas mark the sea-level highstand (Thompson and Borns, 1985).

As the glaciers retreated, land isostatically rebounded and the sea withdrew. Streams cut down into glacial sediment as sea level lowered, and large rivers, like the Merrimack and Kennebec, deposited extensive deltas seaward of their present mouths. We have inferred that these were most active during the late Pleistocene and early Holocene, when relative sea level was near lowstand. The lowstand of sea level apparently varies from -45 m at 12 ka off the Merrimack River (Oldale et al., 1983), to -55 to -60 m at 10.6 to 11 ka off the Kennebec River (Kelley et al., 1992; Barnhardt et al., in prep.)(Figure 2). Radiocarbon dates from cores in these deltas suggest that there is little modern sediment offshore, and that erosion of these "paleodeltas" provides sand to the extensive beaches at river mouths.

At the present time we do not know why a large river like the Saco has no paleodelta; nor do we quantitatively understand what role the large rivers play today in local sand budgets. The historical placement of engineering structures and location of dredged spoil disposal sites near river mouths assumes no sand is presently contributed to the ocean by these rivers.

Sea level rose at varying rates following the lowstand, but slowed down its rate of rise significantly by 5 ka. At that time many of the barrier beach systems developed into their present configuration, and extensive back-barrier marshes began. Radiocarbon dates from basal salt marsh peats suggest varying rates of relative sea-level rise between Cape Cod and the Bay of Fundy (Belknap et al., 1989), with the slowest rate of 1.2 mm/yr from Wells, ME (Kelley et al., in review). This value is about half the rate measured by the Portland, ME tide gauge from 1912-1986 (2.3 mm/yr, Lyles et al, 1988), and the present retreat of the beaches and erosion of the tidal marshes may be a result of the increased rate of historic sea-level rise (Jacobson, 1988; Kelley et al., in review).

## FIELD LOG

We depart from the lobby of the Marriott Copley promptly at 8 AM.

### DAY 1

**DIRECTIONS:** From Boston take the Callahan Tunnel to Route 1A north. Get off at the Winthrop exit after the exit to Logan Airport. This will bring you onto Bennington Street going east. Go approximately 1.3 miles and take a right onto Saratoga Street. At your second traffic light take a right onto Pleasant Street which you will follow all the way to Winthrop Beach.

**STOP 1-1, Winthrop Beach, MA** (Time: 60 minutes) The Town of Winthrop is a peninsula forming the northeast boundary of Boston Harbor. Like much of the surrounding region its topography is dominated by drumlins that have been connected by sand and gravel spits (Figure 3). Beaches consist of moderately-sorted sand with local concentrations of gravel which increase toward drumlin headlands. Cores taken along the beach and offshore region by the Metropolitan District Commission indicate that a typical stratigraphic section consists of a pervasive marine

MAINE COAST  
 LOCAL RELATIVE SEA LEVEL 14,500-0 B.P.  
 EUSTATIC RISE

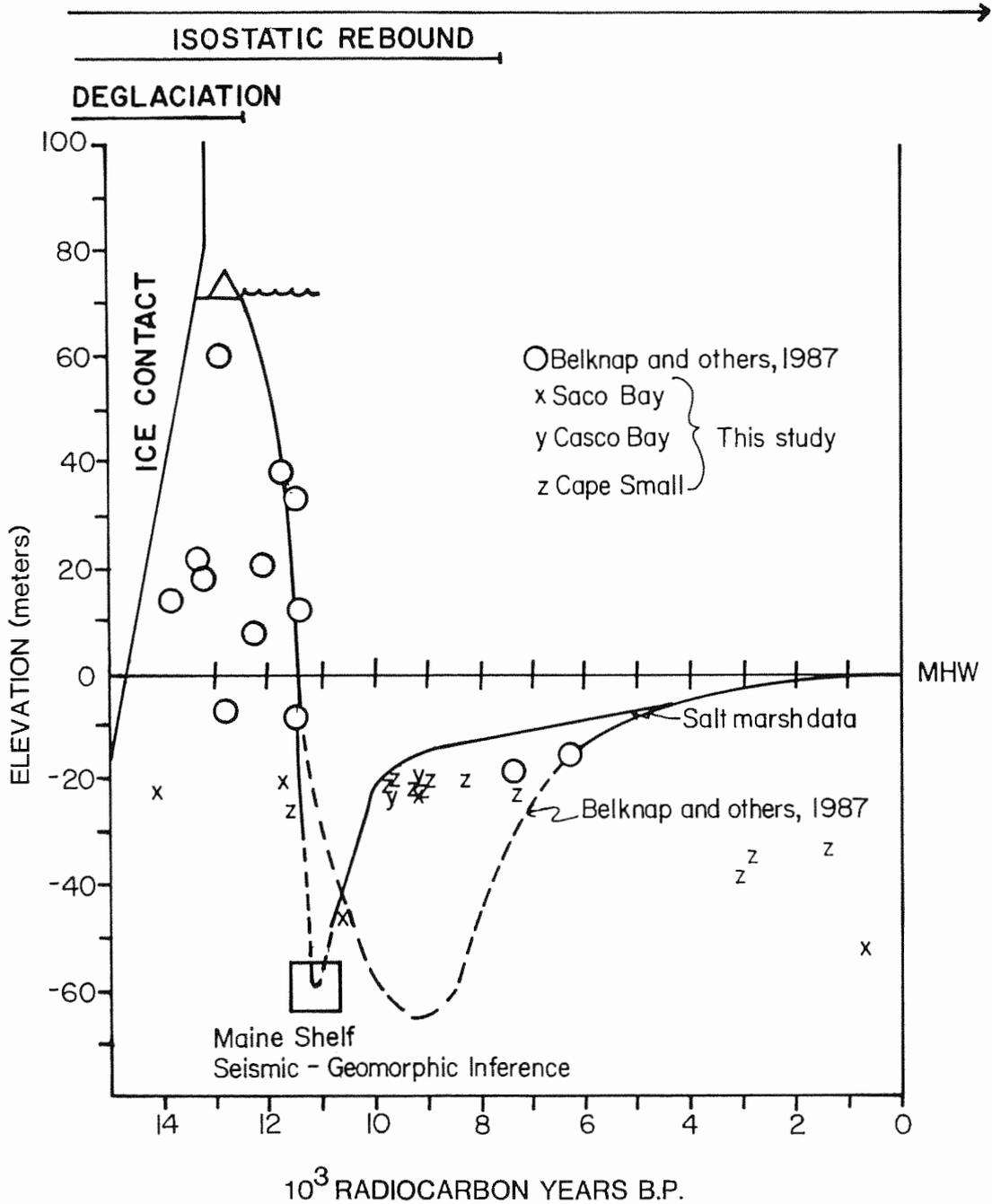


Figure 2. Sea-level curve for central Maine coast (from Kelley et al., 1992).

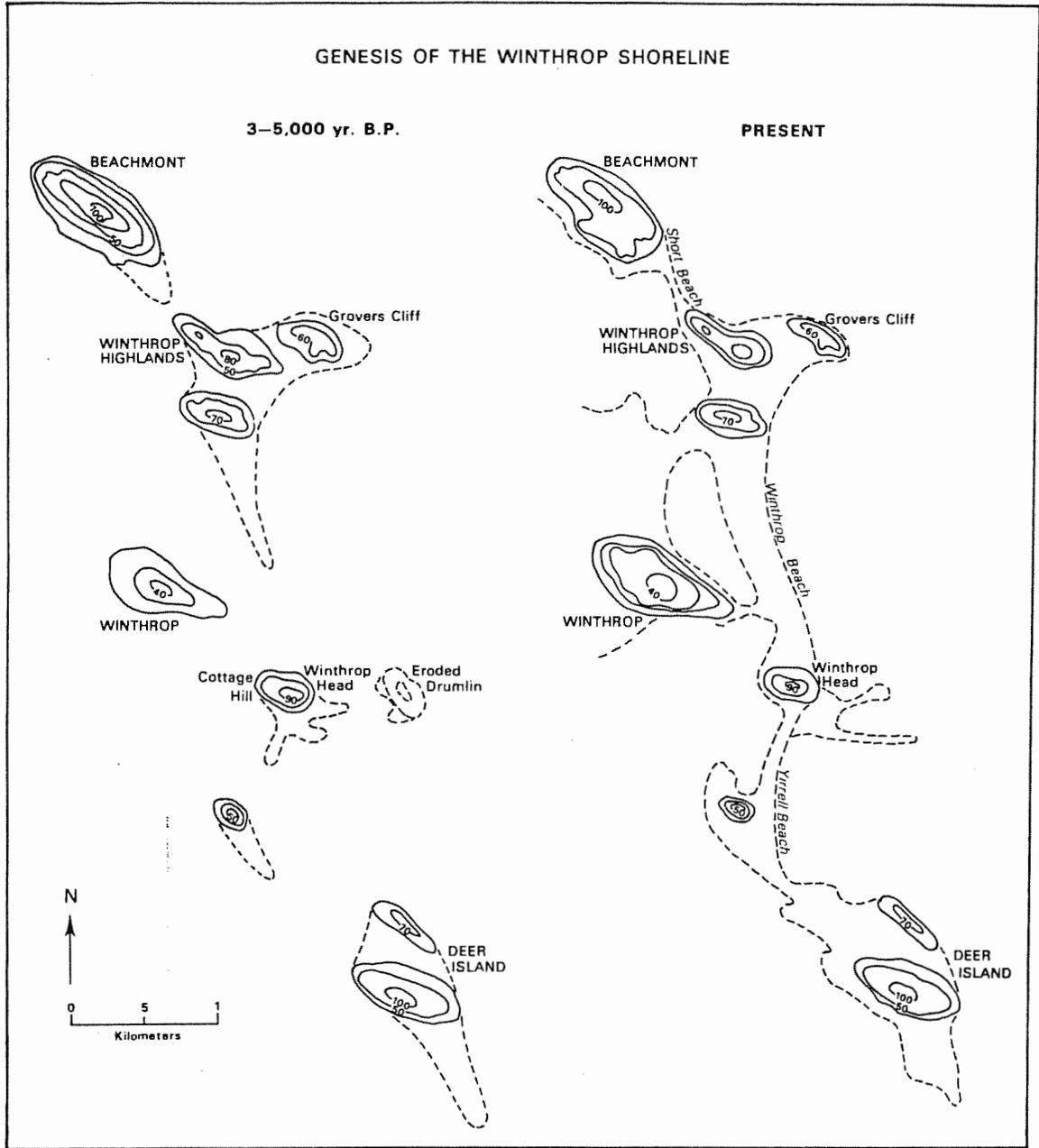


Figure 3. The Winthrop shoreline was formed from a number of drumlins that became connected through spit accretion (from FitzGerald, 1980).

clay (Boston Blue Clay) overlain by a transgressive sequence composed of a thin (< 1.0 m) basal peat and a thick layer (1.0 to 6.0 m) of sand and gravel. The sand and gravel unit thickens in an onshore direction and generally the surface of this deposit consists of gravel.

The entire shoreline in this area is backed by seawalls of various construction except for a small section of beach between Point Shirley and Deer Island that once was the site of a tidal inlet (Shirley Gut). Other coastal structures include five groins, two pedestrian ramps, piled rip-rap, and five closely spaced offshore breakwaters. The accumulation of sediment landward of the breakwaters is the result of wave refraction around these structures generating longshore currents toward the sheltered areas behind them. This section of the shore has also been the site of beach nourishment projects.

Off Winthrop Head two gravel tombolos extend approximately 600 m from the beach and are joined at their seaward ends by a gravel bar. The formation of these features is related to an eroded drumlin located just offshore of this region (FitzGerald, 1980; 1981).

The general configuration of the coast, which consists of headlands and embayed beaches, coupled with the large number of stabilizing structures, produces very low longshore sediment transport rates. Most of the movement of sediment along this coast occurs during major storms and is dominated by onshore/offshore transport. There have been long-term processes along the Winthrop shoreline of gravel movement onshore and the depletion of sand along the beaches, excepting the region behind the breakwaters. North and south of the breakwaters there exist only narrow beaches and at high tide waves break directly on the seawall. Historically, these beaches have experienced over 2 m of vertical erosion. Presumably this sand has been transported to the beach behind the breakwaters as well as to the offshore region. Analysis of vertical aerial photographs dating back to 1945 indicates that gravel has moved onshore between the individual breakwaters and that the seaward ends of the gravel tombolos have been flattened and moved onshore (Figure 3). Beach profile data has documented the landward movement of gravel bars during storms. Additionally, a gravel washover, almost 1 m in thickness was mapped following the February Blizzard of 1978 at the site of the former Shirley Gut (FitzGerald, 1980).

Along some sections of the Winthrop coast major storms are responsible for the transport of sediment over seawalls. The seawall north of the breakwaters has a curved, concave-seaward face which is designed to expend wave energy by propelling the water from breaking waves in a skyward direction. During major northeast storms, such as the February Blizzard of 1978 and the Halloween Eve Storm of 1991, the breaking waves entrain sand and gravel at the base of the seawall and the water and sediment mixture is propelled as much as 15 m above the top of the wall. The strong onshore winds accompanying these major storms blow the much of water, sand, and gravel onto the adjacent street and, in some cases, through nearby cars and houses.

South of Winthrop Head is a section of beach backed by a low seawall that stands about 1 m above the sand. During major storms the beach face is eroded and most of the sediment is moved offshore. However, during the same period the upper beach face sands are transported onshore by wave swash and deposited next to the seawall. This process continues until a sand ramp is formed to the top of the seawall which allows the transport of water and sediment over the wall. During the February Blizzard of 1978 25,000 m<sup>3</sup> of sand were transported over the seawall by this process, inundating cottages and driveways with up to 1 m of sand (FitzGerald, 1980; 1981).

**DIRECTIONS:** Take Route 128 North and get off at the Wingersheek exit. At the end of the ramp, take a left and follow signs to Wingersheek Beach. After a short visit to Coffins Beach, we will proceed to Crane Beach or Plum Island. From Coffins Beach we will go back to the highway and take Route 133 North and follow signs to Crane Beach or Plum Island.

**STOPS 1-2 and 1-3 COFFINS BEACH AND CRANE BEACH/PLUM ISLAND** (Time: 2 hrs)  
New Hampshire and Cape Ann, Massachusetts. With a combined length of 34 km, this is one of the two longest barrier chains in New England. The chain consists of five barriers, the largest of which is 13 km-long Plum Island at the mouth of the Merrimack River estuary (Figure 4). The barriers are backed primarily by marsh and tidal creeks, although bays are prevalent behind inlets. The existence of these backbarrier bays is attributed to drowned river valleys which control the location of all but one of the tidal inlets as well as the overall geometry of most of the backbarriers. The unstructured inlets have well-developed ebb and flood-tidal deltas (e.g. Essex Inlet) whereas the jettied inlets have well-developed flood-tidal deltas and diminutive, subtidal ebb-tidal deltas (e.g. Merrimack and Hampton Inlets) (Hayes, 1975; Boothroyd and Hubbard, 1975).

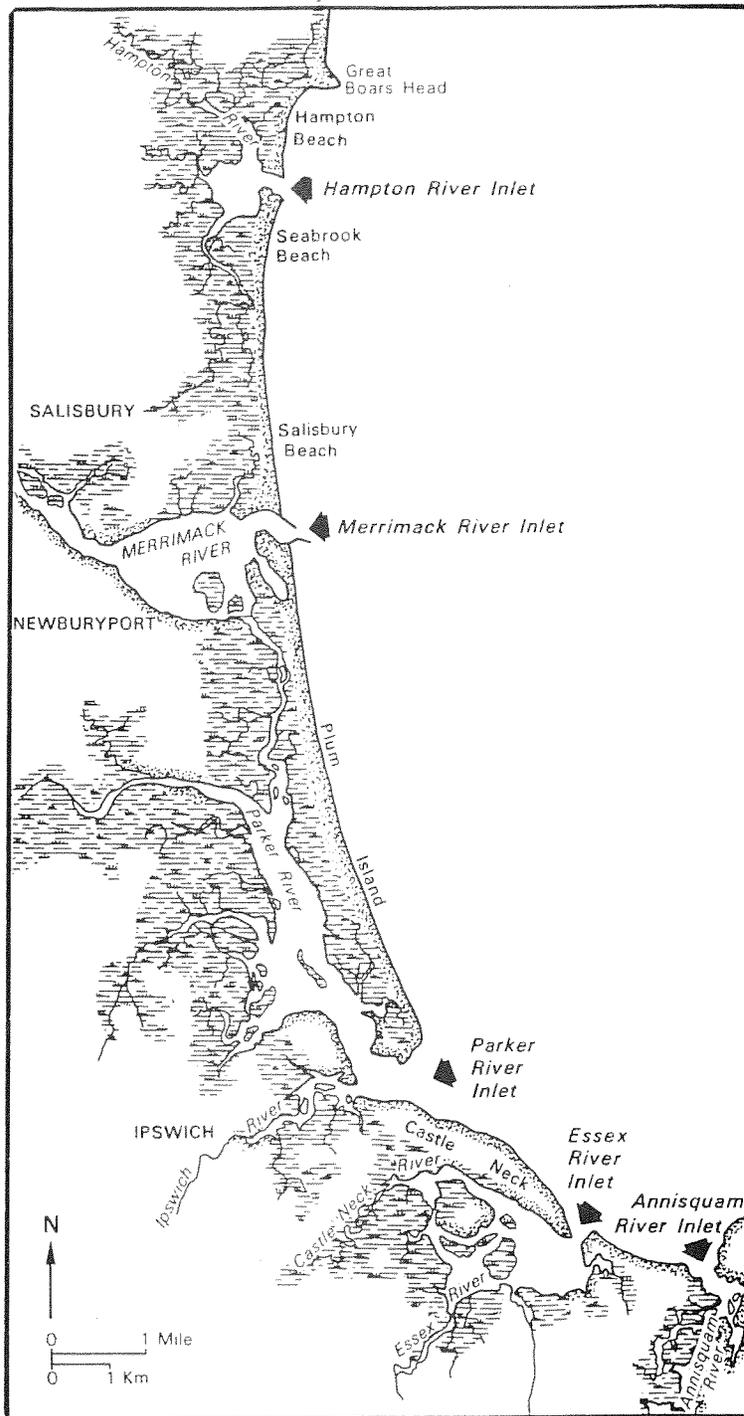


Figure 4. The Merrimack Barrier Chain is one of the longest in New England, and is associated with the Merrimack River paleodelta.

The sediment abundance south of the Merrimack River estuary contrasts with the lack of sediment to the north. The barriers south of the Merrimack are relatively wide with extensive dune development and fairly flat offshore slopes. The barriers to the north have had a history of erosion which has necessitated the construction of numerous protective engineering structures and the implementation of beach-nourishment projects (U.S. Army Corps of Engineers, 1982). Moreover, the lack of sand in the northern offshore region explains the irregular offshore contours and the bedrock exposures.

The asymmetric distribution of sand along this barrier chain is similar to that of the Kennebec River barrier system; it is a product of the location of the lowstand Merrimack River paleodelta and the dominant northeast wave climate of the region. The lowstand delta is positioned 8 to 10 km offshore (at -50 m) and parallels the present-day coast from Hampton Inlet, New Hampshire to Essex Inlet, Massachusetts (Figure 5). A shallow seismic-study of this area has revealed that the delta is approximately 20 km long, 7 km wide, and up to 20 m thick (Oldale et al., 1983).

The onshore reworking of the surface portion of the delta sands during the Holocene transgression formed the present barrier system (Rhodes, 1973; Edwards, 1988; Som, 1990; FitzGerald, 1993). It is likely that the sand moved onshore in the form of transgressive sand sheets and narrow low barriers that periodically stabilized at bedrock and glacial pinning points (c.f. Boyd et al., 1987) (Figure 6) During the transgression, the sand units were exposed to the dominant northeast wave climate driven by northeast storms. The resulting net southerly longshore-transport system coupled with the position of the Cape Ann promontory has produced a preferential accumulation of sediment along the southern portion of the barrier chain. This is shown clearly by the seaward excursion of the -20 m contour, which increases in distance from 1,700 m offshore of northern Plum Island to more than 4,000 m offshore of Coffins Beach. Grain size data along the beaches also demonstrate a southerly transport trend. Mean grain size decreases from 0.2 phi at northern Plum Island to 2.2 phi at southern Coffins Beach (Figure 5), while grain sorting increases from 0.25 phi to 0.41 phi along the same coastal transect (Goodbred and Montello, 1989).

Insights concerning the stratigraphy and evolution of the barrier chain are known from coring and geophysical studies of Plum Island and Castle Neck (McIntire and Morgan, 1964; Rhodes, 1973; Boothroyd et al., 1978), the Castle Neck/Essex Bay marsh system (Som, 1990), and recently of Coffins Beach (McKinlay et al., 1993) (Figures 5,6). Although somewhat concealed by an extensive cover of sand, bedrock and glacial deposits have had an important influence in defining the position of the barriers and the thalweg of the tidal-inlet channels. For example, the southern end of Plum Island, northern Castle Neck, and the flood-tidal delta at Essex Inlet are all pinned to till-covered bedrock. In addition, bedrock outcrops anchor the throat sections of Hampton, Essex and Annisquam Inlets and portions of many backbarrier channels.

Along Coffins Beach a large volume of fine sand has been blown landward forming an extensive dune system. Along the southern portion of the barrier what appears to be the highest dunes, with heights greater than 20 m, are simply sand-covered bedrock. In the northern part of the barrier sand dunes reach 10 m in height. Preliminary analyses of an ongoing ground penetrating radar and coring study of Coffins Beach indicate that the barrier lithosome is greater than 6 m thick and is underlain by backbarrier sediments which overlie marine clays and bedrock. Aeolian sands appear to comprise a large portion of the barrier unit as evidenced by up to 3 m of fine well-sorted sand overlying soil horizons of former sand deposits.

Stratigraphic sections of Castle Neck and Plum Island reveal that the barrier lithosome varies in thickness from 10 to more than 15 m and is underlain by glacial deposits including till and a ubiquitous glaciomarine clay (McIntire and Morgan, 1964; Rhodes, 1973; Boothroyd et al., 1978). In this region the clay is called the Boston Blue Clay (Judson, 1949; Kaye, 1961) which is the equivalent of the fine-grained member of the Presumpscot Formation in Maine (Bloom, 1960). In addition to glacial deposits, upland and backbarrier deposits underlie portions of the Plum Island barrier sands (Figure 7). A single date from a basal freshwater peat beneath the middle of Plum Island yielded an age of 6,280 BP, indicating a more seaward position of the barrier at that time (McIntire and Morgan, 1964).

Along the backside of Plum Island, cores show an interlayering of sand and silty sand which McIntire and Morgan (1964) have interpreted as a reflection of the gradual landward migration of a transgressive barrier. While it is probable that the initial stage in the evolution of this shoreline involved transgressive barriers, the thickness and width of Plum Island, as well as that of Crane Beach and other barriers in the region, would suggest that for the past approximately 6,000 years Plum Island has remained stable by building vertically. The interlayering of clean and silty sands that occurs in the cores can be explained by a widening of the barrier and periodic changes in sedimentation conditions in the backbarrier. The latter are caused by storms, migration of tidal channels and sand bodies (Boothroyd and Hubbard, 1975), rear-dune-ridge construction (Fink et al., 1984), and other processes. In

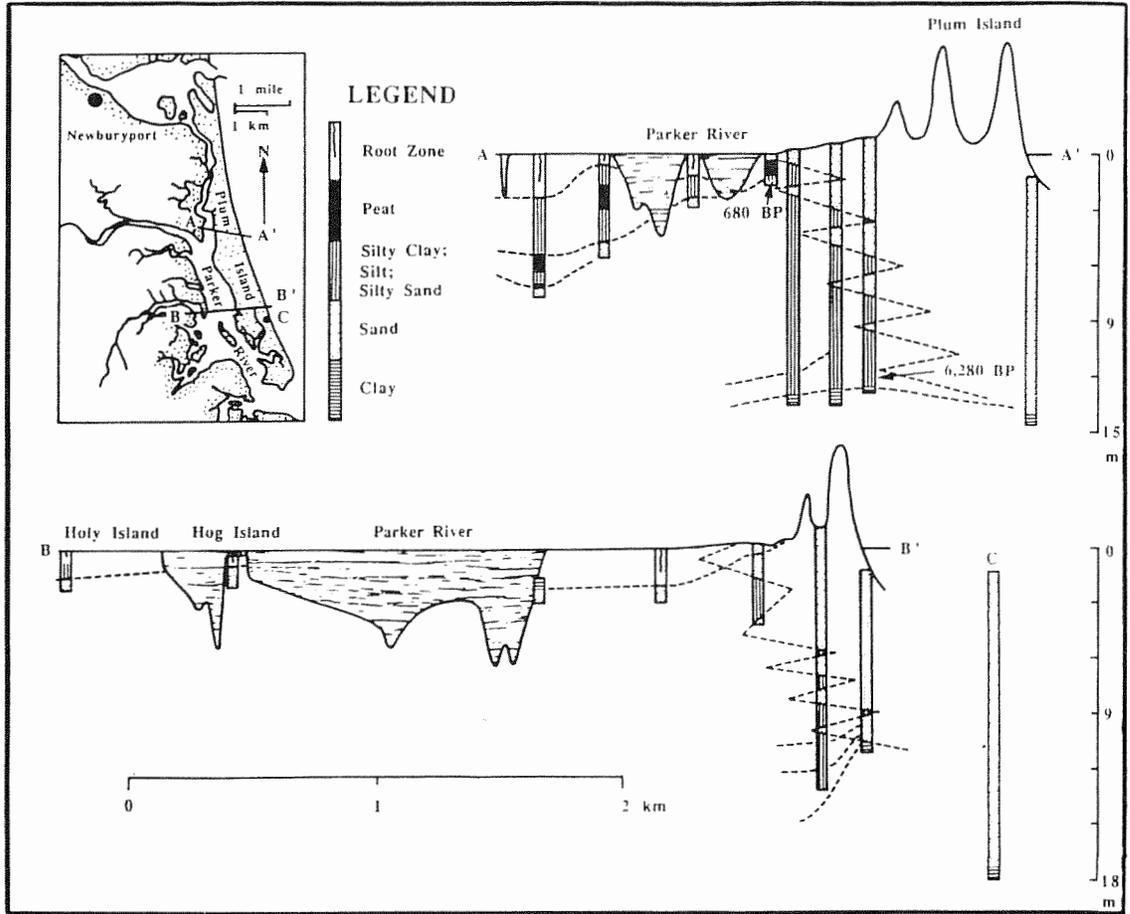


Figure 5. Stratigraphic cross sections of Plum Island (redrawn from McIntire and Morgan, 1964).

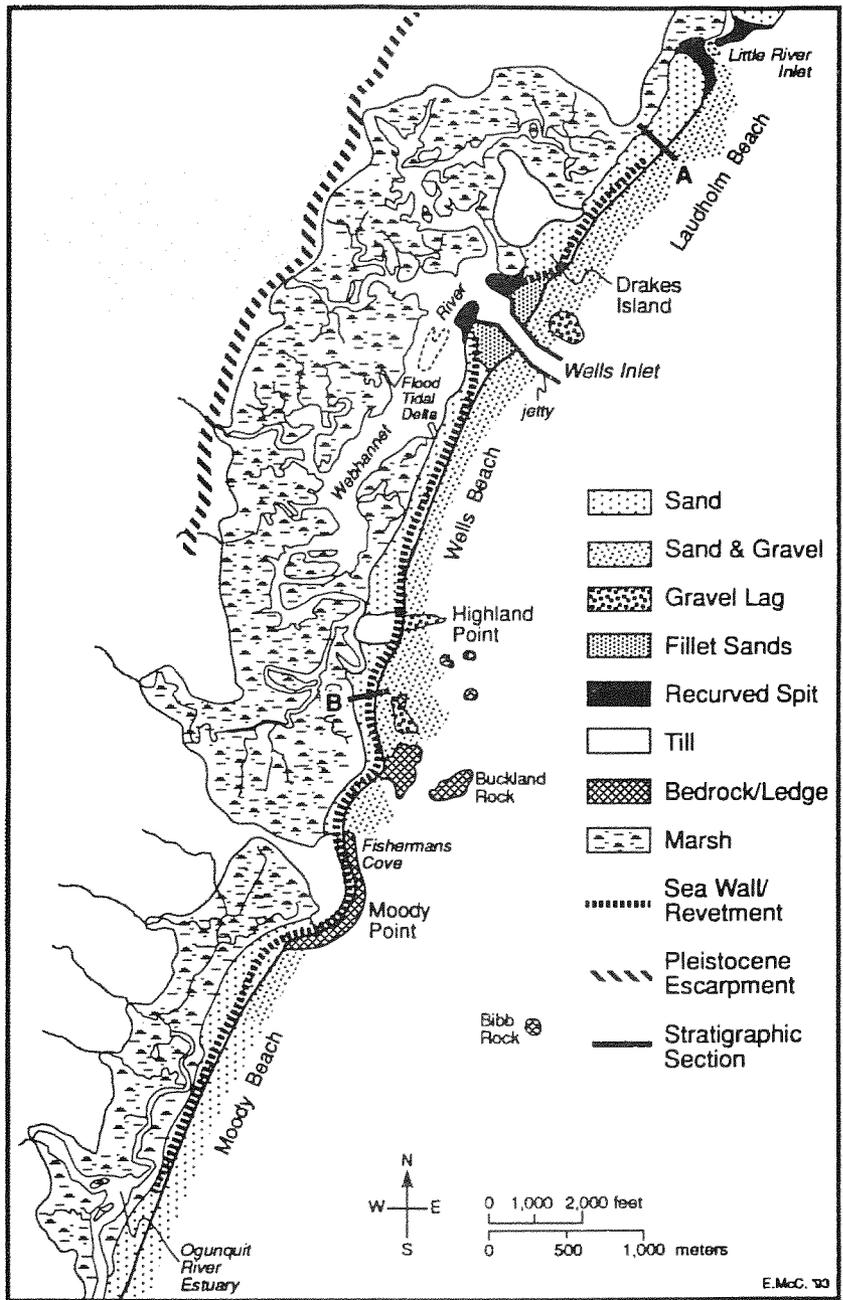


Figure 6. Geomorphology of the Wells/Ogunquit region (from Montello et al., 1992)

# SW-WC MAINE BARRIER COMPARTMENTS

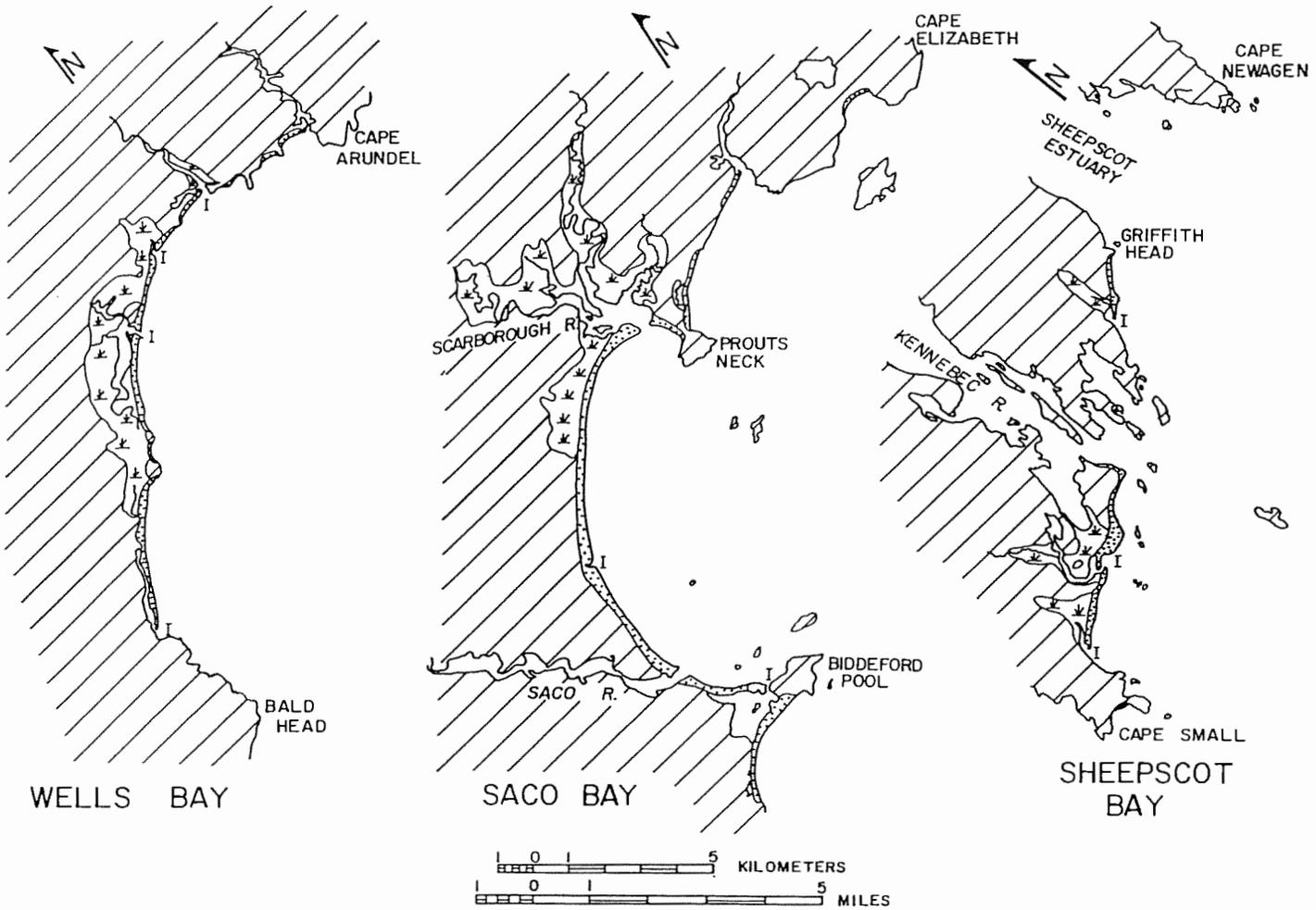


Figure 7. Maine's major beach and marsh systems.

addition, it should be noted that the width of Plum Island (> 500 m), its height (3-6 m), and the extent of its vegetated dunes preclude washovers and landward aeolian transport.

Directions: Leave Interstate 95 at the Wells exit and turn left onto Rt. 9 for 1.7 miles until the red light at the junction of Rt. 1. Turn left on Rt. 1 for 1.6 miles until the sign for Laudholm Farm/Wells Estuarine Research Reserve. Follow signs to parking lot and assemble at farmhouse for lunch.

**STOP 1-4: LAUDHOLM FARM (LUNCH, 60 MINUTES)** The hill upon which the farmhouse rests is composed of till, with glaciomarine mud draped along the sides. Most boulders were cleared for agricultural purposes, although some are visible in the woods and in terraces on the hillsides. The prominent terrace below the porch of the farmhouse (approximately 11 m msl) appears to be a shoreline cut into unconsolidated material (Bloom, 1960). The many boulders protruding from the break-in-slope were probably eroded out by waves. This conjecture is supported by the observation that the terrace is best developed in orientations facing the prevailing storm directions, southwest and northeast. Several other smaller terraces occur at lower elevations, with the most prominent at around 5 m msl. This terrace may correlate with the escarpment on the western edge of the salt marsh (Figure 6). No radiocarbon samples have ever been found in raised shorelines in this area, but the marine limit reached 60 m above present sea level ca. 13.5 ka, 10 km inland from Wells (Belknap et al., 1987). The terraces were presumably formed after the highstand, during the early Holocene regression of the sea.

Directions: We will return to Rt. 1 and turn left for 1.5 miles to the Lower Landing Rd. Cross down the possible shoreline escarpment and drive to the end of the Lower Landing Rd.

**STOP 1-5. WELLS HARBOR AND SALT MARSH (60 MINUTES).** Wells hosts one of the 3 largest beach and marsh systems in Maine (Figure 7). While the other beaches in Saco Bay and at Popham Beach are associated with large river mouths, the Webhannet River in Wells is quite small (Figure 8).

The Wells salt marsh is among the most studied in North America for evidence of past sea-level changes (Figure 9). Hussey (1959; 1970) and McIntire and Morgan (1964), Timson (1978) obtained some of the first radiocarbon dates from peat samples in the region from estuarine deposits exposed on the beach and in marsh cores, respectively. Belknap et al. (1989), Shipp (1989), Kelley et al. (in review), and Gehrels et al. (1993) successively refined earlier work to produce a sea-level change curve (Figure 9). The data depict a relatively rapid rate of sea-level rise, 1.2 mm/yr at 4 ka, until around 2 ka when the rate decreased to 0.5 mm/yr. Contemporary sea level, as evaluated at the Portland tide gauge, is rising at 2.3 mm/yr.

Cross sections of the tidal marsh based on more than 75 "Dutch" hand auger cores and vibracores (Belknap et al., 1989; Kelley et al., 1993), reveal a complex evolution from a simple tidal flat to the present "mature" high marsh community (Figure 10). The pre-marsh surface contained numerous gulleys or ravines cut into glaciomarine sediment by streams during the sea-level lowstand. These were separated by interfluves of relatively flat glaciomarine mud and isolated hills of till (Kelley et al., 1993). As sea-level reached the present estuary, the gulleys were filled with tidal channel sandy mud, while *Spartina alterniflora* colonized the interfluves. Extensive freshwater marshes rimmed the salt marsh system. With the slowing of sea-level rise and sediment accumulation in the estuary, *Spartina patens* succeeded both *S. alterniflora*, in the central areas of the embayment, and freshwater marshes on the margins (Figure 10). The present rapid rate of sea-level rise, greater than at any time since the early Holocene (Figures 2, 9), may be responsible for the observed widening of salt marsh tidal creeks, growth of salt pannes, and retreat of the barrier beach (Jacobson, 1988; Kelley et al., in review).

An additional influence on the estuary has involved creation of a federal anchorage in the Webhannet River by the U.S. Army Corps of Engineers beginning in the early 1960's. The anchorage was located directly on the flood tidal delta of the river, and initial attempts to dredge a harbor failed due to rapid shoaling (Byrne and Ziegler, 1977). Sandy spoils from early dredging were dumped on the adjacent salt marsh, and surrounded by a drainage ditch. Over time the spoils site has apparently subsided, and tidal strams have migrated toward it (Jacobson, 1988). Dredging of the drainage ditch and harbor led to local rejuvenation of the marsh system and downcutting by tidal creeks to the new, lower base level. *S. alterniflora* is now colonizing the margins of the deep (2 m) channels, which require great care when crossing at mid to high tide. The harbor has not been dredged since the middle 1970's for a variety of reasons, one of which is the impact of dredging on the marsh system.

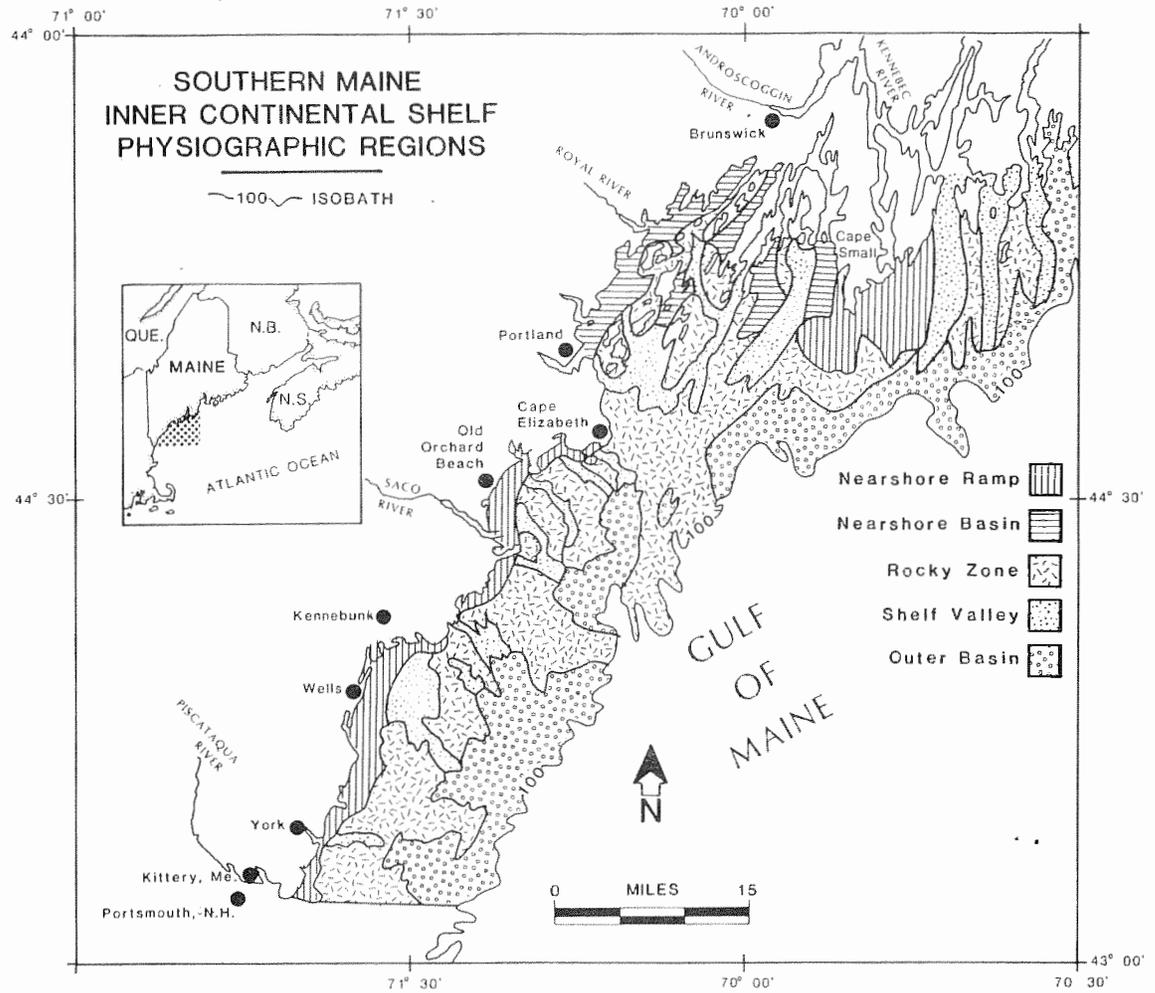


Figure 8. Physiographic environments of the Maine inner shelf depicting the sandy regions seaward of major beaches (from Kelley et al., 1989).

## SEA-LEVEL CHANGE, WELLS, MAINE

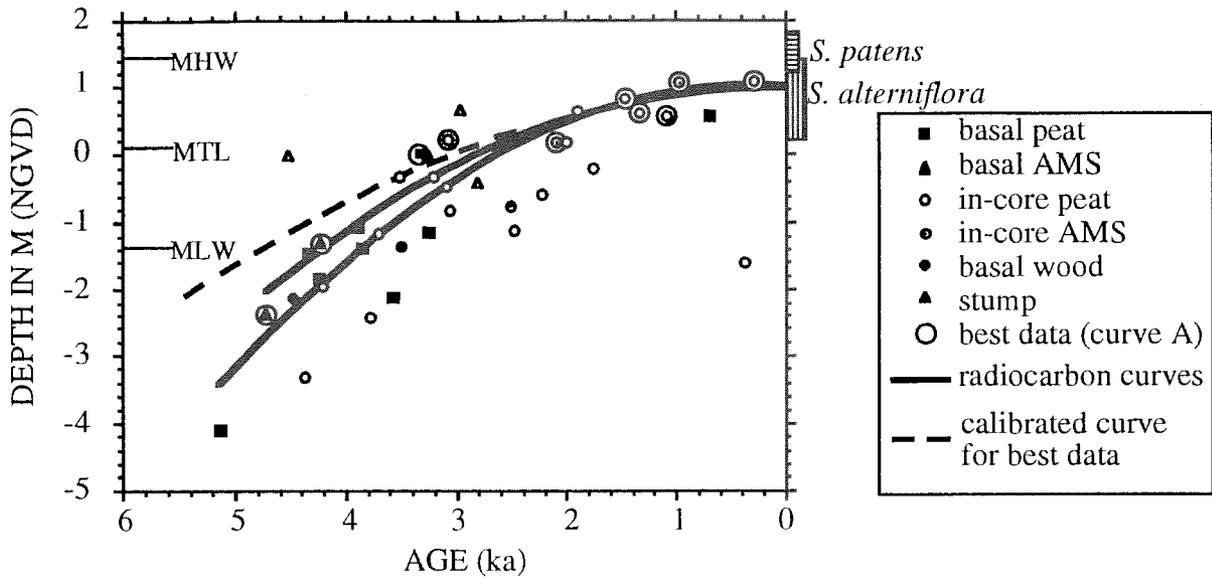


Figure 9. Sea-level change curve for the Wells area, Maine (from Kelley et al., in review).

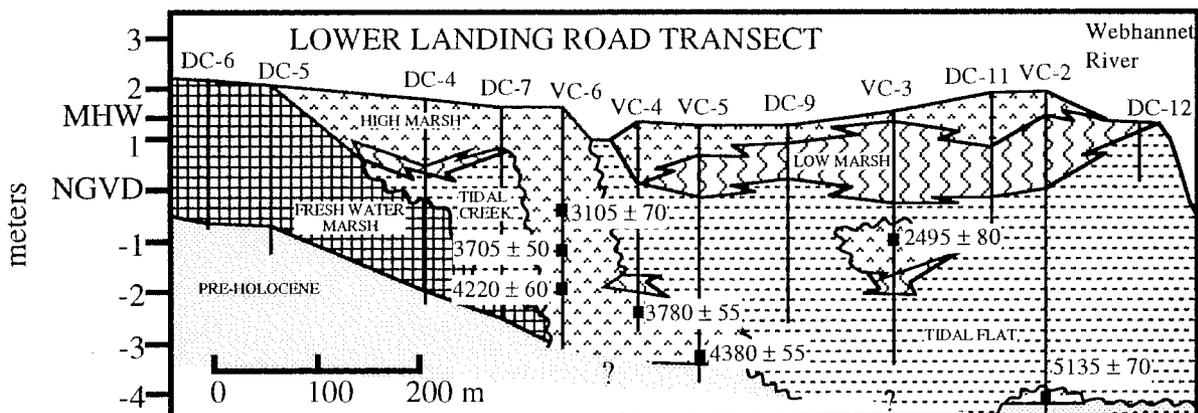


Figure 10. Cross section of the Wells Marsh at Lower Landing Road (from Kelley et al., in review).

Directions: We return on the Lower Landing Rd. to Rt. 1, where we turn left (south) for 1.3 miles. At the Mile Rd. traffic light we turn left onto Mile Rd. until it ends at the municipal parking lot. Here we turn left again onto Beach St. and follow it north to its end at the tidal inlet parking lot.

**STOP 1-6, WELLS BEACH AND TIDAL INLET (60 MINUTES).** The beaches of the Wells region are narrow (80-200 m) compared to river-related barriers in New England (Figure xx), which are more than 300 m wide. The barrier lithosome is < 5 m thick, except in areas of extensive dune development and in topographic lows between headlands (Figure 11). The barrier sands rest on glaciomarine sediment, which lies over till or bedrock. Sections through the back barrier region exhibit interfingering of washover sand and marsh peat. A radiocarbon date from *S. Patens* peat at 4.1 ka demonstrates that the barrier was offshore of its present position before the slowdown in the rate of sea-level rise. The quasi-stabilization of the barrier near the present shoreline was coincident with decelerating rates of sea-level rise over the last 2.5 ka.

The existence of the Wells Embayment barrier beaches appears related to incision by the Webhannet River and other small streams during the regression of the sea (Nelson, 1979). These streams derive sand from the 250 km<sup>2</sup> Sanford Outwash Plain (Thompson and Borns, 1985), and it is estimated that they delivered more than  $200 \times 10^6 \text{ m}^3$  of sediment to the sea (Montello et al., 1992). The present volume of sand in the beach and back-barrier region is only  $33 \times 10^6 \text{ m}^3$ , suggesting much remains offshore.

Examination of nearshore lag deposits, however, suggests that till deposits at Moody Beach, Drakes Island and Great Hill, for example, have supplied longshore transport, resulting in the barrier systems. This model of evolution, noted by Hussey (1970), is similar to the model proposed for the south shore of Nova Scotia (Boyd et al., 1987), or the drumlin-tied barriers of Plymouth, Massachusetts (Hill and FitzGerald, 1992). It is reasonable that this mechanism created a saltating barrier system, in which barriers became more and less stable depending on local sources, as sea-level rose and the shoreline retrograded during the late Holocene.

Shipp (1989) delineated seismic facies offshore, producing an isopach map. Wells Embayment is floored by sands and muddy gravels, with abundant rock outcrops. Shipp (1989) concluded that the mixed gravel and sand beaches in this embayment reflect eroded glacial deposits as well as river-derived outwash sand. Total sediment thickness is greater than 10 m nearshore, but seismic profiles do not clearly delineate Pleistocene sandy outwash and proglacial marine from the Holocene reworked material. We have no offshore cores comparable to those available in Saco Bay and off the Kennebec River (Days 2 and 3). Thus, until more data become available on offshore sand volumes, a sediment budget for this region is incomplete.

Sand is now stripped from southern Wells Beach (and the Drakes Island barrier to the north of the inlet) and concentrated in the equilibrium trap of the Wells jetties (Figure 12). Sea walls also disrupt the natural equilibrium profile and alter the grainsize distribution. Complete removal of sand results in a gravel berm, such as at Fishermans Cove, just north of Moody Point. Longshore transport is evident in the morphology of the spits and the accumulation around artificial jetties, but there is no strong net transport direction, due to seasonal reversals. Wells, Moody and Ogunquit beaches have the appearance of simple spits prograding north and south from Moody Point, but it is unclear that there was enough sand available in the thin till exposed there to create these systems. More likely, they represent equilibrium landforms containing sediments recycled during transgression and trapped within the headlands of the Wells embayment. Other sources of sediment may include long-term reworking of sandy Pleistocene units on the shoreface.

Growth of the beach adjacent to the tidal inlets ceased in the 1970's, and dune grasses have colonized much of the sand. With the loss of their sand, residents of the barrier beaches have long opposed the federal anchorage. In addition to loss of sand, considerable property damage has occurred along the beach. Because Maine's coastal sand dune laws do not permit rebuilding of damaged houses or construction of new seawalls, the sand adjacent to the jetties is viewed as a final hope to save the beach. Recently, with the State's denial of a permit to fully dredge the anchorage, local beach residents have petitioned to have the jetties torn down, or the sand trucked from the jetty impoundment back up the beaches. It has been noted that the jetties, in addition to trapping beach sand, are oriented directly into the prevailing swells, leading to rapid shoaling. Inlets here and at most locations in Maine are localized by till or bedrock, forming at the margins of embayments. Historic inlets are known in the centers of barriers, but they are less common.

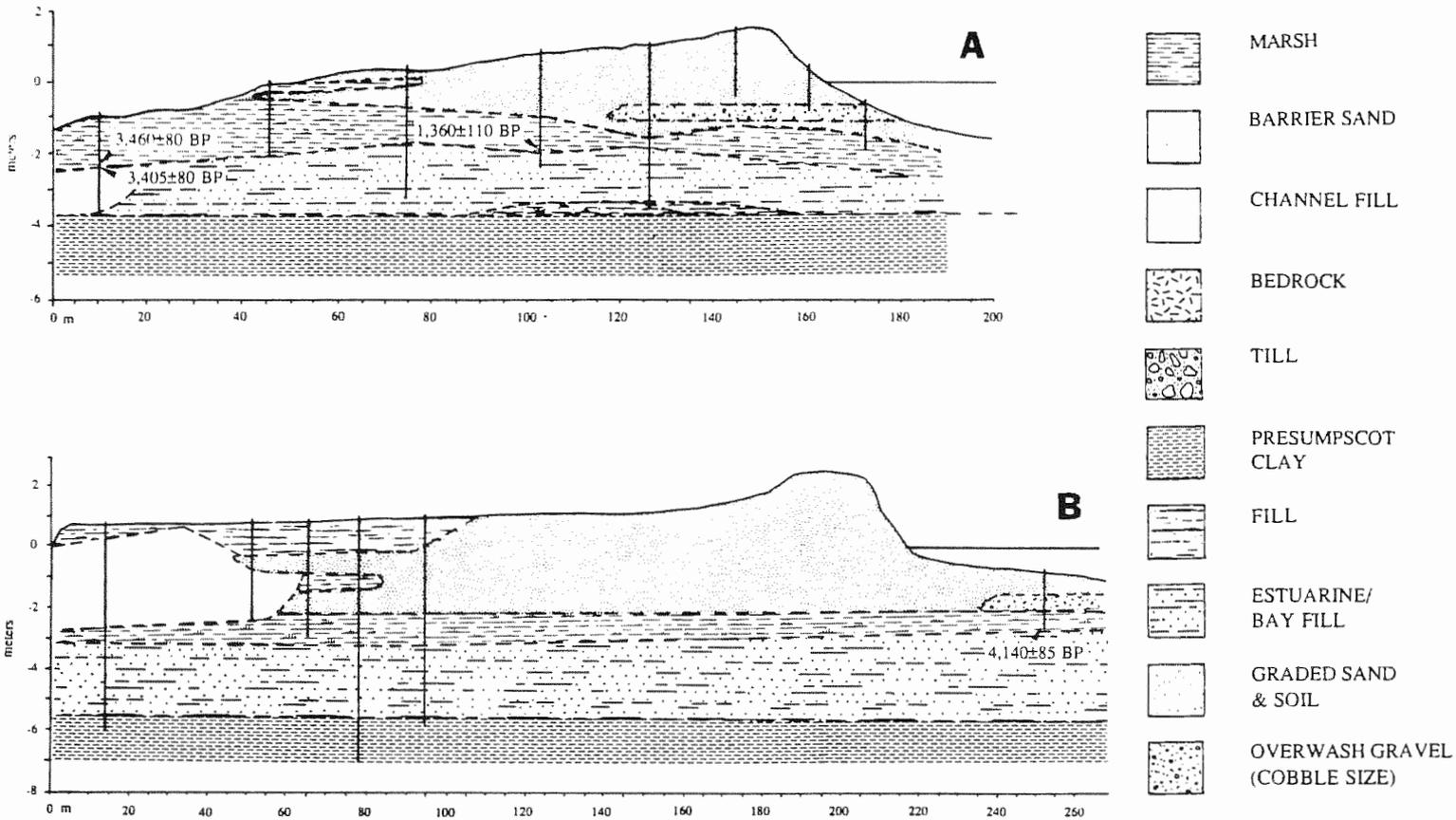


Figure 11. Stratigraphic cross sections of Laudholm Beach (A) and Wells Beach (B) (from Montello et al., 1992).

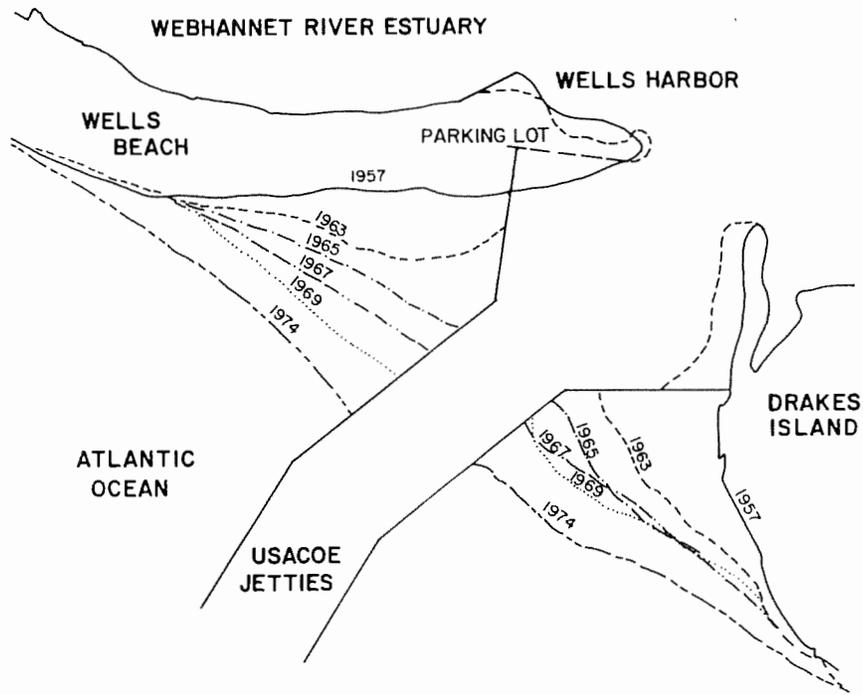


Figure 12. Shoreline change at Wells Inlet (from Kelley et al., 1989).

Directions: Return along Beach Rd. all the way (3.6 mi.) to the southern end of the barrier and enter the parking lot near the sewage treatment plant.

**STOP 1-7, OGUNQUIT BEACH (30 MINUTES).** At the Town line separating Ogunquit and Wells, there is an abrupt change in land use practices. The Wells barrier beach (here called Moody Beach) is fully developed for residences. Local property owners, with 17th century deeds granting them ownership to the low tide line, recently won a State Supreme Court case prohibiting the public from recreational use of the beach (public "fowling, fishing and navigation" uses are all that are allowed). Ironically, erosion of publically owned Ogunquit Beach sheds some sand to the exclusive Moody Beach.

To the south, Ogunquit Beach experiences a net southerly longshore transport, which would be predicted by the dominant northeast wave approach, and has been supported by process measurements and sedimentological data (Lincoln et al., 1985). Once sand reaches the inlet, wave-generated and flood-tidal currents transport the sediment across the spit platform and into the inlet channel. Sand in the outer portion of the channel is moved in a net seaward direction by the dominant ebb-tidal currents, and is eventually deposited on the ebb-tidal delta. Here wave action moves the sand back onshore to the spit platform and adjacent beach, thus completing the counterclockwise sediment gyre.

Sand which is deposited in the inlet channel along the landward portion of the spit platform is transported landward by flood-tidal currents. This sediment accumulates on flood-tidal deltas, in channels, and in other intertidal environments. This landward movement of sand is corroborated by geomorphic evidence. In 1974 the tidal delta was mined of its sand for construction of the artificial dune. One year later the flood-tidal delta had completely reformed to its original size and shape.

Although it appears relatively pristine, compared to Moody Beach, Ogunquit Beach is both fragile and artificial. It terminates at the south in a seawall-protected parking lot and motel complex. Despite damage by many storms, these properties and the bridge connecting them to the mainland, are very armored and less impacted by storms than the sand dunes.

The sand dunes at Ogunquit Beach are a thin veneer of sand over a gravel and sand nucleus. Constructed in 1974 by the Soil Conservation Service with dredged material from the tidal delta and local borrow pits, the fortified dunes are intended to protect the regional sewage treatment plant bordering Moody Beach. Despite encirclement by snow fences and occasional sand replenishment, the dunes do not offer long-term stability and protection of the plant. Nevertheless, an offshore pipeline was recently extended seaward from the plant, signaling a long-term public commitment to the facility.

## DAY TWO

Directions: We will return to Interstate 95 from the hotel by way of Rt. 1. We leave the Interstate at Exit 5, and proceed onto Rt. 1 south (right turn) at the end of the exit ramp. We turn left off Rt. 1 onto Beach St. (Rt. 9) in Saco. We follow Rt. 9 to Camp Ellis (4 miles) and then follow signs through the damaged coastal community to the parking lot at the federal anchorage.

**STOP 2-1, CAMP ELLIS AND THE SACO RIVER MOUTH (60 MINUTES).** Saco Bay contains the largest beach system in Maine (Figure 13). Sand delivered by the Saco River since the lowstand of sea level (Kelley et al., 1986; 1992), or submarine erosion of glacial outcrops (U.S. Army Corps of Engineers, 1992) is assumed to be responsible for formation of the beach. Believing that sand resulting from submarine erosion of glacial deposits moves south along the beach, the Army Corps of Engineers constructed a massive jetty at the mouth of the river beginning in the 19th century. The jetty has been enlarged repeatedly in this century as the anchorage has continued to fill with sand.

Recent current meter observations of flood-current domination over ebb currents, bedform analysis by side-scan sonar, and down-estuary fining of grain size support the alternative hypothesis, that sand is currently delivered to the beach by the river. The accumulation of sand at the mouth of the jetties (Figure 14), and the orientation of paleospits (Figure 13) (stop 2-3) also suggests that northward-moving river sand formed the beach.

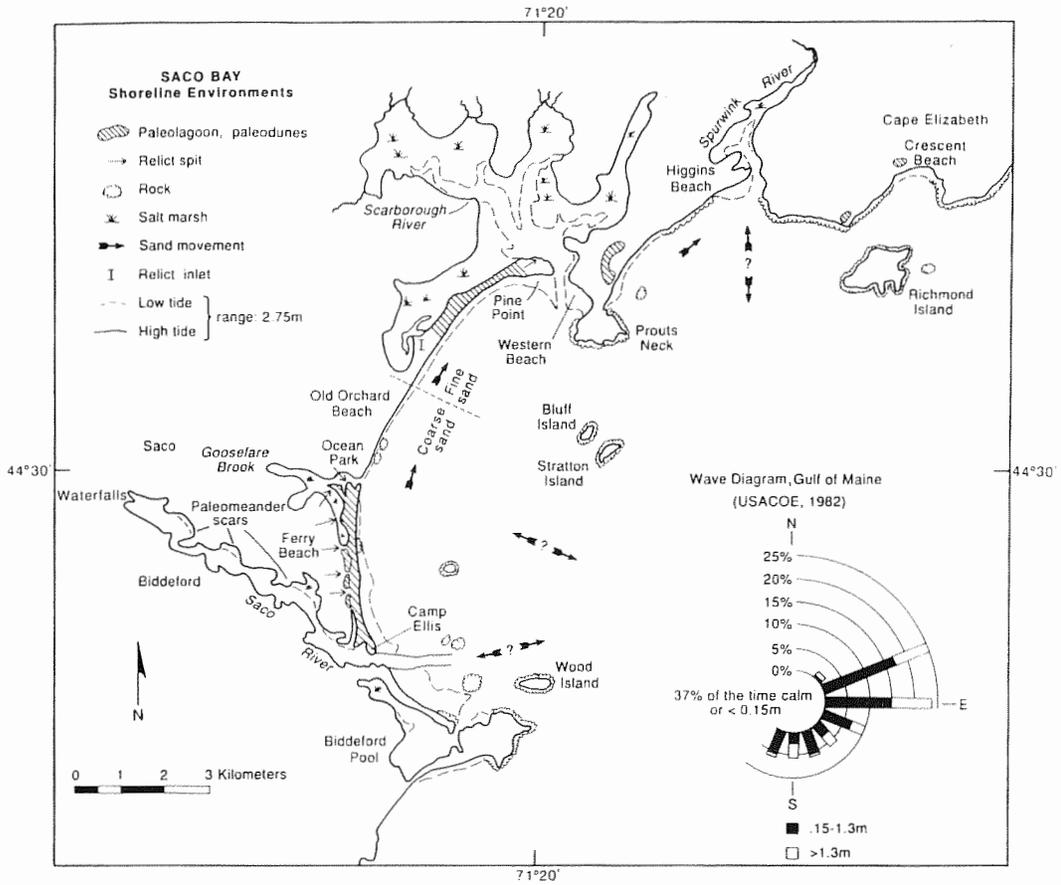


Figure 13. Saco Bay geomorphology (from Kelley et al., 1989).

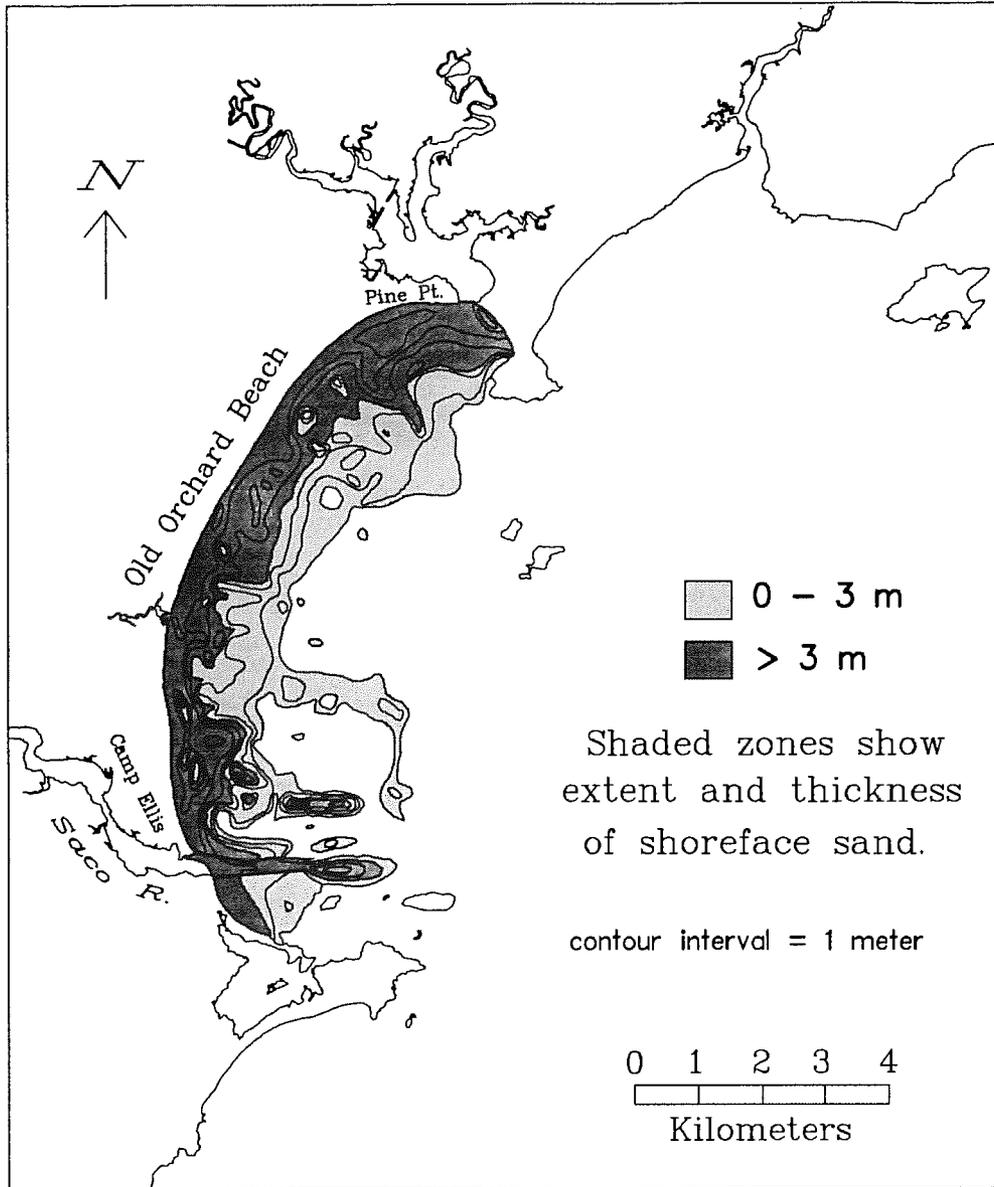


Figure 14. Thickness of shoreface sand in Saco Bay as evaluated by seismic reflection profiles (from Barber, in prep.).

Following the initial dredging of the river's tidal delta, sand accreted at Camp Ellis and was developed for residences. During this century, that sand has eroded and the beach has retreated despite repeated engineering and sand replenishment efforts. Many roads and a railroad line have been lost to the sea, and the Army is presently building a physical model of the area to evaluate the possibility that reflected waves from the jetty may be causing the erosion. The Army has previously recommended construction of a massive seawall along the eroding beach or a series of offshore breakwaters to save the properties. Neither of these solutions are permitted under Maine's sand dune law.

Directions: Return to Rt. 9. Drive north for 0.9 miles to Ferry Beach State Park, and park on side of road.

**STOP 2-2, FERRY BEACH STATE PARK (30 MINUTE)** The stratigraphy of the Saco Bay barriers is variable. Ground-penetrating-radar records, supplemented with vibrocore and bore-log data show evidence of a relatively thin barrier lithosome (averaging < 6 m in thickness), underlain in most places by glaciomarine clays of the Presumpscot Formation (Van Heteren et al., 1992). Locally, the barrier thickens, such as at Pine Point where an up to 20 m thick sand unit has been deposited. The average width of the barriers is 300 m, diminishing at headland anchor points.

Both transgressive and regressive characteristics are present in the Saco Bay barriers. The barrier lithosome is locally underlain by backbarrier tidal flat deposits indicating landward movement of the barrier system. The presence of former beach ridges behind the modern dunes, and ground-penetrating-radar records showing wide (up to 150 m) units of seaward-dipping reflectors are evidence of seaward-prograding barriers. The timing of these phases is presently unknown, but a scenario based on the relationship between rates of local relative sea-level rise and sediment input is expected. Hine et al. (1979) have related barrier development along the East Coast to the Late Holocene decrease in the rate of relative sea-level rise, when sediment accumulation rates began surpassing relative sea-level rates. The main regressive phase of the Saco Bay barriers may be related to this Holocene high sedimentation interval (Van Heteren et al., 1993). Transgressional and regressional phases of the barriers in the past 150 years have been recorded from charts and aerial photographs (Nelson, 1979), particularly at the spit ends and near the Saco River jetties; however, these short-term shoreline fluctuations cannot account for the observed changes along the presently more stable area extending from Ferry Beach to the central section of Goosefare Spit. In this region, remains of old beach ridges with mature soils are fronted by younger barriers that are substantially higher (Figure 15). The latter are partly the result of spit migration along the initially indented shoreline of Saco Bay, but seaward accretion has certainly widened the narrow paleobarriers. A stratigraphic cross section through Ferry Beach, based on vibrocore data and ground-penetrating-radar records, shows the transgressive and regressive phases discussed above (Van Heteren et al., 1993; Figure 15).

Directions: Continue north on Rt. 9 for 0.8 miles to a pullover next to a closed road into the Rachel Carson Wildlife Refuge.

**STOP 2-3, GOOSEFARE BROOK (30 MINUTES).** Goosefare Brook enters Saco Bay through Goosefare Brook Inlet. Relict spits in Goosefare Brook marsh (Nelson, 1979) record progradation of the barrier at a relatively stable inlet position (Figure 13). Marsh coring (Millette, 1992 and in prep.) has recovered peats up to 4 m thick over sandy lagoonal sediments and glaciomarine mud. No dates are yet available, but the marsh peats probably represent 4000 years of back-barrier stability, starting by colonization of tidal flats by low marsh, followed by high marsh colonization, resulting in broad high marsh meadows with extensive salt pannes. The fan-shaped series of spit tips represents sequential shoreline positions, and seaward progradation of the shoreline. This progradation requires input of sediment in the late Holocene.

Seismic reflection profiles demonstrate inlet scars offshore of the present inlet, preserved below a ravinement surface. A seismic profile (Figure 16) offshore of central Old Orchard Beach images the sequence of shoreface stratigraphy. Bedrock is overlain by draped glaciomarine sediment. The location of Goosefare Brook Inlet is stabilized by lows in the glaciomarine sediment, which in turn are controlled by bedrock. The glaciomarine sediment is cut by an angular unconformity, possibly partially a regressive or basal unconformity at the base of channels, and partially a transgressive ravinement surface. Immediately overlying the basal unconformity is backbarrier lagoonal mud with a basal peat and shells. The ravinement unconformity truncates this unit, and is overlain by shoreface sands. These sands are coarse, poorly sorted, and contain abundant shells and gravel. Sands and backbarrier sediments are preserved as a nearshore wedge off the barrier, that pinches out in approximately 20 m water depth. The preservation potential of the barrier is low (Belknap and Kraft, 1985), but some backshore deposits are preserved in depressions below the ravinement surface.

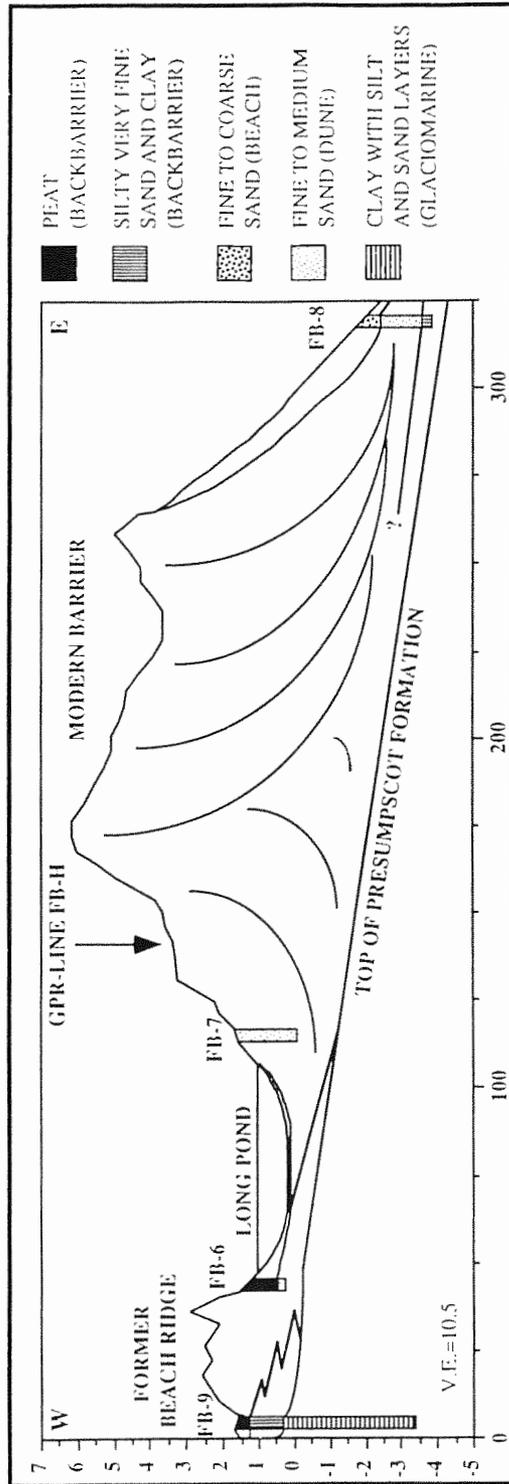


Figure 15. Stratigraphic cross section of the Ferry Beach area based on a topographic profile, ground-penetrating radar records, and vibracores (elevation in meters, 0 is msl)(from Van Heteren et al., 1993).

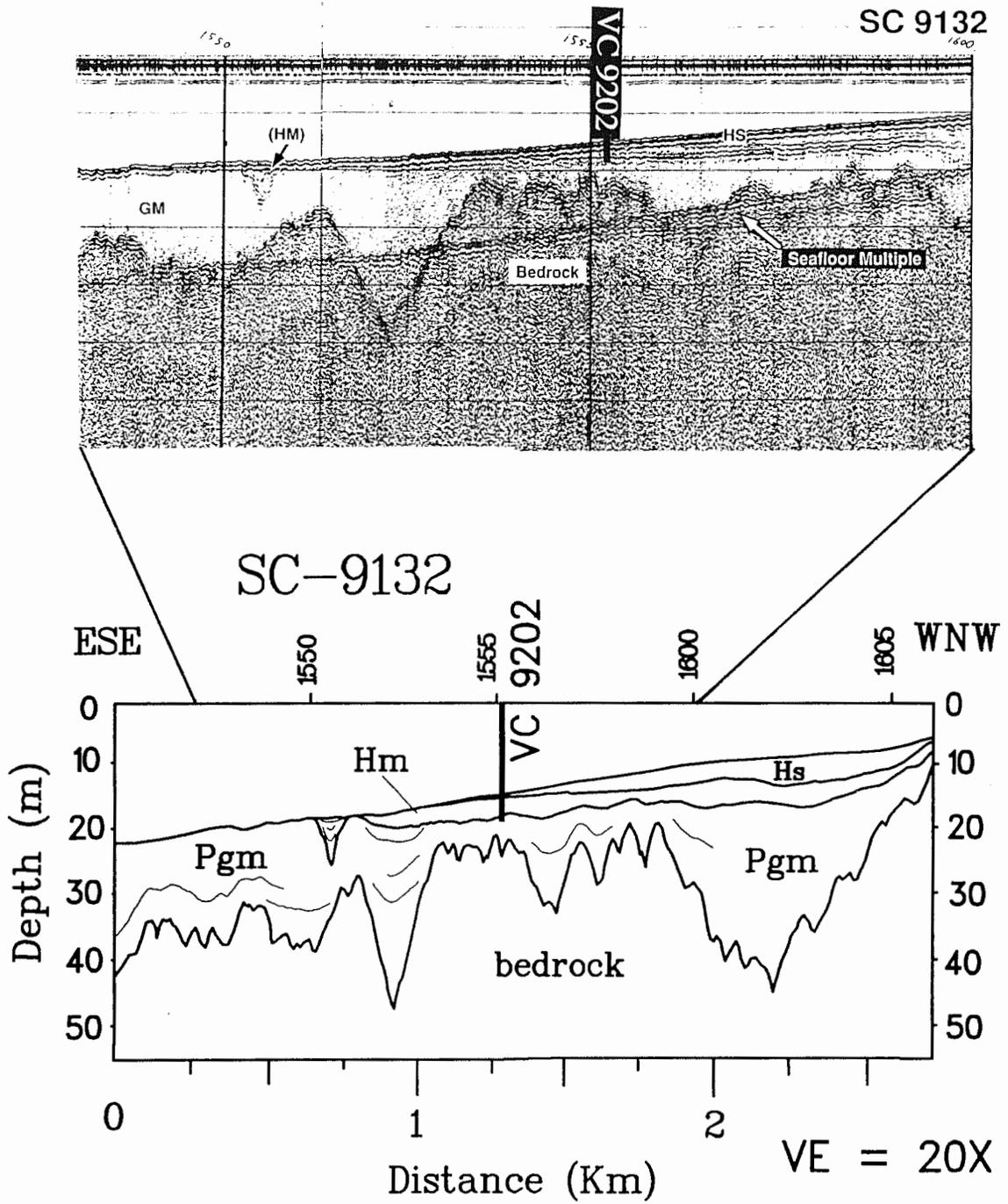


Figure 16. Seismic reflection profile seaward of Old Orchard Beach, with location of vibracore in shoreface sand. Pgm is a Pleistocene glaciomarine muddy unit, Hm represents Holocene mud, and Hs is Holocene sand (from Barber, in prep.)

An isopach map (Figure 14) of Holocene sediments in Saco Bay further illustrates this nearshore sediment wedge. Holocene sediments up to 7 m thick nearshore pinch out two kilometers offshore. This pinchout reflects shoreface reworking at the trailing edge of the transgressive barrier sequence, with coarser sediments carried landward and finer sediments dispersed seaward. The late Holocene progradation and resulting regressive sequence overlies the transgressive units. Unlike the barrier record discerned from GPR records, however, this cannot be clearly distinguished in the offshore.

Directions: Continue on Rt. 9 five miles to Pine Point. Where Rt. 9 makes a sharp left across the salt marsh, turn right and follow beach road to Fishermen's Coop Parking Lot at end of beach.

**STOP 2-4, PINE POINT (LUNCH, 90 MINUTES).** Pine Point is at the northeastern corner of the main Saco Bay barrier cell, adjoining the Scarborough River inlet across from Western Beach (Figure 13). While Western Beach has remained relatively stable historically, Pine Point has prograded eastward 500 meters between 1914 and 1962. In 1962, the inlet was stabilized by a rock jetty on the western side. Ferry Rock, on the eastern side of the inlet and western end of Western Beach is a natural stabilizer. Western Beach is typical of small sandy pocket beaches in Maine, enclosed between two bedrock headlands. Its sediment source includes both the sandy outwash and glaciomarine sediments from the Prouts Neck area as well as from the ebbtidal delta of the Scarborough River. Pine Point's sand is apparently derived ultimately from eroding sand at Camp Ellis. It has recently been suggested that sand dredged from the Scarborough River Inlet be returned to Camp Ellis as beach replenishment.

Directions: Return to Rt. 9, turn right, and follow three miles north to Rt. 1. Follow Rt. 1 for 7.5 miles to Interstate 295 and enter northbound. Exit at Freeport and follow Rt. 1 to Bow St. across from LL Bean in central Freeport. Follow Bow St. for 8 miles to junction with Pleasand Rd. and unnamed dirt road on left. Take unnamed road for 0.5 miles to end and park.

**STOP 2-5, BUNGUNAC BLUFF (90 MINUTES).** If time permits, we will visit Bungunac Bluff, one of the finest exposures of glaciomarine sediment in Maine (Kelley and Hay, 1986).

Directions: Return to Interstate 95 and head north to the Brunswick turnoff. Follow Rt. 1 here for 23 miles through Wiscasset to the junction with Route 27 south (towards Boothbay) in North Edgcomb, where we will spend the night at the Edgcomb Inn.

### DAY 3

Directions: Drive west on Route 1 0.5 mi. to the causeway across Cod Cove.

**STOP 3-1: COD COVE (15 MINUTES)** If time permits, we will examine a tidal flat and marsh system impacted by human alteration by causeways and harvesting of marine worms (Belknap et al., 1986; Wood, 1991).

Directions: Drive south on Rt. 1, and turn left after the Kennebec River bridge onto Rt 209, following the signs for Popham Beach State Park (16 miles). Park in the State Park lot.

**STOP 3-2, MORSE RIVER INLET (60 MINUTES)** Morse River is a small tidal stream that drains and fills Spirit Pond and a sizable marsh system. Repeated migrations of the main ebb channel of Morse River Inlet have caused extensive erosion along adjacent Popham State Park beach. This process has been described by Nelson (1979) and studied in detail by Goldschmidt et al. (1991). Easterly excursions of the main ebb channel is the result of the inlet throat being constricted by bedrock outcrops on both sides of the channel such that ebb discharge is directed to the east. As the channel migrates eastward, it lengthens, causing a decrease in the hydraulic slope and an increasingly inefficient exchange of water between the ocean and backbarrier. This situation is resolved when a new, shorter channel is cut through the spit platform, normally occurring during periods of storm activity. Easterly excursions of the Morse River inlet channel have resulted in remobilization of sand that has been in long-term storage along the State Park beach (FitzGerald et al., 1989).

Directions: We will walk east on the beach through the State park to privately owned Hunnewell Beach.

**STOP 3-3, POPHAM/HUNNEWELL BEACH (90 minutes)** The Kennebec River discharges between the north-south trending bedrock peninsulas of Georgetown and Cape Small. Barriers at the mouth of the Kennebec Estuary are unevenly distributed, which is the result of the dominant east-northeast wave energy in the Gulf of Maine and the structural trend of the shoreline. As sand from the Kennebec paleodelta (Belknap et al., 1987) was reworked onshore during the Holocene

transgression, it is likely that the dominant east-northeasterly storm waves moved the sand in a net westerly direction where it gradually accumulated along the southwestern shoreline of Cape Small, forming the Popham and Seawall Beaches. On the opposite side of the river mouth only Reid Beach, a minor barrier, exists along the southeast-facing shoreline of the Georgetown Peninsula. A large ebb-tidal delta, identified on coastal charts as Pond Island Shoal, is situated at the mouth of the Kennebec Estuary. Like the onshore barriers, the delta is asymmetric about the river mouth. It is almost entirely situated on the west side of the main ebb channel, another indication of the historically dominant westerly transport trend.

Popham Beach is the largest barrier system associated with the Kennebec River paleodelta. It forms the western shoreline of the present-day river mouth. Its evolution and sedimentation patterns have been strongly influenced by both wave and tidal processes. The barrier is anchored to a number of north-south trending bedrock ridges that crop out offshore as elongated islands and submerged ledges (Figure 17). The barrier is cusped in plan form and is approximately 3 km long, extending from the Kennebec River west to the diminutive Morse River Inlet. It is backed by an expansive bay (Atkins Bay), and a wide tidal flat and marsh system. Two of the islands off Popham Beach are connected to the shore by intertidal tombolos. The sediment comprising the barrier ranges in size from fine to coarse sand. Muddy sands occur in Atkins Bay and in the peripheral marsh areas.

Popham Beach is a regressive barrier that has formed from a large offshore sediment supply. The sediment abundance is reflected in the wide sandy offshore area (Belknap et al., 1989a), the thickness of the barrier lithosome including an extensive vegetated dune system, and the areal extent of the barrier. Auger cores and ground-penetrating-radar profiles indicate that the barrier sands are at least 8 m thick on average, and considerably thicker in places (Fink and FitzGerald, unpub. data). Radar sections exhibit successive, sigmoidal-shaped reflectors that have seaward-dipping angles similar to those of the present berm and shoreface. It is apparent that the barrier prograded seaward by the addition of successive accretionary wedges. The timing of initial barrier development and its subsequent evolution are not well known; however, radiocarbon dates of salt-marsh peats from Atkins Bay and the area behind Morse River are in progress.

The present-day cusped shape of Popham Beach seems to imply that the shoreline is swash-aligned (Figure 17 and there is little longshore sediment transport. However, the proximity of the Kennebec River and the strong tidal currents that exist at its mouth produce longshore currents up to 1 m/s along the adjacent beach (FitzGerald et al., 1989). The resulting longshore sand-transport system is part of a large clockwise sediment gyre at the mouth of the river, which circulates sand among Popham Beach, the seaward portion of the Kennebec River, and Pond Island Shoal (FitzGerald and Fink, 1987). Sand moves onshore in the form of a large bar complex which can extend along the beach for more than 1 km and is fronted by a 2 to 3 m high slipface. Bar complexes weld to the shoreline every 6 to 10 years, adding  $0.3$  to  $0.5 \times 10^6 \text{ m}^3$  of sand to the beach.

During the past century Popham Beach has experienced shoreline fluctuations greater than 150 m over periods ranging from 15 to 50 years (Nelson, 1979; FitzGerald and Fink, 1987; Fink et al., in press)(Figure 18). The major driving force behind these changes is the aforementioned sand-circulation process. Other, less uniform factors affecting long-term shoreline trends at Popham Beach include: 1. frequency of major storms, 2. migrational behavior of Morse River (Goldschmidt et al., 1991), 3. configuration of the Wood Island bar (Fink et al., in press), 4. major floods of the Kennebec River, and 5. sea-level trends. There have been major transferrals of sand among Riverside Beach, Hunnewell Beach and the beach between Morse River and the Fox Island tombolo. For example, the far landward position of the Hunnewell Beach shoreline in 1856 was coincident with the farthest seaward shoreline of both Riverside Beach and the beach west of Morse River. By 1953 the converse was true.

Shoreline change has figured prominently in the development history of Hunnewell Beach as well as in the evolution of Maine's sand dune laws. The loss of houses along this beach in the middle 1970's led to a widespread public awareness of the danger posed by houses built too close to the sea. Following the 1978 storms, Maine's sand Dune law was enacted, and its standards prohibited seawalls and the reconstruction of severely damaged houses. It was made illegal to construct a house in an area where it was likely to be damaged by shoreline change in the next century. When a house was constructed without a permit at Hunnewell Beach on the site of a building destroyed in 1976, the State required that it be torn down. The Maine Supreme Court upheld the decision and awarded no compensation to the owner of the building. Considerable debate had ensued over the future location of the Hunnewell Beach shoreline. Those favoring removal of the house believed that the site had no future. In the past 5 years the beach has grown seaward by 50 m, but the long-term future remains clouded.

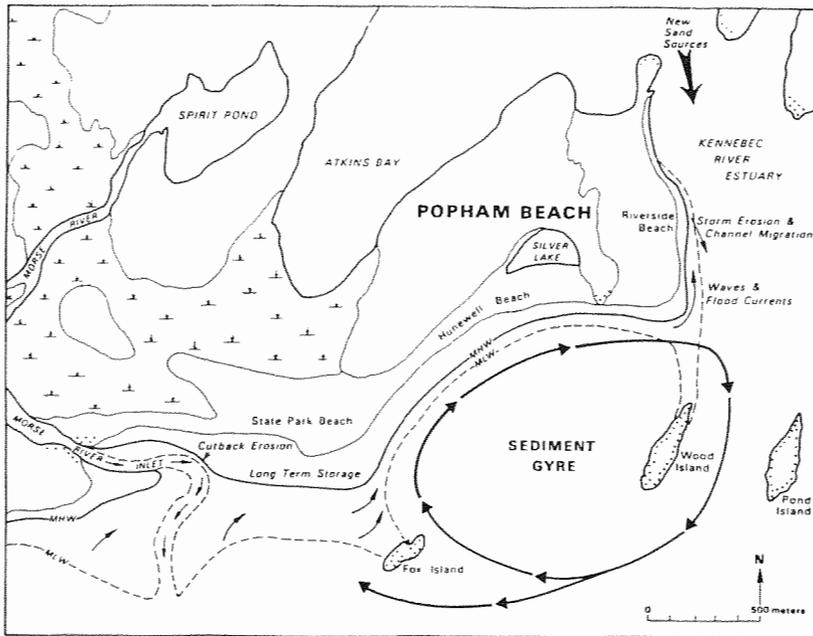


Figure 17. Sand circulation at the Kennebec River mouth.

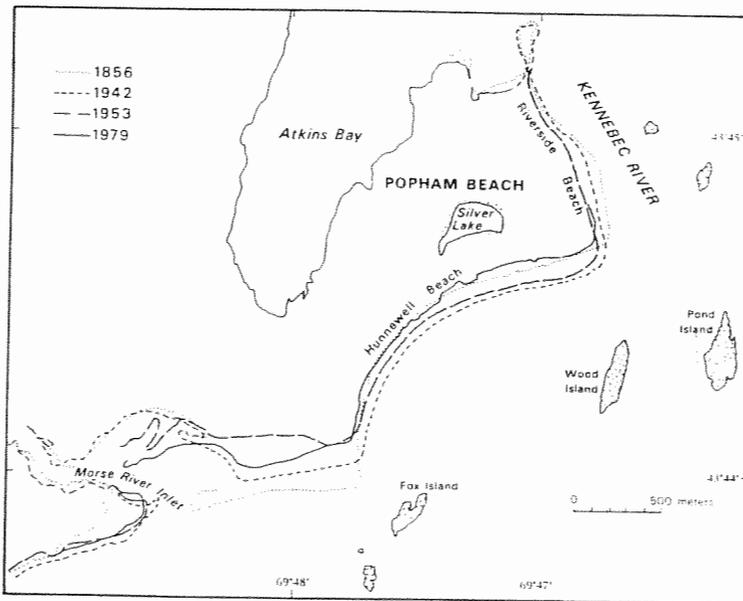


Figure 18. Shoreline change at Popham Beach (from FitzGerald and Fink, 1987).

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Chapter H

# Alleghanian Assembly of Proterozoic and Paleozoic Lithotectonic Terranes in South Central New England: New Constraints from Geochronology and Petrology

By R.P. Wintsch, J.F. Sutter, M.J. Kunk, J.N. Aleinikoff, and J.L. Boyd

Field Trip Guidebook for the Northeastern United States:  
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**ALLEGHANIAN ASSEMBLY OF PROTEROZOIC AND PALEOZOIC  
LITHOTECTONIC TERRANES IN SOUTH CENTRAL NEW ENGLAND:  
NEW CONSTRAINTS FROM GEOCHRONOLOGY AND PETROLOGY**

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**PURPOSE OF FIELD TRIP**

This field trip is designed to contrast the rock types, metamorphic grades, and metamorphic histories of the rocks of southeastern New England that constitute its lithotectonic terranes: the Avalon, Putnam-Nashoba, Merrimack, Central Maine, and Bronson Hill (Fig. 1). We will focus on the lithologic, metamorphic, and geochronologic evidence for the definition of lithotectonic terranes, and show how contrasting Pressure-Temperature-time (PTt) paths reflected by the mineral cooling ages of the rocks in the various terranes require that significant assembly of these terranes post-dates middle to late Paleozoic metamorphism. We will show that much metamorphic history can be extracted from a metamorphic rock by dating its minerals, in spite of a low variance mineral assemblage that does not narrowly constrain the metamorphic conditions. We use the term "terrane" to refer to a body of rock isolated by faults, with internal lithologic continuity, and a uniform metamorphic history.

The most contentious issue regarding the timing of terrane assembly has centered on the arrival of the Avalon terrane. Some argue for a Middle Ordovician arrival, causing the Taconic orogeny; some favor an early Devonian arrival causing the Acadian Orogeny; and some hold that at least the last stages of assembly occurred during the late Paleozoic Alleghanian orogeny. Stockmal *et al.* (1990) and van der Pluijm *et al.* (1990) have proposed that initial collision occurred in the Silurian, and continued in a protracted way into the late Paleozoic. Wintsch *et al.* (1992) go on to suggest that the arrival of western parts of Avalon (or its outboard terranes) underplated the eastern terranes, uplifting them, and causing the end of the Acadian orogeny. On this field trip we will examine some of the evidence for the latter proposal by contrasting middle and late Paleozoic PTt paths of the various terranes, and show how they provide evidence for the Late Paleozoic assembly of these terranes.

**METAMORPHIC PRESSURE-TEMPERATURE-TIME PATHS**

Identification of lithotectonic terranes has been based on both lithologic assemblages and on metamorphic history. Thermochronologic data in the form of mineral ages strongly compliment P-T data derived from mineral thermobarometry by defining the *time* interval of metamorphism. Mineral ages are well known to be reset by heating events, and in fact the ages of many minerals in igneous or high grade metamorphic rocks reflect the time of cooling from the last metamorphism, rather than the time of crystallization. Closure temperatures, or the temperatures below which minerals effectively retain radiogenic daughter isotopes, are for lead: monazite  $720 \pm 20^\circ\text{C}$ , and sphene  $575 \pm 50^\circ\text{C}$  (Cliff and Cohen, 1980; Copeland *et al.*, 1988); and for argon: hornblende,  $500 \pm 50^\circ$ ; muscovite,  $350 \pm 25^\circ\text{C}$ ; biotite,  $300 \pm 25^\circ\text{C}$ ; and K-feldspar,  $200 \pm 50^\circ\text{C}$  (McDougall and Harrison, 1988),

On this trip we take advantage of mineral ages in several ways. For instance, in rocks east of the Willimantic window (Fig. 2), hornblende ages reflect cooling from pre-Alleghanian metamorphism. However, in and southwest of the window, their younger Permian ages represent middle amphibolite facies Alleghanian overprinting. U-Pb ages of sphene behave in the same way except that sphene crystals that grow below their closure temperatures reflect the time of *crystallization*, because temperatures were not high enough to allow the diffusive loss of radiogenic lead. In the Avalon terrane, the Pennsylvanian ages of sphenes reflect prograde metamorphism to lower amphibolite facies conditions. Integration of these and other geologic and petrologic data provide very strong evidence for the Alleghanian assembly of these terranes. Some of the data upon which the trip is based is presented in Wintsch *et al.* (1991; 1992), and Wintsch (1992). Prior reference to these is suggested.

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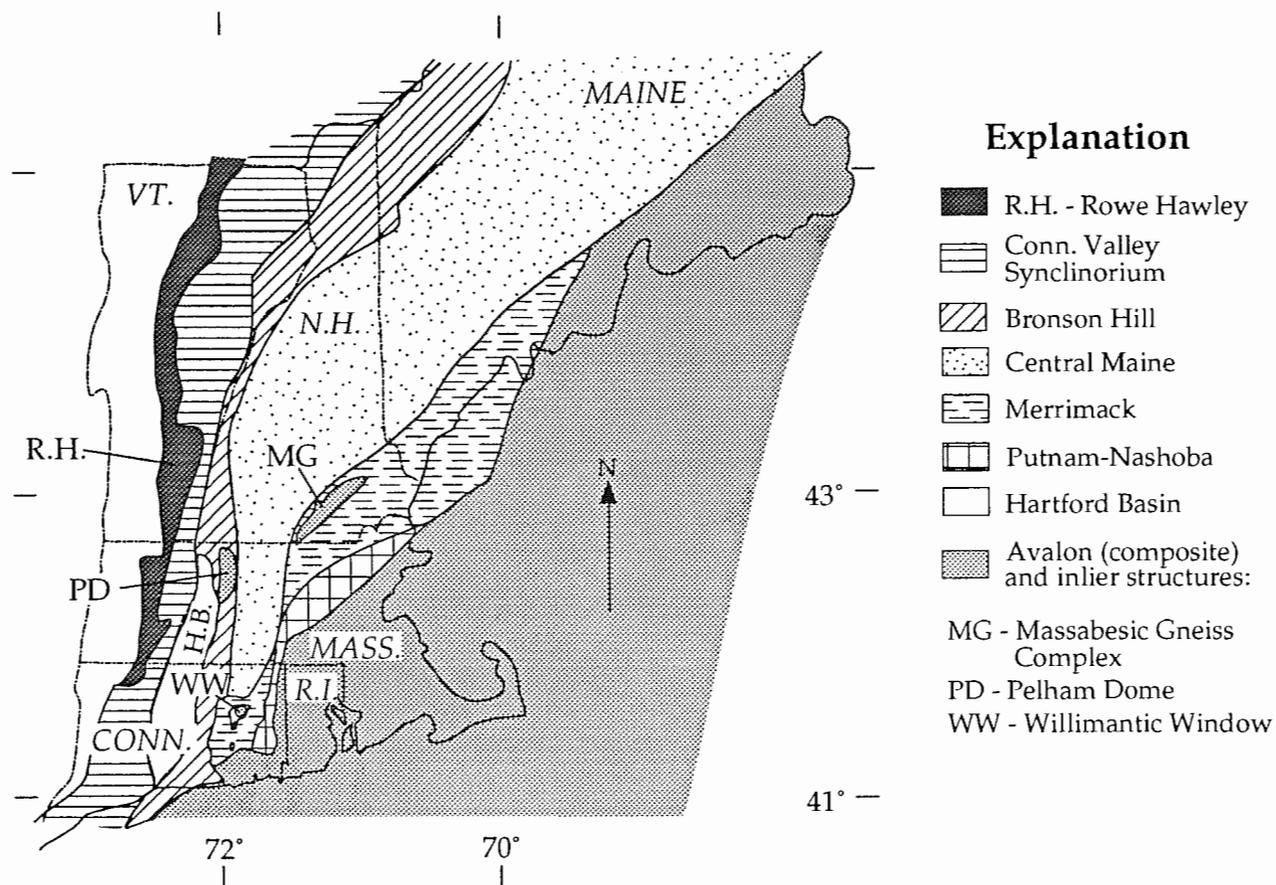


Figure 1. Generalized map of New England, showing the distribution of the lithotectonic terranes (modified from Zartman, 1988; Hutchinson et al., 1988; Robinson and Goldsmith, 1991). The locations of the Avalonian inliers in domal structures is indicated.

### METAMORPHIC AND GEOLOGIC SETTING

The distribution of the high grade metamorphic rocks that comprise the terranes in the area of this field trip is shown in Fig 2. The overall NNE strike of the terranes reflects the gentle dips to the west of all the rocks, except in the vicinity of the Willimantic window. The Avalon terrane underlies all terranes to the west, and is exposed again in the Willimantic window, as are rocks of the Putnam-Nashoba terrane. All terranes are separated from one another by ductile (mylonitic) faults, which commonly mark discontinuities in metamorphic grade and/or regional cooling histories. Field trip stops included on this trip (Fig. 2) are selected to show the maximum information about the terrane, primarily because of excellent exposure, but also because of the availability of petrologic and thermochronologic information.

#### Avalon Terrane

The Avalon terrane consists of Late Proterozoic rocks of coastal New England, and three inliers of rocks in windows or domes in central New England west of the main coastal outcrop area (Fig. 1). These three structures (the Willimantic window, Pelham dome, and Massabesic Gneiss complex) share with the main body of the Avalon terrane a Late Proterozoic crystallization age of igneous rocks and only a late Paleozoic metamorphism, and are thus correlated with it (Wintsch *et al.*, 1992). The presence of Avalonian rocks in these structures indicates that Avalonian rocks structurally underlie all terranes to the west at least as far as the Hartford basin. This common history further suggests that faults that cut this terrane (e.g. Hunts Brook, Hope Valley, Beaverhead, Fig. 2) are less regionally significant than those that separate Avalon from structurally higher terranes.

In the area of this field trip, the Avalon terrane contains primarily Late Proterozoic felsic plutonic (Stops

### Explanation

- terrane boundary
- fault
- - - metamorphic isograd
- ◇ 249 sample location and hornblende age (Ma)
- 20** location and number of field trip stop

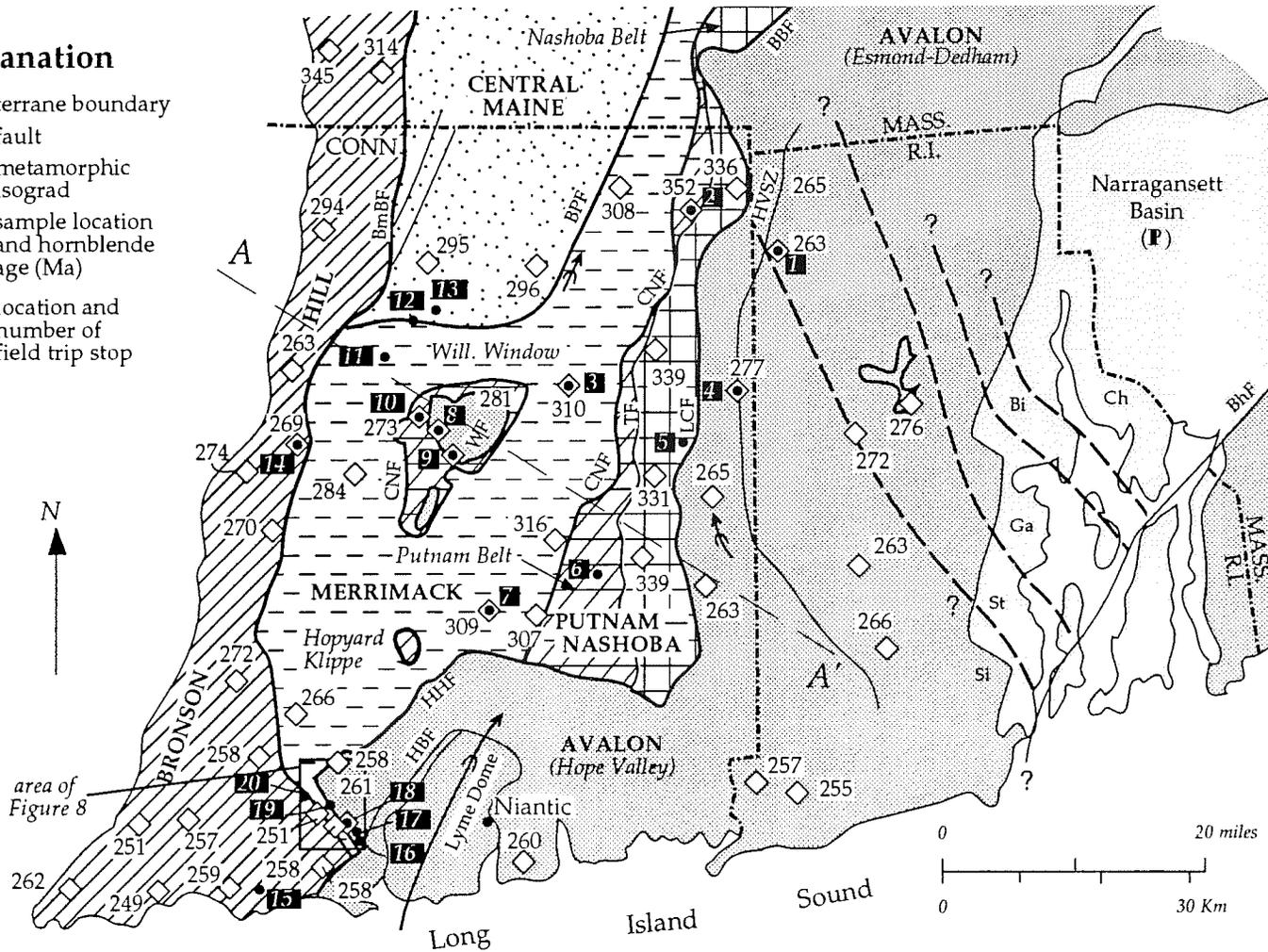
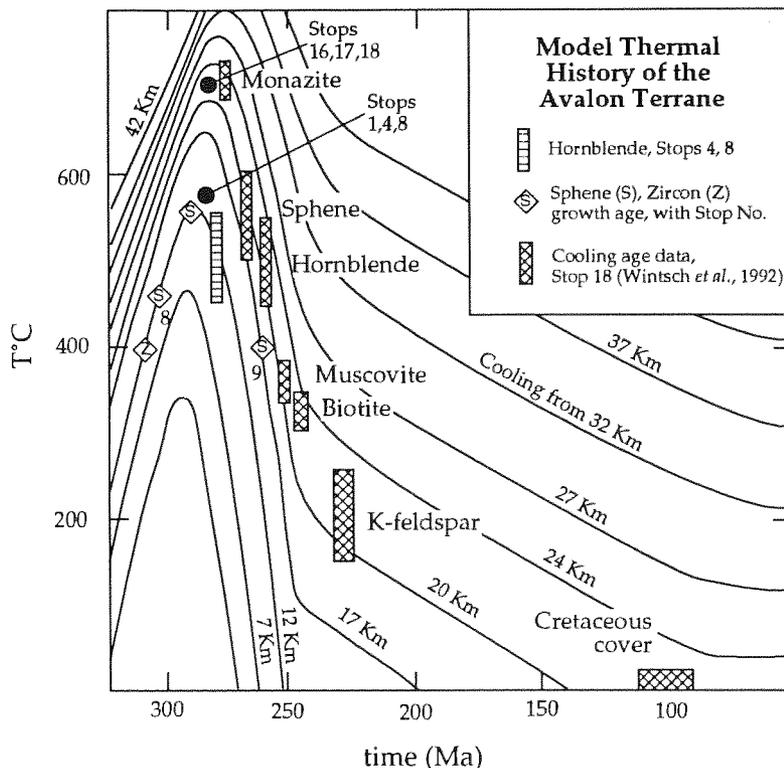


Figure 2. Map of southeastern New England, showing the distribution of terranes (after Wintsch, *et al.*, 1992). Faults at terrane boundaries are: HHF, Honey Hill fault; LCF, Lake Char fault; BPF, Bloody Bluff fault; BNI, Clinton-Newbury fault; BPF, Black Pond fault; BmBF, Bonemill Brook fault. Faults within terranes: HBF, Hunts Brook fault; HVSZ, Hope Valley Shear zone; BhF, Beaverhead fault; TF, Tatnic fault. Alleghanian isograds in Rhode Island separate the chlorite (Ch), biotite (Bi); Garnet (Ga); Staurolite (St); and Sillimanite (Si) zones. The location of Niantic (east of Lyme Dome), where we are staying, is indicated.

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Figure 4. Cooling history of the southwestern part of the Avalon terrane, and model thermal history (from Wintsch, et al. 1992). Hornblende cooling ages from the structurally higher parts of the Avalon terrane (Stops 4, 8) are superimposed on the model cooling curves. Temperature coordinates for sphene and zircon ages interpreted as crystallization ages from Stops 8, 9 (Getty and Gromet, 1992b) and north of Stop 19 (unlabeled, Aleinikoff, unpublished) are plotted to be consistent with the calculated Temperature-time curves.



1,16,18), volcanic (Stops 4, 8), and minor metasedimentary (Stop 4) rocks. These are cut by Ordovician, Devonian, and late Paleozoic intrusive rocks, and several pre-, syn-, and post-metamorphic faults. The structural top of the Avalon terrane is defined by the Honey Hill (Stop 18) - Lake Char (Stop 5) - Willimantic (Stop 9) (and Bloody Bluff, Massachusetts) fault system. In the west, the fault was last active under middle to upper amphibolite facies conditions, but in the east it was active (or reactivated) at higher structural levels and greenschist facies conditions. Metavolcanic and metasedimentary rocks are cut out at a low angle and folded along the foot wall of the Honey Hill fault system (Fig. 3).

The grade of metamorphism varies from unmetamorphosed Late Proterozoic igneous rocks and Cambrian and Pennsylvanian sedimentary rocks in eastern Massachusetts to upper amphibolite facies in southwestern Rhode Island, and southern Connecticut. Permian isograds (Fig. 2) have a general northwest strike, with the Ponaganset Gneiss in northwestern Rhode Island lying approximately between the staurolite and sillimanite isograds (see Stop 1). In the Late Proterozoic rocks of eastern Connecticut this metamorphic gradient is manifest in muscovite + biotite  $\pm$  garnet-bearing rocks conspicuously lacking migmatites and feldspar-bearing veins (Dixon, 1974; Moore, 1983; Stops 1, 4), whereas to the south and west, abundant pegmatites and migmatites reflect anatectic conditions (Lundgren, 1966; Wintsch and Aleinikoff, 1987; Dipple *et al.*, 1990; Stops 16-19). Amphiboles in structurally higher parts of the Avalon terrane are fine grained and acicular (Stop 4), but south and west of Stop 4 (e.g. Stops 8, 16, 18) they are coarse grained, nearly equant, show straight extinction, and generally share smooth, straight boundaries with all adjacent grains.

The time of metamorphism of these rocks is quite uniform; hornblende cooling ages are all late Paleozoic (Table 1; Fig. 2), which precludes any km-scale tilting since the late Permian. However, the apparent metamorphic gradient south of the Honey Hill fault strongly suggests 10 km or more tilting of the Avalon terrane toward the north. A similar tilting to the west under the Lake Char fault is also likely. The history of metamorphism of the Avalon terrane rocks reflects relatively rapid loading and unloading during the Alleghanian orogeny. Rocks close to the structural top of this terrane, modeled by the 12 km isopleth of Fig 4, were heated and grew metamorphic sphene in the Pennsylvanian, reached peak metamorphic temperatures of  $< 600^{\circ}\text{C}$  at about 280 Ma, and have hornblende cooling ages of slightly less than 280 Ma. Rocks deeper in the terrane have monazite ages of  $\sim 280$  Ma (Stop 18), and younger Permian sphene and hornblende cooling ages (Stops 1, 18). The exposure of these deeper rocks reflects folding or tilting of the Avalon terrane between 280-265 Ma.

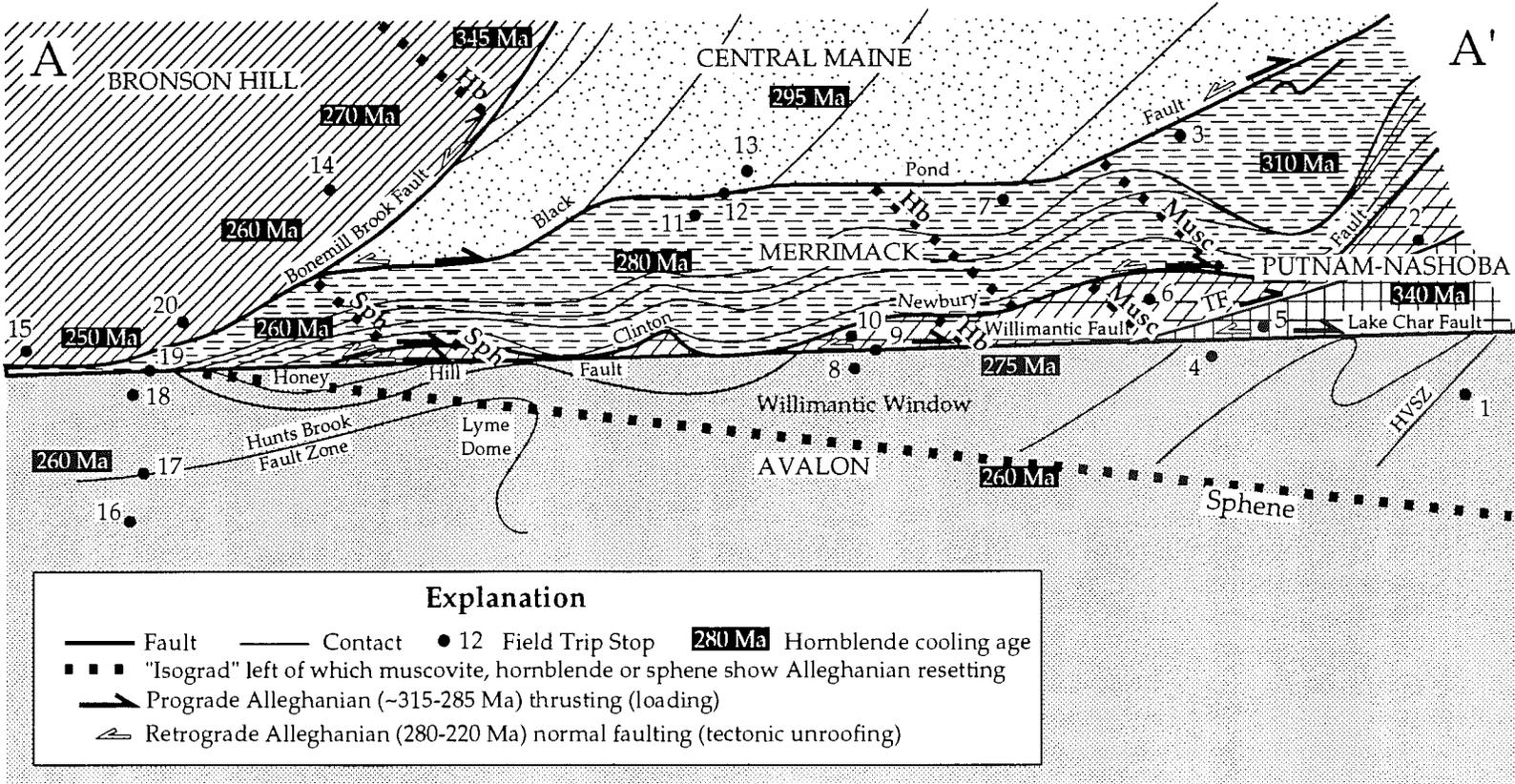


Figure 3. A partially schematic section across lithotectonic terranes in SE New England reconstructed from regional, quadrangle, and detailed (Wintsch, unpublished) mapping. The unfolded Honey Hill-Lake Char-Willimantic fault system is the plane of reference. The geometry of the terranes is projected onto plane A-A' (Fig. 2) with considerable vertical exaggeration. Alleghanian metamorphic history of each terrane is indicated by the sphene, hornblende, and muscovite 'isograds', that separate rocks containing mineral ages reset in the Alleghanian on the West from those to the east with unreset, pre-Alleghanian ages. A net thrust motion ( ➔ ) is required by the structural position of terranes with younger cooling ages overlying terranes with older ages; significant thrust motion must postdate peak metamorphic conditions. Later (Permian-Triassic) reactivation of these faults in a normal sense ( ↔ ), contributed to tectonic exhumation.

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Table 1. Summary of Mineral Apparent Ages

Terrane	West / SW	Center	East
<b>Bronson Hill, Stops</b>	<b>14,19</b>		
Sphene	350 / 284		
Hornblende	345 / 250		
Muscovite	250		
Biotite	245		
K-feldspar	220		
<b>Central Maine</b>			
Monazite		365 (345)	
Sphene			
Hornblende		295	
Muscovite		250	
Biotite		248	
K-feldspar		>230	
<b>Merrimack, Stops</b>	<b>19</b>	<b>11</b>	<b>2,6</b>
Sphene	305 / 281		315
Hornblende	266 / 251	285	310
Muscovite			250
Biotite	243		
K-feldspar			230
<b>Putnam-Nashoba, Stops</b>		<b>9,10</b>	<b>2,6</b>
Monazite		400	
Sphene	- / 285	335	350
Hornblende	- / 264	280	340
Muscovite			275
K-feldspar			247
<b>Avalon, Stops</b>	<b>18</b>	<b>8,9</b>	<b>1,4</b>
Zircon	270 L.I. (310)		
Monazite	278		
Sphene	270 (295)	(305)	600 (315?)
Hornblende	260	280	277
Muscovite	250	247	250
Biotite	243	243	245
K-feldspar	230	228	>232

Table 1. Monazite, sphene data: U-Pb system. Hornblende, muscovite, biotite, K-feldspar data:  $^{40}\text{Ar}/^{39}\text{Ar}$  system. All ages are cooling ages except those in ( ), which refer to crystallization ages. L.I. refers to lower intercept of discordia. Columns refer to positions on section A-A', Fig. 2, where West, Center and East refer to the western, Willimantic, and eastern parts of the section. Bold numbers refer to field trip stops. SW refers to the area of Figure 8 (Deep River). Data is from this study and Wintsch *et al.* (1992), Tucker and Robinson (1990), Getty and Gromet (1992b), Thomson *et al.* (1992), and Boyd *et al.*, (1992).

**Putnam-Nashoba Terrane**

Rocks of the Putnam-Nashoba terrane are exposed in eastern Massachusetts (Nashoba belt) and in eastern Connecticut (Putnam belt), and immediately overlie the Avalon terrane. These rocks are Late Proterozoic (?) pelitic metasedimentary and mafic metavolcanic rocks (Goldsmith, 1991; Stops 2, 6, 10). They were intruded by several post-metamorphic Silurian mafic rocks (Zartman and Naylor, 1984) and cut by several ductile faults (Dixon and Lundgren, 1968; Zen *et al.*, 1983), including the Tatnic fault (Fig. 2). The base of this terrane is strongly cut out by the Honey Hill fault system, such that west of stop 6 the lower mafic volcanic unit (Quinebaug Formation) is cut completely out, and farther west the entire terrane is missing (Fig. 3, Stop 19). The top of the terrane is cut out by the Clinton-Newbury fault.

Upper amphibolite facies metamorphic conditions of  $\sim 700^\circ\text{C}$  and 5.7 kb (Moecher and Wintsch, 1991; in review) predate Silurian ( $\sim 425$  Ma, Zartman and Naylor, 1984) intrusions in the Putnam belt and monazite mineral ages in the Willimantic window ( $\sim 400$  Ma, Getty and Gromet, 1992b). Available thermochronology suggests very slow cooling ( $3^\circ\text{C}/\text{m.y.}$ ) from a Silurian or older metamorphism to the Triassic (Table 1; Fig. 5). Data from both the Putnam belt and the Willimantic window define parallel cooling curves of nearly identical age that support the lithotectonic correlation of the two belts of rock. In eastern exposures (Stops 2, 6) resetting by Alleghanian metamorphism was minimal; even muscovite apparent ages are consistent with slow cooling (Fig. 5). However, in the Willimantic window (Stop 10) Alleghanian metamorphism ( $\sim 600^\circ\text{C}$ , 8  $\rightarrow$  5 kb) has reset hornblende but not sphene, suggesting first sillimanite-grade Alleghanian metamorphic overprint (Moecher and Wintsch, 1991; in review).

**Merrimack Terrane**

Rocks of the Merrimack terrane structurally overlie Putnam-Nashoba rocks (Figs. 2, 3). Rocks of particular concern to this trip are now correlated with the Oakdale

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Formation in Massachusetts (Pease, 1989; Robinson and Goldsmith, 1991) in Zartman's (1988) Merrimack Trough zone. East of the Willimantic window the rocks are dominated by calcareous metasilstones (Oakdale [Stops 3, 7]) and argillaceous metasediments (Scotland Schist member [Stop 7]). The base of the terrane is marked by the Clinton-Newbury fault, except in the west where the base is cut by the Honey Hill fault (Stop 19, Fig. 3). In the north of Fig. 2, the top is cut by the Black Pond fault, which locally cuts out parts of the upper section, and is locally associated with low amplitude asymmetric folds. In the southwest, the top is cut out by the Bonemill Brook fault (Fig. 3).

These rocks are metamorphosed to staurolite-kyanite zone of the epidote-amphibolite facies. Preliminary thermochronology in these rocks shows that hornblendes in the east have not been affected by Alleghanian metamorphism; they show plateau or isochron ages of about 310 Ma whereas south of the Willimantic window they yield Permian ages (Fig. 2) suggesting Alleghanian overprinting. Muscovite from the Scotland Schist member shows a ~250 Ma age, which is consistent with slow cooling from pre-Alleghanian metamorphism (Fig. 5), and cannot be unambiguously assigned to cooling from Alleghanian metamorphic overprinting.

**Central Maine Terrane**

Rocks in the Central Maine terrane are dominated by calcareous and pelitic metasediments, metamorphosed to

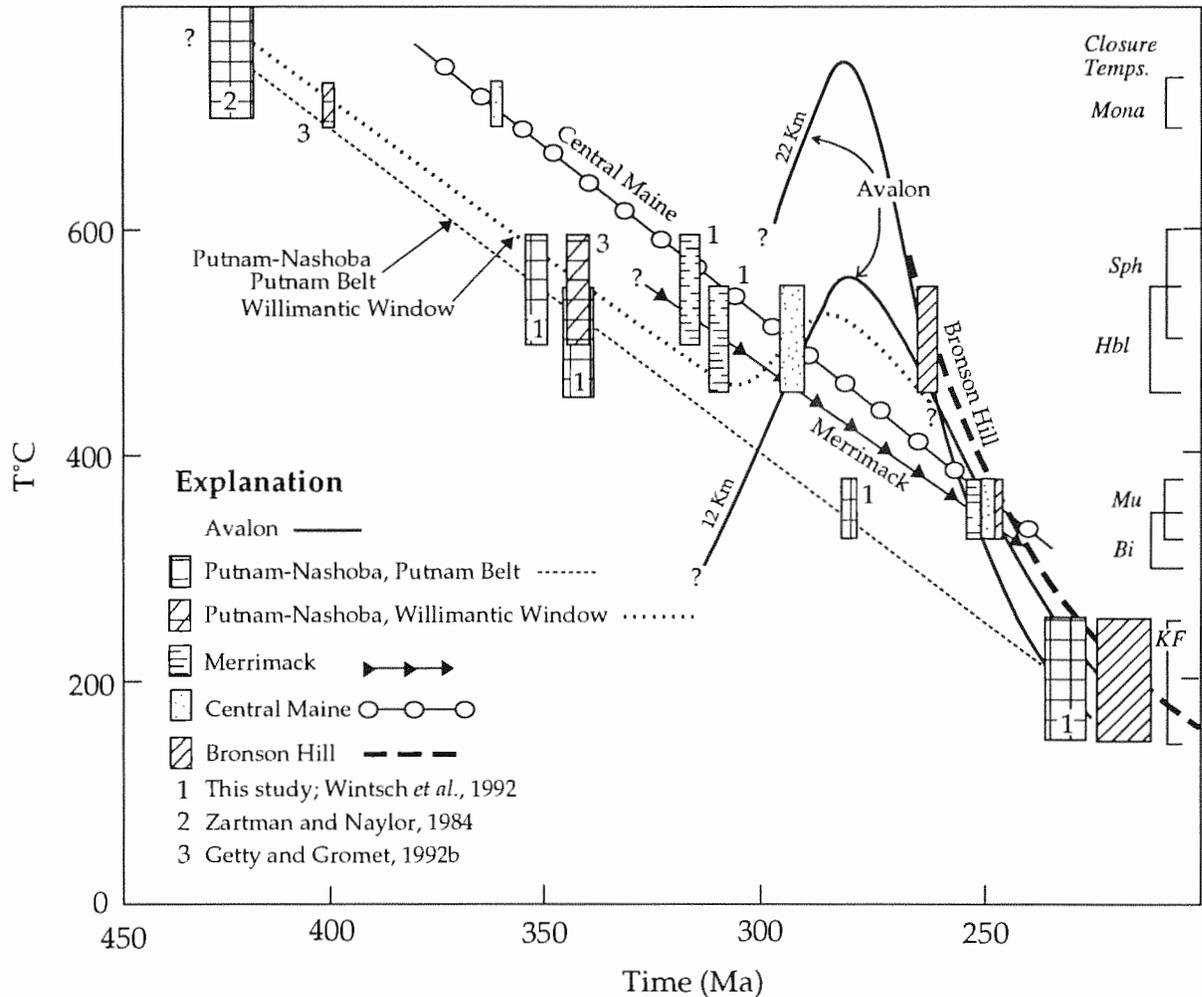


Figure 5. Comparison of the thermochronologic data from the Avalon, Putnam-Nashoba, Merrimack, Central Maine, and Bronson Hill terranes. Although some data are preliminary, the trend of decreasing cooling age from the eastern Putnam-Nashoba to the western Bronson Hill terrane is evident.

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upper amphibolite and lower granulite facies conditions (Thompson *et al.*, 1992). Summaries of rock descriptions are given in Peper *et al.* (1975) and Robinson and Goldsmith (1991). The basal Black Pond fault (Figs 2, 3) cuts out the base of most units. The upper part of the terrane is cut by the Bonemill Brook fault.

Available geochronology on monazites (Thompson *et al.*, 1992) suggests peak metamorphic conditions ceased by the late Devonian (~365 Ma) with younger monazite ages probably recording crystallization events. Our preliminary hornblende and muscovite ages of ~295 Ma and 250 Ma, respectively, suggest the end of amphibolite facies metamorphism by the Permian. Again, late Permian muscovite ages are consistent both with slow post-Acadian cooling, and with cooling from Alleghanian overprinting (Table 1).

### **Bronson Hill terrane**

Rocks of the Bronson Hill terrane are dominated by granodioritic orthogneisses (Stop 14), and dacitic, dioritic and basaltic metavolcanics (Stop 20); Webster and Wintsch, 1987; Leo *et al.*, 1984), but metapelitic, calcsilicate and manganese-rich (cotecule) rocks are also present. Igneous activity is late Ordovician (Zartman and Leo, 1985; Tucker and Robinson, 1990). The base of the Bronson Hill terrane is marked by the Bonemill Brook fault. It deforms an amphibolite along much of its length, and locally cuts it out completely. A 100-million-year north-to-south age gradient in hornblende ages (Fig. 2) shows cooling from middle amphibolite facies Acadian metamorphism in the north, and from upper amphibolite facies Alleghanian metamorphism in the south. The latter is probably an Alleghanian overprint on an earlier metamorphism.

### **DUCTILE FAULTS AND TERRANE BOUNDARIES**

All terranes are separated by ductile faults. All are pre- or syn-metamorphic, but some have been reactivated under post-peak metamorphic greenschist facies and even brittle conditions. Consequently, a wide variety of fault rock types are present among terrane boundaries, from mylonitic schists and gneisses (and straight gneisses [Stops 2, 5, 10, 12, 17, 19, 20]), to blastomylonites, phyllonites, and cataclases (Stops 5, 10, 11, 12, 14). The spectrum of faults and fault rocks (Fig. 6) is remarkable because some of the most spectacular mylonitic fault zones occur within terranes (Stop 17), and some of the most regionally significant faults are the least impressive in outcrop as high strain zones. Lithologic boundaries are quite sharp, with a minimum of imbrication (Stops 9, 19).

Tectonic blocks are common in many ductilely deformed rocks; they probably formed during extension under simple shear, and can be thought of as huge boudins in various stages of rotation (Wintsch, 1979; Fig 7). They are characteristic of the pelitic unit in the Putnam-Nashoba terrane, especially near the base of the Tatnic Hill Formation and can be seen in various stages of development (compare Stops 2, 6, 10) along the Tatnic and Willimantic faults, and at the Black Pond fault and at Stop 12. They are also exposed along the Honey Hill fault (Lundgren, 1968, Stop 3; Wintsch, 1985, Stop 7).

Most ductile faults have been intruded by pegmatite bodies, commonly at several stages of strain in the fault zone. Consequently, a wide spectrum of textures and structures are exposed in some fault zones, from strongly foliated, layered, and boudinaged gneisses to weakly layered and foliated tabular bodies, to undeformed dikes. We argue that at least some of these bodies were deposited from hydrothermal fluids rather than crystallization from silicate liquids. This opens the tantalizing question of what the compositions and pressures of aqueous fluids were during ductile faulting in the lower crust.

### **TECTONIC SIGNIFICANCE OF METAMORPHIC HISTORY**

#### **Putnam-Nashoba terrane, and History of Motion of the Honey Hill Fault System**

Several lines of evidence reflect important motion on the Honey Hill fault system. The regional truncation and folding of structurally higher rocks (e.g. between Stops 1 and 4) below the plane of the Honey Hill fault reflect thrust motion of the hanging wall. The gentle east dip of the boundary between reset and unreset schist, and the decrease in hornblende ages to the west reflect the deeper level and higher grade of Alleghanian metamorphism in the west (Stops 16-19). Rocks of the Putnam-Nashoba terrane are also progressively cut out to the west. In the east, the Lake Char fault separates the Avalon terrane from the basal, mafic unit (Quinebaug formation) of the Putnam-Nashoba terrane. In the Willimantic window, the Willimantic fault places the upper, pelitic unit (Tatnic Hill formation) directly on Avalon terrane rocks (Fig. 7). Thus the Avalon terrane boundary apparently cuts out the lower portion of the Putnam-Nashoba terrane. Farther west (Stop 19; Fig. 8) the entire Putnam-Nashoba terrane is missing.

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The cooling history of the rocks from the Putnam-Nashoba terrane contrasts strongly with that of the Avalon terrane. Peak metamorphic conditions of about 700°C and 6 kb (Hudson, 1982) in these rocks occurred in the Silurian or older, while in the underlying Avalonian rocks they were reached in the late Paleozoic (Fig. 4). The 80-million-year difference between the hornblende cooling ages (Fig. 5) thus precludes the possibility of thermal equilibrium between Putnam-Nashoba and Avalon terrane rocks during the Alleghanian heating of the latter (Wintsch *et al.*, 1992). On the contrary, it requires that Avalonian rocks were undergoing high-grade metamorphism, while Putnam-Nashoba rocks were at lower greenschist facies conditions. Given normal geothermal gradients, this approximately 300°C temperature difference requires the presence of 8-10 km of rock between the Putnam-Nashoba and Avalonian rocks at the Late Pennsylvanian time of peak metamorphism. This thickness of rock must have been removed in the Permian, because by Triassic time Avalon and cover rocks were adjacent, and cooling together (Wintsch *et al.*, 1992). The truncation of isograds at the western edge of the Avalon zone by the Lake Char and Bloody Bluff faults (Fig. 2) further indicates important motion on this fault zone that post-dates peak metamorphic conditions in both the hanging wall and foot wall rocks. The data require that: (1) Alleghanian metamorphism in the Avalon composite terrane occurred elsewhere, remote from rocks now exposed in the Putnam-Nashoba zone, and (2) peak metamorphism in the Avalon terrane predates significant late movement on the Honey Hill fault system (contrary to O'Hara, 1986).

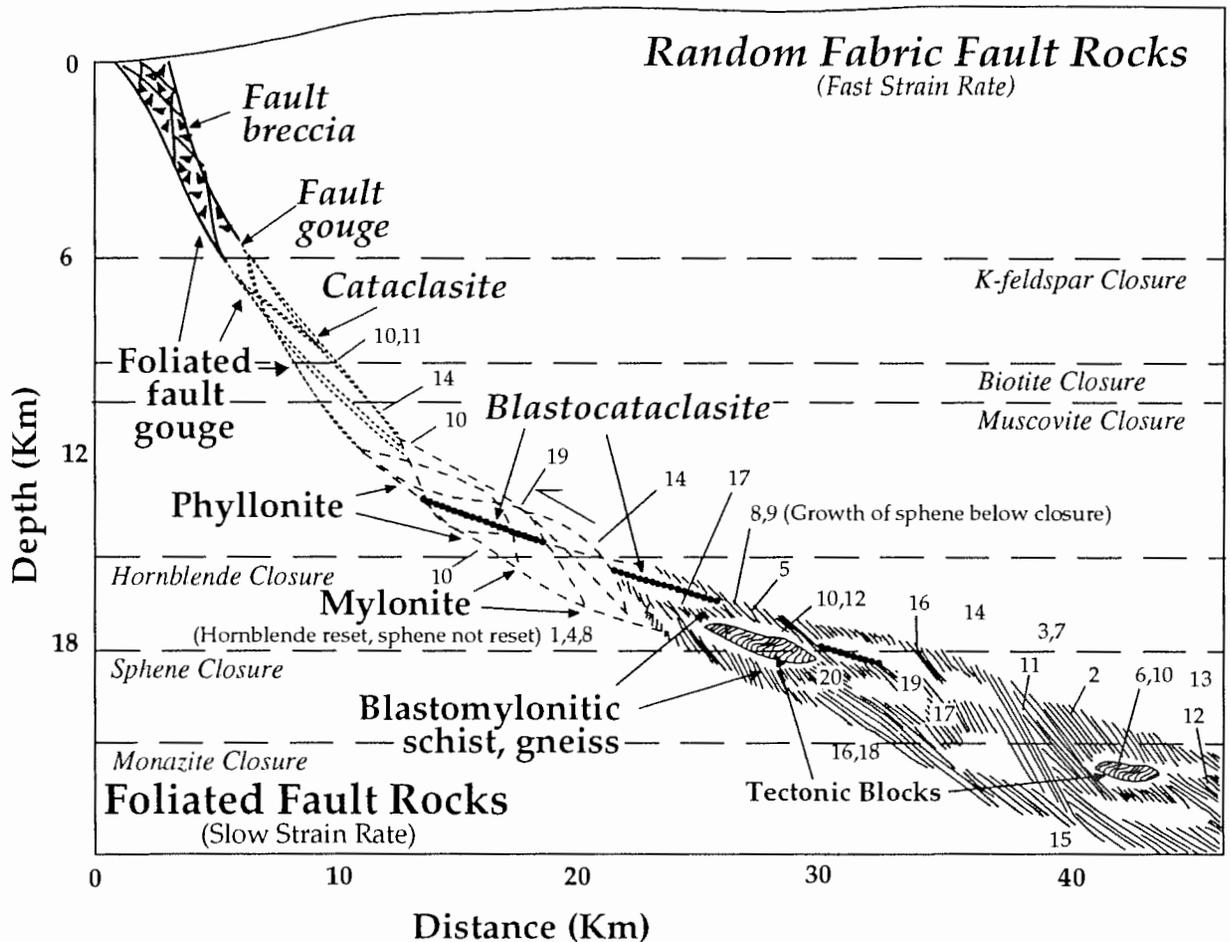


Figure 6. Conceptual model of a fault zone, showing possible P-T conditions of formation of various fault rocks: foliated fault rocks are in bold type; random fabric fault rocks are shown in italics. Numbers refer to field trip stops. The reactivation of ductile fault rocks under lower P-T and even brittle (Stops 10, 11, 14) conditions indicates that exhumation of the fault block was partially tectonic, and that the position of the hanging wall (or footwall) relative to the active fault did not change during exhumation.

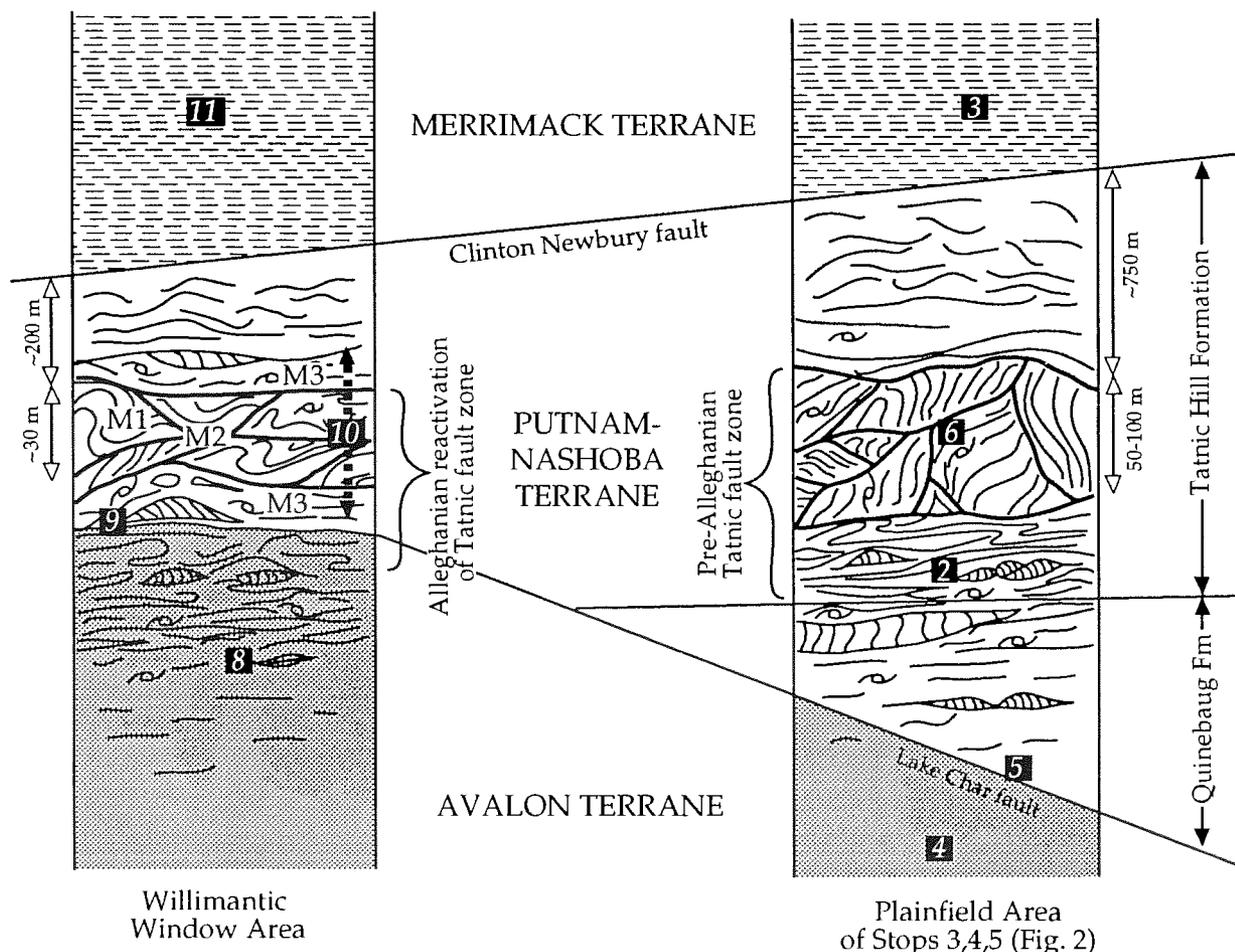


Figure 7. Partially schematic summary of the ductile structures in the Putnam-Nashoba terrane (after Wintsch, 1975, ms.), showing the relative position of pre-Alleghanian tectonic blocks in the Putnam belt (Fig. 2) with the mafic Quinebaug Formation below. In the Willimantic window the Quinebaug Formation is cut out, and the zone of tectonic blocks is reactivated in the Alleghanian. All Putnam-Nashoba rocks were metamorphosed in pre-Devonian times (Dixon, 1982), but in the Willimantic window this early metamorphism (M1) has been overprinted by Alleghanian high pressure staurolite + kyanite grade (M2) and lower pressure sillimanite-(andalusite) (M3) metamorphisms (Moecher and Wintsch, 1991; in review). Numbers in sections refer to relative position of field trip stops.

The same argument can be made for the Willimantic window. Avalonian core rocks have exclusively an Alleghanian metamorphic history (Wintsch *et al.*, 1992) where Late Proterozoic and Pennsylvanian sphenes are not fully reset (Getty and Gromet, 1992b), showing that temperatures did not exceed  $\sim 600^\circ\text{C}$  (Stop 8). In the overlying Putnam-Nashoba rocks, hornblende ( $\sim 280$  Ma) is reset (Wintsch *et al.*, 1992) but sphene and monazite (Getty and Gromet, 1992b) still record cooling from a pre-Acadian metamorphic event. The resulting event is well preserved in local syntectonic metamorphic mineral assemblages and fabrics (Stop 10), recording prograde Alleghanian P-T conditions of nearly  $600^\circ\text{C}$  and from 8 to 5 kb (Moecher and Wintsch, in review), superimposed on the earlier event (Stop 10). Only in the actual terrane boundary fault (Stop 9) are the Putnam-Nashoba rocks so fully reconstituted.

#### Merrimack terrane and History of Motion on the Clinton-Newbury Fault

The Clinton-Newbury fault separates the Putnam-Nashoba and Merrimack terranes, but its location in Connecticut is not known in detail. Its position on Figure 2 is put at the base of the lowest calc-silicate-bearing-Merrimack-like rock (locally named the Fly Pond member of the Tatic Hill formation, previously assigned to the Putnam-

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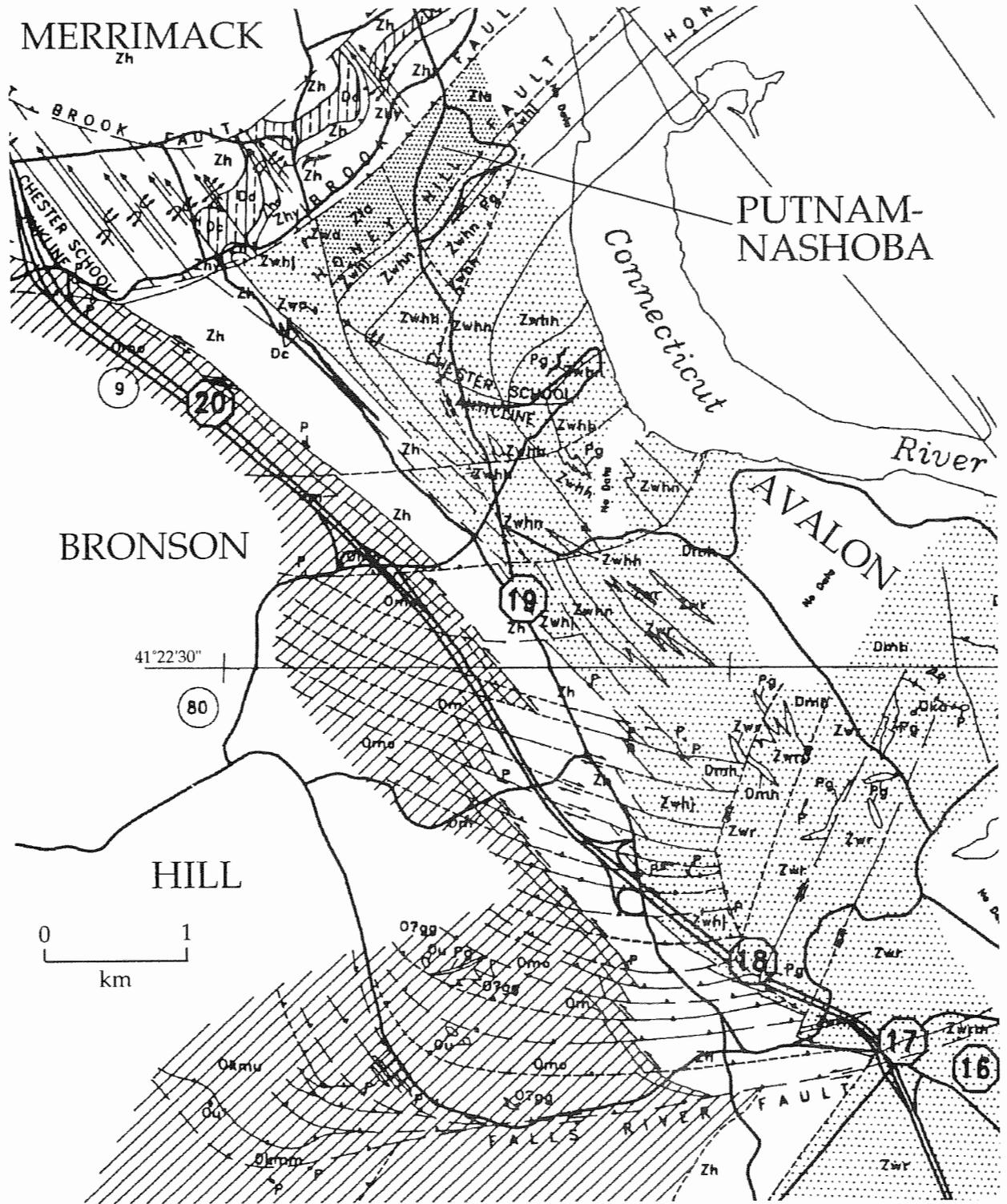


Figure 8. Detail of the geology of the Deep River area (after Wintsch, 1985), showing the distribution of terranes, major and minor faults, and the Chester School anticline that overturns all rocks southwest of its trace. The locations of Stops 16-20 are indicated.

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Nashoba terrane). Rocks in this zone are blastomylonitic suggesting high strain, and this position allows correlation of the Pennsylvanian cooling ages of hornblende samples just above this boundary (307 and 316 Ma) between Stops 6 and 7 with similar Merrimack and terrane cooling ages and contrast them with cooling ages of Putnam-Nashoba rocks. Thus these hornblende cooling ages contribute to defining the terrane boundary.

The geometry and metamorphic history of these rocks suggest net thrust displacement along this boundary. Rocks of the Putnam-Nashoba terrane (north of Stop 3) in the footwall of the Clinton-Newbury fault are locally cut out, as are some Merrimack rocks in the hanging wall in the west, where the Honey Hill and Clinton-Newbury faults touch (Stop 19, Fig. 8; see Fig. 3). Hornblende and sphene ages reflect Pennsylvanian cooling, but are younger than the same minerals in the underlying Putnam-Nashoba rocks. Because rocks structurally overlying and in thermal equilibrium with Putnam-Nashoba rocks would have cooled *before* the rocks below them, hornblende cooling ages from Merrimack rocks would be expected to predate hornblende cooling ages from Putnam-Nashoba rocks by approximately 5-10 million years. Merrimack rocks could not have cooled after the rocks of the Putnam-Nashoba zone if they had been overlying them during the cooling of both. This is another line of evidence for thrust motion in the Alleghanian prograde metamorphism of Putnam-Nashoba rocks in the Willimantic window. In the window Putnam-Nashoba rocks show Alleghanian loading of ~ 3 kb and have hornblendes that are reset. Thus, the rocks of the Merrimack terrane must be allochthonous, and must have overridden the Putnam-Nashoba rocks since ~310 Ma, probably in the early Permian.

#### **Central Maine Terrane and the Black Pond Fault**

Rocks of the Central Maine terrane are separated from those of the Merrimack terrane by the Black Pond Fault. The upper part of the Merrimack terrane is cut out and map scale folds are present in the foot wall (north of Stop 3). Rocks at the base of the Central Maine terrane in the hanging wall are also strongly cut out. Motion sense here is readily argued to be thrusting.

Parallel to the arguments above, the thermal disequilibrium between rocks of the Central Maine terrane and the underlying Merrimack terrane suggested by the higher grade and younger age of hornblendes in the former require a net thrust motion. Moreover, Merrimack hornblendes in the west, deeper in the orogen, have been reset, and yield Permian cooling ages, which also require Alleghanian loading, presumably from rocks in the hanging wall. Work on these rocks is in progress.

#### **Bronson Hill terrane and the Bonemill Brook fault**

The Bronson Hill terrane is separated from rocks to the east everywhere by the ductile Bonemill Brook fault system. Rocks and faults at the top of the Central Maine terrane are cut out by the Bonemill Brook fault (NW of stop 13). The rocks in the Bronson Hill terrane hanging wall are apparently less deformed. However, there is an important drop in metamorphic grade in rocks of the Bronson Hill terrane relative to rocks to the east. North and west of the Willimantic window the boundary dips west, where rocks of the Bronson Hill terrane cooled to hornblende closure temperature 10-15 million years later than the rocks of the Central Maine terrane to the east. Here, again, reverse motion on the Bonemill Brook fault system is implied. Stronger evidence for thrust motion is the presence of the small Hopyard Klippe of Bronson Hill rocks south of the Willimantic window (Fig. 2). In the southwest (Fig. 8) the Bonemill Brook fault cuts out all of the Central Maine terrane, and much of the top of the Merrimack terrane. Here (Stop 20) the last motion was strike-slip. Work on this boundary is in progress.

#### **Assembly of Lithotectonic Terranes in Southeastern New England**

In summary, there is a progressive decrease in metamorphic age in terranes from east to west, from structurally lower to higher level rocks. Net thrust motion is required by the presence of metamorphic rocks with younger cooling ages overlying older, and in some cases by higher grade metamorphism overlying lower grade. A sequence of hinterland propagating thrust nappes is required by these results. This stack of nappes was thicker in the west, as shown by the reset ages and younger metamorphism in the west, and occurred at a time equivalent to the oldest reset ages, or ~280 Ma. By late Permian, however, rocks of all terranes were cooling (Fig. 5). The cooling reflects exhumation, in part tectonic exhumation, with west dipping faults and terrane boundaries reactivated in a normal sense (i.e. hanging wall moves NW). This superposition of NW normal motion upon SE verging structures has led to spirited debate over the significance of kinematic indicators that reflect the various stages of motion. Our results show that the older loading history must have been achieved by thrusting. Kinematic indicators suggesting normal

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motion found in the younger Permo-Triassic (Goldstein, 1989; Getty and Gromet, 1992a) retrograde fault zones reflect the overprinting and destruction of earlier, ductile fabrics. It does not, however, provide evidence for a lack of loading during Alleghanian orogenesis in southeast New England; rather, it provides evidence that post-Alleghanian exhumation denudation was in part tectonic.

### CONCLUSIONS

The distribution of lithotectonic terranes in southeastern New England reflects post-Acadian, Alleghanian, uplift and stacking in a westward (hinterland) propagating sequence of thrust nappes, followed by reactivation of these boundaries during NW verging tectonic exhumation by normal motion. Thus all terrane boundaries were active as thrust faults with regionally significant motion, but many were reactivated as normal faults, along which ductile fault fabrics and mineral assemblages were overprinted by more brittle fabrics and lower grade mineral assemblages.

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## ROAD LOG

**October 22, 1993: Day 1.** The purpose of today's trip is to establish the lithologic sequence of rocks in the Avalon, Putnam-Nashoba and Merrimack terranes, and to document the discontinuities in metamorphic grade and time of metamorphism across the terrane boundaries.

**Mileage**

Assemble at Conn. Yankee Inn at the intersection of I95 and SR 161, Niantic, Conn.

- 0.0 Northbound entrance ramp to I95, interchange 74.
- 1.0 Follow Exit 76 to I395
- 12.0 Crossing the Honey Hill Fault separating the Avalon terrane from the structurally higher Putnam-Nashoba and Merrimack terranes.
- 22.0 Crossing the trace of the Tatnic fault, internal to the Putnam terrane.
- 47.0 Follow exit 97 (passing mylonitic Putnam rocks) to US44. Turn R (E) on US44
- 50.5 Approximate trace of the Lake Char fault, separating Avalon and Putnam-Nashoba rocks.
- 51.6 Rhode Island state line
- 53.7 Low road cuts on the left (N) side of US Rt. 44 between Bowdish Reservoir and Lake Washington; 350 m east of small bridge over Bowdish Reservoir, Rhode Island. Carefully make U-turn on north side of road.

**STOP 1. AVALON TERRANE (30 MINUTES)** Thompson Quad (Dixon, 1974; Wintsch, 1992). The purpose of this stop is to show a typical Avalonian granodioritic orthogneiss, and to show how much history can be extracted from it in spite of relatively poor exposure, and lack of constraining metamorphic mineral assemblages. The rock is a grayish-tan, massive weathering well foliated, plagioclase, quartz, microcline, biotite, epidote, hornblende, sphene gneiss, probably orthogneiss: the Late Proterozoic Ponaganset Gneiss (Dixon, 1974). Foliation and a conspicuous lineation (30°N15E) is defined by the parallel alignment of disseminated biotite flakes, but biotite-rich folia also give the rock a locally banded appearance. In contrast to K-feldspar porphyroclasts up to 1 cm in diameter, equant to elongate matrix grains of K-feldspar with weak cross-hatch twinning are 200 to 500 μm long. Hornblende (100 x 200 μm) is only ~2% of the rock, occurring as anhedral equant to elongate grains, always surrounded by, and locally embayed by biotite and epidote. These textures suggest the retrograde replacement reaction: hornblende + K-feldspar + Quartz + H<sub>2</sub>O = biotite + epidote.

This outcrop contains a strong record of cooling from Alleghanian metamorphism. Hornblende from this outcrop plateaus at 263 Ma, biotite total fusion age is 249 Ma, and K-feldspar gives early Triassic ages (Table 1). However, sphene from this rock gives a concordant age of 602 Ma. Thus the temperature of Alleghanian metamorphism in this rock exceeded the closure temperature of hornblende, but not of sphene, and must have been ~550°C. In fact the age of 602 Ma is very easily interpreted to be a cooling age following late Paleozoic intrusion and cooling. This rock, then, was heated to about first sillimanite isograd conditions in the Permian, but last experienced second sillimanite grade temperatures in the late Precambrian. Given this cooling history, and a temperature estimate for hornblende replacement (above) of ~450°C, the lineation is constrained to be latest Permian, and a conspicuous flat quartz vein crystallizing in the lowest greenschist facies must be early Triassic.

- 53.7 Follow US 44 west from STOP 1.
- 56.4 Turn R (N) on SR21 at Ma Frosty's.
- 58.2 Turn R (N) on SR 193.
- 58.6 Jct. of SR 193 and 200 in the village of Thompson, and the approximate trace of the Tatnic fault. Continue NW on SR 200.
- 60.4 Intersection of Ct. Rts. 200 and 12. Turn left (S) on Ct. 12.
- 61.4 Bear left, leaving Ct. 12 (to the right), and enter the ramp to I395.
- 61.5 Park off the right shoulder of the ramp, before the RR overpass. Walk 100m up the ramp to Stop 2.

**STOP 2. PUTNAM-NASHOBA TERRANE (45 MINUTES)** Putnam Quad (Dixon, 1982; Wintsch, 1992). The purpose of this stop is to show several characteristic rock types and structures of the upper pelitic unit of the Putnam-Nashoba terrane, here close to the base of the Tatnic Hill formation near the Tatnic fault. Foliations dip

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gently west, and steep ductile shear zones begin to define tectonic blocks (Fig. S2.1). The rocks of this zone are significant because they are clearly very high grade, above the second sillimanite isograd of the upper amphibolite facies. Garnets from this and nearby outcrops show only retrograde zoning (Fig. S2.2), and garnet-biotite geothermometry suggests temperatures of at least 650°C (Hudson, 1982). These rocks are higher in grade than rocks in the footwall Avalon terrane. The limited data available from the hanging wall Merrimack terrane (F, Y; Fig. S2.2) suggests lower grade, and even prograde zoning, again suggesting lower grade metamorphic conditions in these rocks.

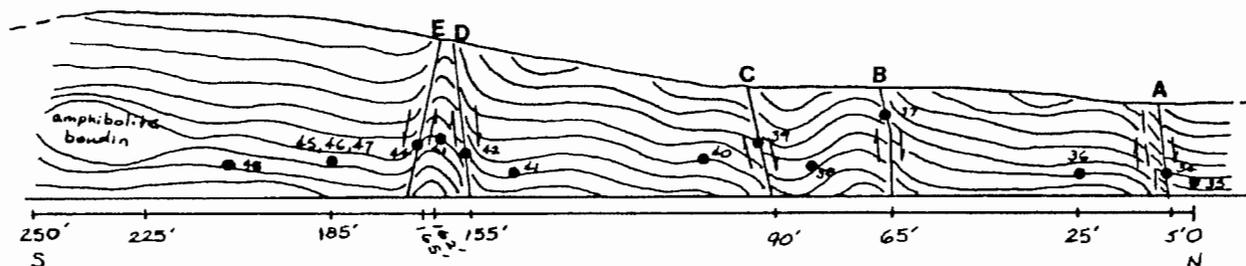
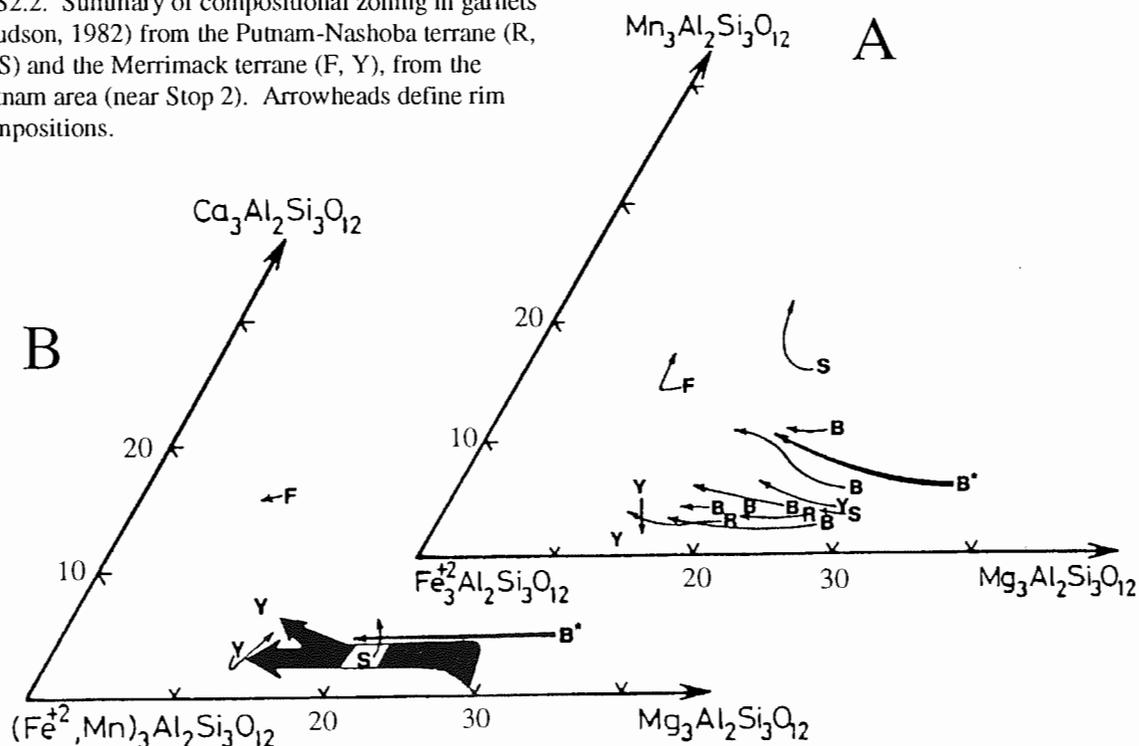


Figure S2.1. Field sketch (from Hudson, 1982) of the western-most exposure at Stop 2, showing pre-Alleghanian foliation (strike parallel) categorized by dominantly north dipping, normal faults.

Part of the argument for a net thrust motion on the Lake Char fault comes from the higher grade Putnam-Nashoba rocks occurring over lower grade Avalonian rocks across this fault. In sharp contrast to Stop 1, amphiboles and sphenes from nearby rocks yield late Devonian cooling ages (Wintsch *et al.*, 1992; Wintsch, 1992), and indicate that peak metamorphism and most of the deformation in these rocks is Devonian or older.

Figure S2.2. Summary of compositional zoning in garnets (Hudson, 1982) from the Putnam-Nashoba terrane (R, B, S) and the Merrimack terrane (F, Y), from the Putnam area (near Stop 2). Arrowheads define rim compositions.



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- 61.5 Return to bus. Proceed south on I395.
- 63.3 Exit at interchange 97, following US 44 W through Putnam.
- 64.6 US44 joins SR12. Follow US 44.
- 66.6 Pomfret town line.
- 73.2 Jct. SR97. Turn L (S) on SR97 in Albington.
- 79.9 Village of Hampton
- 79.9 Jct. US 6. Turn L (E) on US 6.
- 80.1 Road cuts on N side of US 6, Freshly blasted in May, 1993.

**STOP 3. MERRIMACK TERRANE (30 MINUTES)** Hampton Quad (Dixon and Pessl, 1966; Dixon *et al.*, 1968; Wintsch, 1992). The purpose of this stop is to show the ductile deformation of the well layered biotite schist and calc-silicate-granofels characteristic of the Merrimack terrane. This rock is dark greenish gray and purplish gray, medium grained, layered (mm-cm scale) hornblende granofels and biotite schist. The granofels layers 1/2-2 cm thick, contain 100-300 mm grains of essential scapolite (porphyroblastic), hornblende, plagioclase and quartz, all in random orientation, with accessory zoisite, clinozoisite, calcite, sphene and zircon. Subhedral porphyroblastic hornblende grains are between 100-1000 mm long, and 50-300 mm wide. They very commonly contain rounded inclusions of scapolite, quartz, and plagioclase, but show straight extinction, and uniform pale green birefringence. The biotite schist layers 0.5-1.5 cm thick contain essential quartz, plagioclase, biotite and scapolite, with accessory zoisite, hornblende, sphene and zircon. All grains are anhedral except biotite, which is subhedral and evenly disseminated; it does not define biotite-rich folia. Several isoclinal Z-folds are typically overturned to the east (Dixon *et al.*, 1968), and here plunge  $\sim 15^\circ$  N60W. They probably reflect eastward vergence of thrust nappes over and under these rocks.

The age spectrum from hornblende in this outcrop plateaus at  $310 \pm 2$  Ma, and sphene yields an age of  $315 \pm 5$ . In contrast, muscovite and biotite are Permian. The Pennsylvanian age of this sample is distinctly different from hornblende ages in both the Avalon and Putnam-Nashoba terranes of Stops 1 and 2. In fact the *younger* age of this hornblende requires that it cooled about 30 m.y. after the hornblende in the structurally lower Putnam-Nashoba terrane rocks. Normal cooling in a stack of rocks in thermal equilibrium would require that structurally higher rocks cool before those below. Thus the younger age of these hornblendes and sphenes requires that Merrimack rocks cooled to below about  $500^\circ\text{C}$  elsewhere (to the west?), and have been thrust over Putnam-Nashoba rocks since amphibolite facies conditions, or in post-Pennsylvanian times.

- 80.1 Continue E on US6.
- 88.3 Road work in June 1992 destroyed out crops of Putnam Nashoba terrane rocks across from McDonalds.
- 94.5 Drive just into Rhode Island, turn around, and proceed back west on I395 access road.
- 97.8 Road cuts on the right (N) side of the road just behind the Exit 90 road sign.

**STOP 4. AVALON TERRANE (30 MINUTES)** East Killingly Quad. (Moore, 1983; Goldstein and Owens, 1985; Wintsch, 1992). The purpose of this stop is to examine metasedimentary and metavolcanic rocks metamorphosed to only lower amphibolite facies near the top of the Avalon terrane. The most abundant rock here is a gray to tan, massive to locally slabby weathering, well foliated quartz, muscovite, biotite quartzite of the Late Proterozoic Plainfield Formation. At the northeastern limit of the road cut, a massive, dark olive green weathering, moderately foliated hornblende, chalcopyrite, quartz, plagioclase, sphene, epidote amphibolite. Hornblende needles range in size from 30 x 30 to 150 x 350 mm. The grains are untwinned, and fewer than 10% of the grains contain small (<5 mm) inclusions.

Muscovite from the quartzite here yields a plateau age of  $250 \pm 1$  Ma, and biotite gives a total gas age of  $244 \pm 1$  Ma. These are both consistent with the regional Permian cooling throughout the Avalon zone. However, the acicular hornblende at the east end of this outcrop produces on an isotope correlation diagram, a strong linear array containing 98.3% of the gas, with a correlation age of  $277 \pm 8$  Ma, and a  $^{40}\text{Ar}/^{39}\text{Ar}$  of  $850 \pm 103$  (Wintsch, 1992). We interpret this locality, at a structurally high level in the Avalon terrane, to follow an isopleth on Fig. 4 of about 12 km, while deeper rocks (Stop 1) would have followed a deeper (e.g. 20 km) isopleth. The further implication of this Permian age of amphibole, is that the rocks of the Hope Valley subzone did not experience a metamorphism of greater than middle amphibolite grade prior to the Permian.

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- 97.8 Continue west on I395 connector.  
 104.9 Leave highway at exit 88. Bottom of exit ramp is Stop 5.

**STOP 5. PUTNAM-NASHOBA TERRANE (30 MINUTES)** Plainfield Quad (Dixon, 1965; Lundgren, 1968; Wintsch, 1992). The purpose of this stop is to examine mylonitic rocks close to the Honey Hill fault system, but derived in part from the amphibolite-rich Quinebaug Formation of the Putnam-Nashoba terrane. Most of the deformation in the Willimantic and Tatnic faults occurred under upper amphibolite facies conditions, whereas many fault rocks along the Lake Char fault developed under lower amphibolite and greenschist facies conditions. Here strongly layered, well foliated and folded mylonites and blastomylonites are well exposed. The common FeO/TiO<sub>2</sub>/MnO ratios (Wintsch, unpublished data) of these mylonites with an andesite porphyry in Dixon's (1965) qlc unit in the Quinebaug Formation support her suggestion that these mylonites were derived from this porphyry.

A complicated relationship between the granodiorite porphyry and intruded mylonites is clear here. Mylonite and blastomylonite is intruded by dikes of granodioritic porphyry, while continued deformation of the mylonite zone folded and mylonitized the porphyry. An interesting and complicated set of crosscutting relationships results, especially on the south side of SR14A. On the N side of SR14A, strongly foliated and layered rocks derived from this porphyry are interlayered with mylonitic rocks derived from amphibolites of the Quinebaug Formation (Putnam-Nashoba terrane). These mylonites may be associated with the Lake Char fault, or they may be internal to the Putnam-Nashoba terrane. However, the source of the granodiorite porphyry cannot be found structurally below this outcrop, and so is probably cut out by the Lake Char fault. Thus these mylonitic rocks are probably within the Putnam-Nashoba terrane.

- 104.9 Return to bus. Return to I 395 south.  
 113.2 Crossing the Tatnic fault, internal to the Putnam-Nashoba terrane.  
 116.2 Leave highway at exit 83. Turn R (S) on SR97, drive under highway 97, drive under I395, to north bound entrance ramp and STOP 6, 5.5 km northeast of Norwich.

**STOP 6. PUTNAM-NASHOBA TERRANE (40 MINUTES)** Norwich Quad. (Snyder, 1961; Wintsch, 1987). The purpose of this stop is to show assemblages and structures typical of the quartz-plagioclase-biotite schists and gneisses in the hanging wall of the Tatnic fault (Putnam-Nashoba terrane, Fig. 2) and to show the boudin-tectonic block structures characteristic of this fault. Most of the schists and gneisses of these rocks contain only quartz-plagioclase-biotite, sphene, K-feldspar, with accessory garnet and sillimanite, and the assemblage quartz-K-feldspar-garnet-biotite-sillimanite, typical of the upper amphibolite facies are present locally. Abundant pegmatitic and migmatitic structures, and the lack of prograde muscovite further reflect these high grade conditions. A muscovite age of ~275 Ma (Wintsch *et al.*, 1992) from this outcrop is evidence against a greater than greenschist facies Alleghanian overprint in these rocks.

Large boudins, some rotated, are the most important structure in these rocks. The shear zones separating the blocks generally dip steeply west, and biotite-rich assemblages lacking muscovite and chlorite suggest that they developed under amphibolite facies conditions (see Fig. 7), and must therefore be pre-Alleghanian structures.

- 116.2 Return to I395, S-bound  
 120.8 Exit 82 to SR2, 32, Norwichtown  
 121.6 Follow SR32 North, through village of Yantic  
 122.6 Follow SR87 NW (L)  
 124.1 Road cuts under power line on SR87. STOP 7.

**STOP 7. MERRIMACK TERRANE (30 MINUTES)** Fitchville Quad. (Snyder, 1964a). This road cut exposes the contact between the Oakdale (Hebron) Formation and the Scotland schist (muscovite-schist) member. The Oakdale of this outcrop is a thinly layered quartz-plagioclase-biotite schist and quartz-plagioclase-hornblende ± scapolite granofels. Layering occurs on the scale of 1-10 mm. Hornblende rich layers 1-5 mm thick are locally boudinaged. The very gently dipping contact between the Oakdale and the Scotland schist member, a quartz-muscovite-biotite-plagioclase-garnet-staurolite schist is parallel to compositional layering within the Oakdale suggesting that all compositional layering in the Oakdale is inherited from differences established at deposition. Hornblende and

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muscovite from these rocks produce ages of 306 (isochron) and 251 (Plateau). Although sphene (Stop 3) is reset, the presence of staurolite + kyanite in these rocks shows that the grade of metamorphism here is less than that in the Putnam-Nashoba terrane. The younger age of hornblende (~310 Ma) in these rocks shows that Merrimack terrane rocks were still under middle amphibolite facies conditions 30 m.y. later than the Putnam-Nashoba rocks below.

- 124.1 Return to bus. Follow SR87 back to SR32.
- 127.4 SR 2. Follow it S to I395, to I95 Exit 74,
- 143.7 Connecticut Yankee Inn.

*October 23, 1993: Day 2. The purpose of today's trip is to document Alleghanian overprinting in the Willimantic window, and to contrast the grade of metamorphism of rocks of the Merrimack, Central Maine and Bronson Hill terranes in and NW of the Willimantic window with the rocks of Day 1.*

- 0.0 Leave Connecticut Yankee Inn
- 15.4 Exit 81 to SR 2, west
- 16.6 Exit SR 32
- 28.2 Bear R (N) at flashing light, to downtown Willimantic.
- 28.3 Turn L (W) on SR66, rejoin Rt. 32, follow 32
- 30.6 North to road cuts on entrance ramp to Rt. 6 at SR32. STOP 8.

**STOP 8. AVALON ZONE, WILLIMANTIC WINDOW (45 MINUTES)** Willimantic Quad. (Snyder, 1964b; Wintsch and Fout, 1982; Wintsch, 1992). The purpose of this stop is to show the metavolcanic rocks at the structural top of the Avalon terrane, to document middle amphibolite facies metamorphism and ductile deformation, and to show that feldspar-rich sill- and dike- like structures can not be magmatic. These rocks contain Late Proterozoic metavolcanic Mansfield Hollow Lithofacies of the Hadlyme Formation (Wintsch *et al.*, 1990), and are correlated to part of the Waterford complex in the New London area. The most abundant rock type is a pale gray, massive weathering, well foliated and layered plagioclase, quartz, biotite K-feldspar, sphene, (muscovite or hornblende) gneiss and granofels. A second conspicuous rock type is a dark gray to black, massive, unlayered, well foliated hornblende, plagioclase, biotite, magnetite, sphene amphibolite. The outcrop is cut by several 0.5-1.0 m thick zoned granitic pegmatites, and by felsic sill-like layers parallel to the layering in the gneiss. The pegmatites are cored by quartz, have K-feldspar intermediate zones, and muscovite bearing margins. The sill-like structures are dominated by and are thickest where microcline-rich porphyroblasts bow out the surrounding foliation. They are joined together in necklace fashion by thinner layers of quartz and plagioclase.

Hornblende from the amphibolite produces an isochron age of 281 Ma on an isotope correlation diagram (Wintsch 1992). Muscovite from a cross-cutting pegmatite, and biotite, and K-feldspar from the plagioclase gneiss yield cooling ages of 247, 243, and 228 Ma respectively. The cooling curve produced by these data (Fig. 4) is very similar to that for the Hope Valley zone, except for a 15 m.y. older hornblende. Incompletely reset Late Proterozoic and Pennsylvanian sphene (Getty and Gromet, 1992b) in these rocks shows that the rocks did not reach the upper amphibolite facies temperatures of sphene closure (550-600°C) in the Alleghanian.

Knowledge that these rocks never exceeded 600°C is critical to interpreting the pegmatites and feldspar-rich 'sills,' in that they could not have been produced by the crystallization of silicate liquids. Partial melting is not a possibility, because even the most volatile-rich liquids do not form below 600°C. Thus all of these structures must have crystallized from an H<sub>2</sub>O-rich liquid. The rocks also show moderate ductile deformation, but the necks between amphibolite boudins are rather straight, and are filled primarily with quartz. This is consistent with lower to middle amphibolite facies deformation, and with fracturing of the relatively strong amphibolite during ductile deformation of the relatively weak quartz-feldspar host gneiss. A locally strong lineation defined by biotite streaks and quartz and feldspar rods apparently formed at about 400°C (Wintsch and Fout, 1982), and by use of the T-t curve of Fig. 4, they formed during latest Permian.

- 30.6 Return to bus. Drive south on SR 32
- 31.4 Rejoin SR66, follow SR32 south.
- 32.2 Turn south at light, cross railroad tracks, and Willimantic River.

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- 32.4 Proceed straight through intersection onto SR289.  
 33.0 Park under power line that crosses the highway traverse uphill (W) across several road cuts on the east side of Hosmer Mountain to STOP 9.

**STOP 9. WILLIMANTIC FAULT (30 MINUTES)** Willimantic Quad. (Snyder, 1964b; Wintsch, 1981; 1992; Wintsch and Fout, 1982; Getty and Gromet, 1992b). The purpose of this stop is to show the profile across the terrane boundary (Willimantic fault) between Avalon terrane plagioclase gneisses and Putnam-Nashoba terrane pelitic schists (Fig. 7). The Avalonian plagioclase gneisses are similar to (but more mafic than) those at Stop 8, and to many south of the Honey Hill fault, but here are more highly strained. Here, an apparent strain gradient across the contact is suggested by an increase in the development of boudinage, of small scale folding, and of plagioclase and hornblende porphyroblasts, porphyroclasts, and tectonic inclusions, and a reduction in grain size of the matrix (Wintsch, 1979). The Tatnic Hill Formation here is a mylonite schist, and relative to the precursor rock has undergone a 10-50 X grain size reduction (Wintsch, 1979). Across the contact into the Putnam-Nashoba terrane, strain is very high, and the rocks contain totally reconstituted blastomylonites, as evident by kyanite and andalusite-bearing mineral assemblages, chemical composition, and mineral and isotopic composition (Wintsch, 1981; Moecher and Wintsch, 1991, in review; Getty and Gromet, 1992b).

Strain is penetrative below the zone of tectonic blocks, where strain becomes very heterogenous. The lowest several meters of Putnam-Nashoba terrane rocks contain both kyanite and andalusite as well as sillimanite. The first two aluminosilicates embay, and andalusite also includes biotite, and both are associated with magnetite, suggesting the oxidation reaction (Wintsch, 1981):  $\text{Biotite} = \text{Al}_2\text{SiO}_5 + \text{magnetite} + \text{quartz} + \text{ions}$ . Hematite is also present in these rocks as thin blades intergrown with biotite. As grain size is reduced in these rocks, there is a decrease in  $\text{SiO}_2$ ,  $\text{Na}_2\text{O}$  and  $\text{CaO}$  relative to  $\text{Al}_2\text{O}_3$  and an increase in the  $\text{Fe}_2\text{O}_3/\text{FeO}$  ratio. This correlation of oxidized assemblage with small grain size supports the proposal of Wintsch (1981) that a pH increase caused by surface exchange could have been responsible for this oxidation. Thus the mineralogy and even the composition of these rocks do not reflect the upper amphibolite facies conditions which the rocks once experienced. Rather, they reflect a complex set of metasomatic reactions which by some path were probably strain induced at conditions that crossed the stability fields of all three  $\text{Al}_2\text{SiO}_5$  polymorphs. The apparent discordance of this schist with the rocks both above and below suggests that some of the later strain in these rocks cut across both units. Further south along the road rocks higher in the structural section are exposed. They contain upper amphibolite facies assemblages and structures, including intrafolial folds, boudinage, tectonic blocks, and feldspathization — evidence of the earlier (Devonian or older), completely ductile deformation.

- 33.0 Return to bus, retrace path to Stop 8.  
 35.4 Enter SR6, westbound  
 36.8 Road cuts on SR6, 300 m northeast of its intersection with Ct. Rt. 66, 3 km west of Willimantic. STOP 10.

**STOP 10. PUTNAM-NASHOBA TERRANE, WILLIMANTIC WINDOW (60 MINUTES)** Columbia Quad. (Snyder, 1967; Wintsch and Fout, 1982; Wintsch, 1987; 1992; Getty and Gromet, 1992b). The purpose of this stop is to examine the pre-Devonian upper amphibolite facies rocks and structures of the Putnam-Nashoba terrane, and their overprint by prograde Alleghanian metamorphism. The most stunning features exposed are the large and very large tectonic blocks containing upper amphibolite facies metamorphic assemblages separated by anastomosing shear zones containing middle amphibolite facies assemblages (Fig. 7). The dome shape of these blocks leads to quaquaversal foliation patterns in natural exposures in the Willimantic and Columbia quadrangles (Snyder, 1964b; 1967) that is diagnostic of the basal Putnam-Nashoba rocks in the Willimantic window. Because they are usually 100 ft (30 m) or more long in the E-W direction, they are best viewed from a distance of 100 ft (30 m) or more. Face I (Fig. S10.1) is particularly well suited for viewing from a distance because the highway is not finished; but for those especially interested in these structures, a walk through all the cuts is imperative. The lack of continuity of the blocks or the shear zones on either side of the road along the north side of the interchange (at II, Fig. S10.1) indicates that these structures are not longer than 100 ft (30 m) in the N-S direction, and thus the blocks must be lens-shaped. Some of these blocks may have developed as large drag folds (e.g., 100m, Fig. S10.2) with axes striking N 30 E, evolving into these discrete blocks as strain was concentrated on the long limbs (Wintsch, 1979, Fig. 5; see also Soula and Bessiere, 1980). However, many blocks do not appear to be rotated, and some degree of mega-

boudinage (e.g. 200 m, Fig. S10.2) was probably also involved. Rotation of these blocks and many other small-scale structures were probably produced during southeast motion of the Merrimack and Putnam-Nashoba terranes over the Avalon terrane during the loading stage of the Alleghanian orogeny.

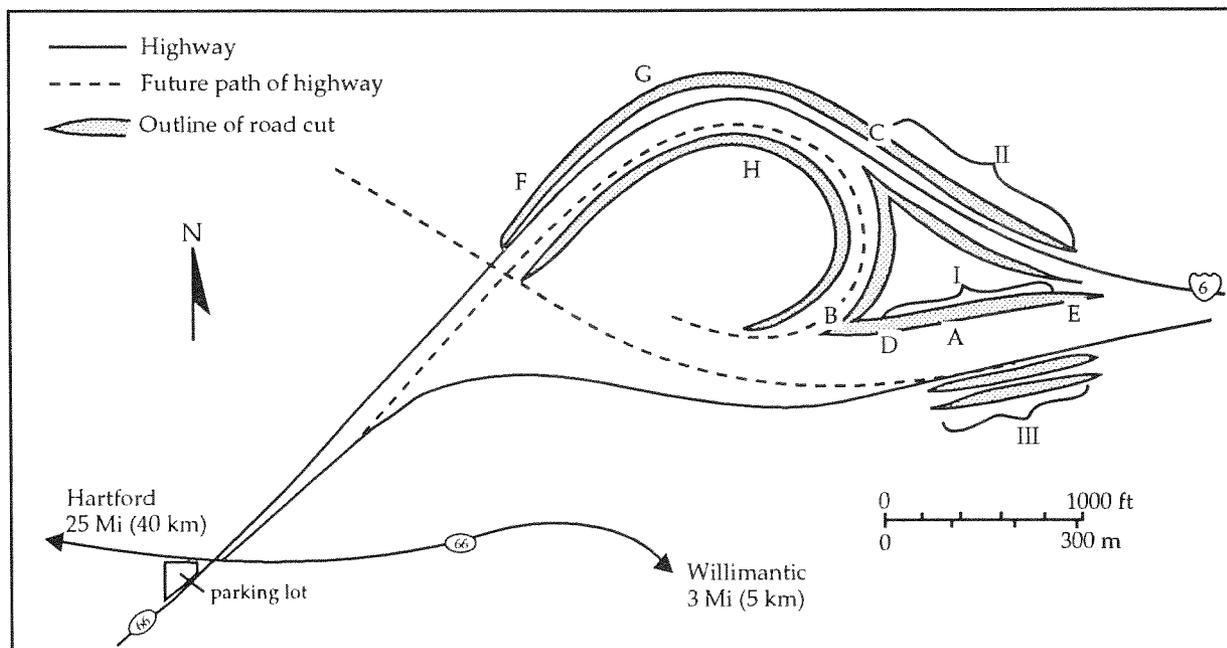


Figure S10.1. Map showing the distribution of outcrop (shaded), road cuts and natural exposures along the unfinished interchange of US6 and SR66. Letter symbols refer to locations given in text.

Augen gneiss, blastomylonitic gneiss, mylonitic schist, and mylonite are present in order of decreasing abundance in these rocks. The highest grade assemblages are best preserved in the augen gneisses inside and away from tectonic blocks (e.g., east end, 150 to 250 m, Fig. S10.2). These quartz-plagioclase-biotite-K-feldspar-garnet-sillimanite bearing gneisses locally contain K-feldspar and plagioclase augen up to 5.5 in (14 cm) in diameter. These augen probably grew as porphyroblasts, and did not crystallize from a melt. Evidence for this comes from rotated biotite inclusion trails in some crystals. The incorporation of these rotated inclusions demands growth of a rigid crystal in a solid matrix. Moreover, the bulk composition of the augen gneisses is strongly syenitic or monzonitic, and does not reflect the minimum melt composition of these pelitic gneisses, which should be close to a granitic eutectic. Sillimanite is ubiquitous in these gneisses, and commonly occurs as randomly oriented needles in fibrolite mats. This metamorphism (M1, Fig. 7) to pre-Devonian (Getty and Gromet, 1992b), and reached 700°C and 8kb (Moecher and Wintsch, in review).

The shear zones surrounding and cutting these blocks of high-grade gneiss contain blastomylonitic schists and gneisses with lower grade kyanite-bearing assemblages (e.g., localities A, B, C Fig. S10.1, if not collected out). This Alleghanian metamorphism (M2, Fig. 7) exceeded a pressure of 8kb (Moecher and Wintsch, in review). These high pressure shear zones are in turn cut by lower pressure (3-5 kb) shear zones, where sillimanite occurs with a strong preferred orientation trending approximately N 60° W, parallel to locally developed biotite streaks and quartz-feldspar rods. Still lower grade, slabby, strongly lineated and layered mylonitic schists and blastocataclastic rocks (Fig. 6) are well exposed on the natural cliff face III (Fig. S10.1), south of the road. This schist projects under all rocks exposed in road cuts, and its strong N-S trending lineation parallel to tight isoclinal folds suggests a similar change in fault motion direction. Another less slabby, even textured, and finer grained mylonitic schist forms a gently folded 3 to 4 m thick layer at D (Fig. S10.1) between 60 and 120 m; (Fig. S10.2). This later foliation is itself foliated into small isoclinal folds, but they are difficult to find because the rock almost totally lacks compositional layering. True mylonites are rare, but late, fine-grained, middle to lower greenschist facies mylonite up to 2 cm

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thick cuts the other gneisses in the steeply west-dipping shear zone which cuts the entire exposure at A (Fig. S10.1; 140 m, Fig. S10.2). A brittle fracture zone occurs at E (Fig. S10.1) where the K-feldspar-chlorite-epidote bearing assemblage reflects very shallow alteration of this zone. Together these shear zones and their associated mineral assemblages demonstrate repeated deformation in this fault zone during shallower and lower metamorphic grade conditions (Fig. 6).

Several other rock types are present in these cuts. Interlayered in the pelitic blastomylonitic schists and gneisses are at least 28 thin (30 cm) amphibolite layers, all boudinaged (between A and D, Fig. S10.1). Successive boudinage of a larger (1 m) amphibolite boudin at its tapering neck can be seen at 215 m (Fig. S10.2). Several 30 to 50 cm thick layers of diopside-bearing marble are present at F (Fig. S10.1). Sphe from one of these marbles yields a U+Pb age of 335 Ma. The margins of one of these contained hornblende porphyroblasts up to 10 cm in diameter (now apparently collected out). A 2 m diameter pod of ultramafic rock, now chlorite-talc schist is present at G (Fig. S10.1).

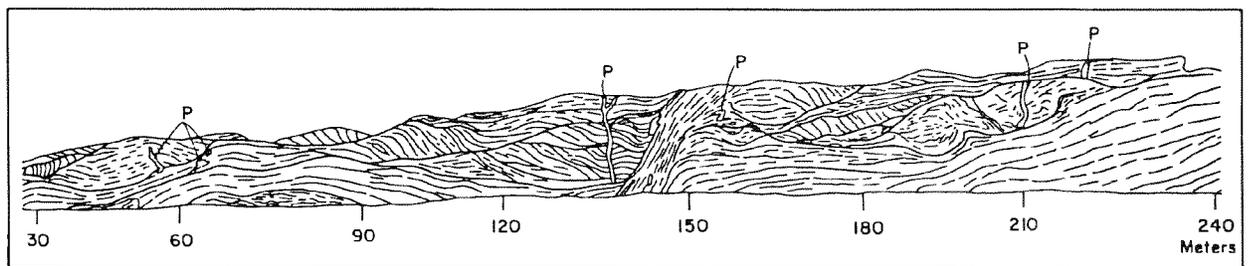


Figure S10.2. Profile traced from photographs of roadcut face I (Fig. S10.1) showing anastomosing shear zones that define tectonic blocks. The clockwise rotation of foliation from gently west-dipping (e.g. 210-240 m) to horizontal, and east dipping within tectonic blocks, provide some of the evidence for top moves east (to the right) or thrust motion (Wintsch, 1979). Local reactivation of west dipping shear zones in a normal sense occurred in the late Permian (Getty and Gromet, 1992a).

Hornblende from a massive weathering, well foliated hornblende, plagioclase, biotite, chlorite amphibolite, from the (unfinished) westbound entrance ramp (H, Fig. S10.1) was dated. Hornblende grains range from 0.5 to 3 mm in diameter, some with a few plagioclase inclusions. Most grains contain rectangular exsolution lamellae from 10-100  $\mu\text{m}$  long. A few larger grains are composed of smaller rounded, mutually embayed, randomly oriented grains, suggestive of recrystallization and grain growth. Other evidence for a secondary event includes the partial replacement of biotite grains 1/2 to 3 mm in diameter and by equally coarse grained chlorite, and the strong development of deformation-twinning in plagioclase grains, 200-500  $\mu\text{m}$  in diameter. This hornblende (Wintsch, 1992) produced a U-shaped age spectrum with an age minimum of 306 Ma. However, an isochron containing over 98% of the  $^{39}\text{Ar}_K$  in the sample defines an age intercept of  $\sim 273$  Ma, and is probably a better reflection of the time of cooling. This hornblende age is not consistent with monazite and sphene ages of 401 and 335 Ma respectively (Getty and Gromet 1992b), but is consistent with Alleghanian cooling in the Avalon zone. The monazite and sphene apparent ages are easily interpreted as a cooling age from pre-Silurian metamorphism, and define a cooling curve almost indistinguishable from that obtained from the rocks in the east (Fig. 3). Moecher and Wintsch (1991; in review) interpret the hornblende age of  $\sim 270$  Ma as reflecting cooling following Alleghanian reheating. This is consistent with the occurrence of overprinting metamorphic assemblages containing andalusite and kyanite in the ductile, Willimantic fault zone within 100 m above the Avalon zone contact (Wintsch, 1980). Indeed, complex exsolution lamellae (not present in hornblendes in the east) may reflect such heating. This Permian resetting may have been generated either by conduction of heat across the fault or by frictional heating (Barr and Dahlen, 1989). In view of all the other data that support isotopic resetting in this fault zone (Getty and Gromet, 1992b), this hornblende apparent age is probably best interpreted as locally thermally reset to a temperature  $> 500^\circ\text{C}$ , but less than  $\sim 600^\circ\text{C}$ , because sphene is not reset.

Wintsch (1979) interpreted the overall motion sense in these rocks to reflect SE directed transport. Since then Goldstein (1989) Getty and Gromet (1992a) have observed that some kinematic indicators suggest a NW normal

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sense of motion. The conflicting evidence for motion sense probably arises from the SE directed motion on thrusts that loaded and rotated the blocks, overprinted by nearly opposite, NW directed tectonic unloading during Permo-Triassic exhumation.

- 36.8 Return to bus, follow US6 W, make U-turn on US6, retrace path to SR32, following it north to US44a.
- 43.3 Turn left (west) on US44a.
- 44.6 Follow US44 west to STOP 11.

**STOP 11. MERRIMACK TERRANE (30 MINUTES)** **This is a very dangerous highway cut! Stay off the pavement! Spread out along the cut!** South Coventry Quad (Fahey and Pease, 1977). The purpose of visiting this outcrop are to examine rocks of the Merrimack terrane on the NW side of the Willimantic window, and the evidence for ductile and brittle deformation in the foot wall of the Black Pone fault. These 2 m high road cuts expose quartz-plagioclase-biotite, and hornblende-diopside schists and granofels. Foliation and axial planes of intrafolial folds dip gently N, and fold and boudin axes plunge gently N. At least three generations of concordant pegmatites cut these rocks (e.g. east end of outcrop). Older pegmatitic layers are strongly foliated and boudinaged, younger pegmatitic sheets are weakly foliated, and the youngest pegmatites are undeformed and tabular. All these structures are cut by planes of microbreccia and rare breccias, some showing a sinistral sense of displacement. Hornblende ages in these rocks are late Pennsylvanian, and probably postdate most of these ductile structures, making the latter Acadian or older. Regional data on K-feldspar ages (total gas) suggest that these breccias are Triassic.

- 44.6 Return to bus. Follow US44 east.
- 45.1 Turn left (north) on Brigham Road.
- 46.1 Follow Brigham Road north to natural cliff exposures on the west side of the road, STOP 12.

**STOP 12: CENTRAL MAINE-MERRIMACK TERRANE BOUNDARY BLASTOMYLONITES (30 MINUTES)** South Coventry Quad, (Fahey and Pease, 1977). The purpose of this stop is to show the very highly strained rocks of the Black Pond fault, the boundary between the Merrimack and Central Maine terranes. These exposures show very well these north-dipping mylonitic and blastomylonitic schists and gneisses of the Black Pond fault. Amphibolite layers are strongly thinned, but not boudinaged in the surrounding biotite-plagioclase-garnet bearing schist, probably reflecting very high temperatures and/or slow strain rates. These rocks also contain rootless intrafolial folds, boudin-like, tectonic inclusions, and steeply dipping mylonites that cut the mylonitic schist, and begin to define tectonic blocks similar to those in the Tatic fault zone (Fig. 7).

- 46.1 Return to bus. Follow Brigham Rod south.
- 47.1 US44. Turn left (east) on US44, cross SR32.
- 49.8 Turn left (north) into correctional facility.
- 50.0 Exit vans, at trail up the hill to water towers, and STOP 13.

**STOP 13: CENTRAL MAINE TERRANE (30 MINUTES)** (South Coventry Quad., Fahey and Pease, 1977). The purpose of this stop is to contrast the lithology, structure, and metamorphism of rocks assigned to the Central Maine terrane in the hanging wall of the Black Pond fault with those of the Merrimack terrane. The rocks around this water tower are strongly layered quartz-plagioclase-K-feldspar-garnet-biotite granular schists. The rocks contain conspicuous porphyroblasts of feldspars (1-3 cm) and garnet (1/2 - 3 cm). Steeply dipping NNE trending foliations wrap to more easterly trending strikes with shallower dips, mimicking the larger scale structure of the Black Pond fault.

- 50.0 Return to bus. Return to US44.
- 50.2 Turn right (west) on US44. Follow US44 to SR84.
- 58.8 Turn left (south) on SR85. Follow SR85 south.
- 65.4 Turn right (west) on SR94.
- 68.6 Turn left (south) onto Mimichang Farm to STOP 14.

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**STOP 14: GLASTONBURY GNEISS, BRONSON HILL TERRANE (45 MINUTES)** Marlborough Quad. (Snyder, 1969; Ambers, 1988). The purpose of this stop is to compare the structure, lithology, intrusive relationships, and metamorphism of rocks of the Bronson Hill terrane to those of the Central Maine terrane. The rocks in this exposure, studied by Ambers (1988), include a xenolith-rich strongly lineated granodioritic gneiss and mylonitic derivatives, an intrusive granitic gneiss, and cross cutting pegmatitic veins, all occurring in a sub-horizontal, right-lateral zone of deformation. Strain can be monitored using aspect ratios of xenoliths in the deformed granodiorite, that increase from a low of 12-15 in xenolith measurement domains 7-12 (Fig. S14-1) to over 135 in the mylonite domain 3. Regional studies show that hornblende + K-feldspar in the precursor assemblage has reacted to biotite-epidote-bearing assemblages. Contact metamorphism of this assemblage around granitic intrusions in the southern part of the outcrop remetamorphoses the biotite and epidote schist to a hornblende granofels. Thus, some of the ductile deformation of the granodiorite predates the granite, but it too, is deformed with the strong horizontal N30°E lineation of the gneiss. Many interesting small scale structures are present in the mylonite and around the pegmatites.

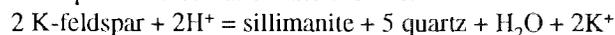
The mylonite in domain 3 is derived from the granodiorite by syntectonic metasomatism, where microcline originally in the mylonite was replaced by plagioclase and biotite, and plagioclase in the shoulders of the mylonite zone was replaced by microcline as porphyroblasts. Mass balance calculations among mylonite, quartz-K-feldspar veins, and metasomatized gneiss show nearly isochemical and isovolumetric deformation on the scale of the outcrop (Ambers, 1988).

*Day 3. The purpose of today's trip is to show the ductile thinning, extension and even omission of terranes along ductile faults in the Deep River area (Fig. 8) and to compare Alleghanian upper amphibolite facies metamorphism of the southern Bronson Hill and Avalon terranes in the 'core' (Lundgren's appendix?) of the orogen to the pre-Alleghanian metamorphism of similar intensity viewed stops from Days 1 and 2.*

Depart Connecticut Yankee Inn at 8 a.m. sharp.

- 0.0 Enter I95 at SR 161 westbound
- 20.6 Exit I95 at exit 64, Horse Hill Road. Turn left (south) crossing I95 to large road cut, west side of the road.
- 20.8 STOP 15.

**STOP 15. PERMIAN FOLIATED PEGMATITE (30 MINUTES)** Essex Quad. (Lundgren, 1964). The purpose of this stop is to show that sillimanite (fibrolite) can occur as a product of syntectonic-metasomatic reactions involving feldspars and need not reflect a sedimentary, aluminous protolith. This road cut exposes one of the larger granitic pegmatites cutting gneisses of the Bronson Hill terrane. The pegmatite is deformed, with porphyroclasts of K-feldspar up to 5 cm in diameter in a matrix of 1/2 - 1 cm K-feldspar-plagioclase-quartz-garnet-(biotite)-(muscovite) gneiss. Most surfaces expose chalky-white sheaths of nearly pure fibrolite similar to those described by Wintsch and Andrews (1987). These folia show a locally strong preferred orientation of sillimanite needles, and the folia themselves cut adjacent K-feldspar and plagioclase, and are locally partially replaced by retrograde muscovite. These relationships reflect metamorphic-metasomatic reactions like:



This reaction can be driven by high local mean stress and strain, but requires local metasomatism (loss of alkalis). The strong preferred orientation of the fibrolite within folia provide evidence that sillimanite was syntectonic and grew during a strain event in an open system. The presence of sillimanite need not reflect an aluminous sedimentary protolith.

- 20.8 Return to bus. Reenter I95 eastbound to Exit 65.
- 22.9 Exit I95 at exit 65 on Rt. 153. Turn left (north) on Rt. 53.
- 28.0 Turn right (south) on Rt. 154.
- 28.3 Turn left (east) into Wildwood Medical Specialists lot.
- 28.4 Drive to the far eastern end of the driveway to STOP 16.

## WINTSCH, SUTTER, KUNK, ALEINIKOFF AND BOYD

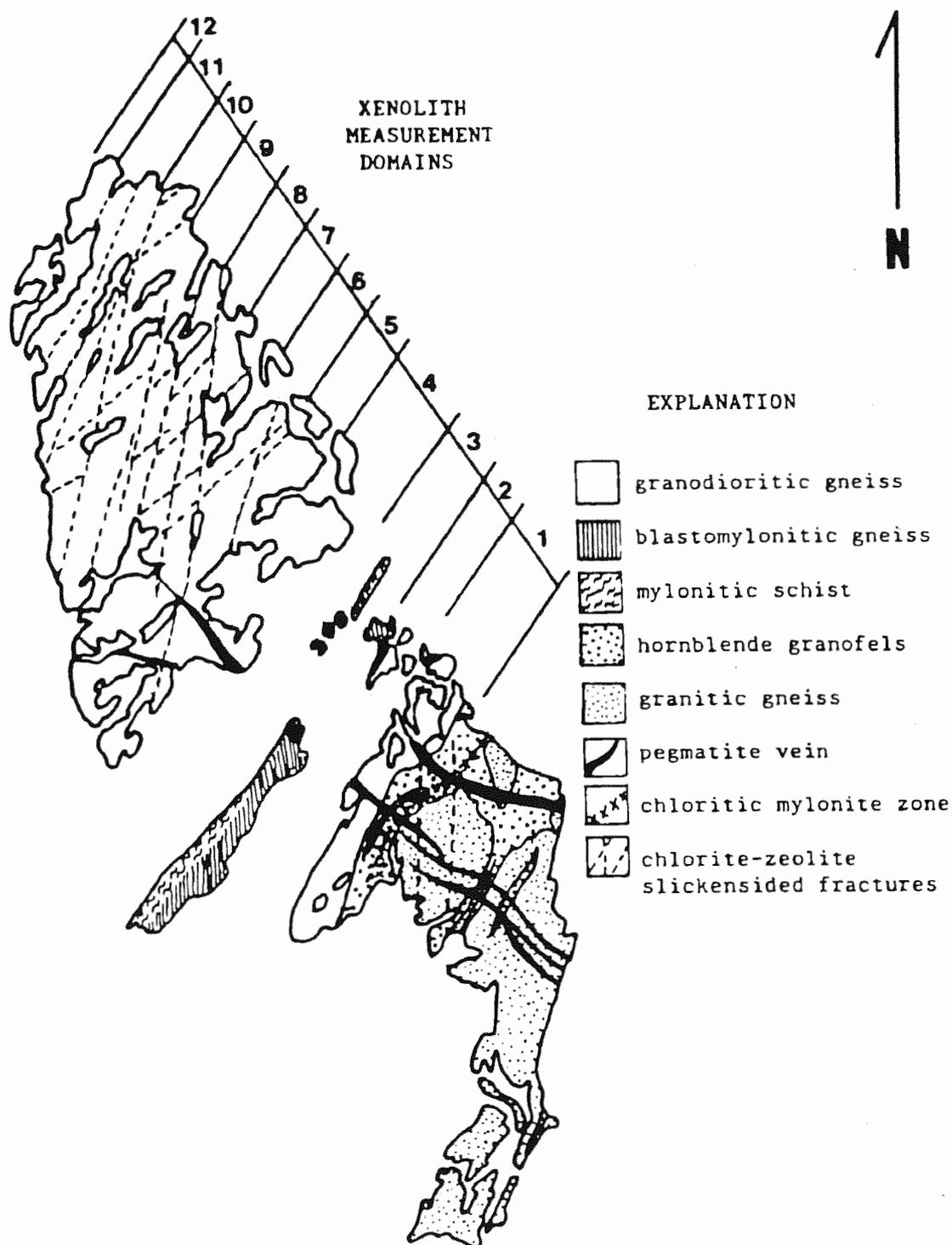


Figure S14.1. Pace and compass map of Stop 14 (from Ambers, 1988), showing the distribution and spectrum of rock types present, and xenolith measurement domains used for semiquantitative strain analysis.

## WINTSCH, SUTTER, KUNK, ALEINIKOFF AND BOYD

**STOP 16. GRANODIORITIC ORTHOGNEISS, AVALON TERRANE (30 MINUTES)** Essex Quad. (Lundgren, 1964; Wintsch 1985). The purpose of this stop is to show that high-grade, upper amphibolite facies metamorphism or not destroy alone primary igneous textures or compositions; only in shear zones and near migmatites are rocks modified. The outcrop shows many conspicuous amphibolite (andesitic) inclusions in a granodioritic gneiss (Fig. S16-1). These blocks could be xenoliths incorporated 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 blocks of diopside-bearing granofels with amphibolite (Wintsch, 1985).

The outcrop exposes several ductile shear zones cutting the coarse grained gneiss (Fig. S16.1). The gneissosity defined by disseminated biotite flakes dips 50°NW, and hosts a lineation defined by biotite streaks and stretched inclusions plunging 40°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. In the north-central part of the outcrop a ductile shear zone and a syntectonic migmatite form a conjugate set of shear surfaces cutting NW plunging fabrics, and indicating NE directed shortening. The migmatite monitors the temperature of this shortening, at ~650°C, which relates to the late Permian (Fig. 4) and at essentially the same time as the NE lineations of Stop 1; the temperature-time paths allows correlation of these structures. The presence of several shear fabrics of differing generations highlights the complications when mapping in these rocks. Foliations can not be correlated to identify larger scale structures unless it is first established that the foliations are of the same generation. The second message of this outcrop is that strongly layered gneisses need not reflect metavolcanics; the layering can be totally produced by strain.

28.4 Return to bus.

28.5 Turn right (north) on SR 154.

28.7 Turn right (east) at traffic light.

28.8 Turn left (north) into the north-bound entrance ramp to Conn. 9 at Interchange 3 where Rts 9 and 154 meet 2 km west of Essex, to STOP 17.

**STOP 17. MYLONITIC 'ORTHOSCHIST' DERIVED FROM AVALONIAN GRANODIORITIC ORTHOGNEISS (40 MINUTES)** Essex Quad. (Lundgren, 1964; Wintsch 1985; Dipple *et al.*, 1989). The purpose of this stop is to show that in very highly deformed rocks, metasomatism accompanying strong deformation can strongly modify the textures and compositions of rocks, but that in spite of highly strained rocks, the fault may not have very strong tectonic significance. These cuts expose a strongly layered and folded sequence of blastomylonitic biotite-sillimanite schists and biotite-plagioclase gneisses. Their southern contact is not exposed, but their northern contact with late Proterozoic orthogneiss of Stop 16 is gradational. Moving to the south, it is characterized by an increase in the quality of layering defined by the concentration of biotite (replacing hornblende) into discrete foliation planes, and by an increase in the amplitude and a decrease in the wavelength of folding in these layered rocks. The presence of these intercalated sillimanite-bearing schists and the assumption of isochemical metamorphism caused Lundgren (1964) to map these rocks as part of the pelitic Putnam-Nashoba terrane (Stops 2, 6, 10), but the gradational contact between these schists and the orthogneiss makes this interpretation difficult to defend.

To test whether these schists could have been derived from the enclosing orthogneiss of Essex, (Dipple *et al.*, 1989) analyzed the rocks and minerals of the blastomylonites and the enclosing orthogneiss. Using two-sample mass balance calculations among 80 bulk rock analyses they show that the orthogneiss of Stop 16 was segregated into biotite-rich schists, and biotite poor quartz-plagioclase gneisses through cm scale transfer of Si, Al, Ca, Na, Ba, and Zr from schists to gneisses. However, incongruent dissolution of plagioclase shifted the rocks from diopside to corundum normative and stabilized sillimanite and garnet in some schists (Fig. S17.1). Three and four-sample mass balance calculations show that Al, Ca, Na, and Ba were lost and Si and K gained in these rocks. The extraction of quartz and plagioclase from the orthogneiss: (1) concentrated biotite into discrete foliation planes, and (2) enriched the gneissic layers in quartz and plagioclase. This 'unmixing' occurred on the scale of a few to tens of cm to create the strongly layered rock present in these cuts (Fig. S10.2). Some of the quartz and plagioclase removed from the

WINTSCH, SUTTER, KUNK, ALEINIKOFF AND BOYD

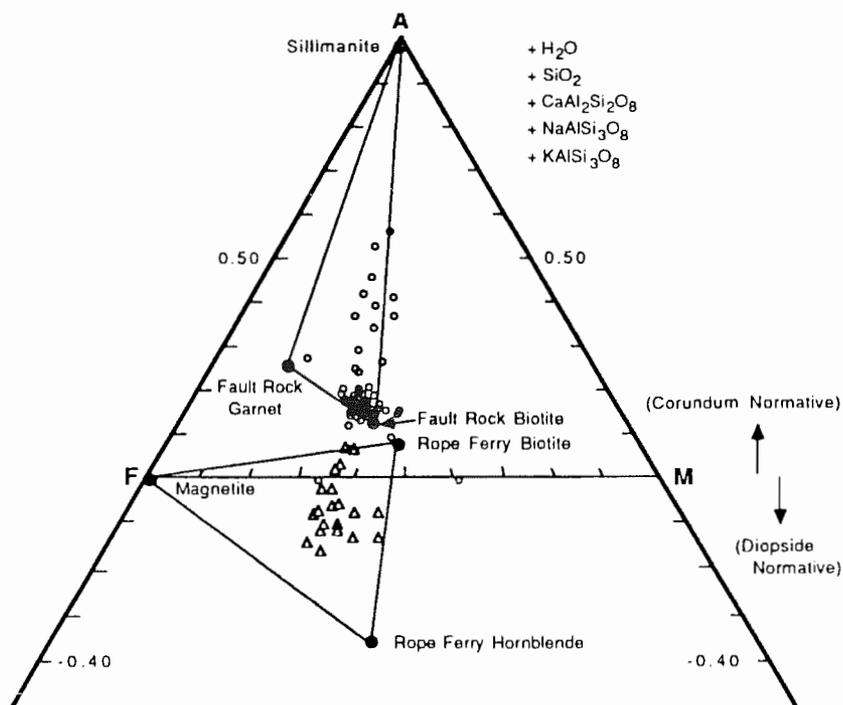
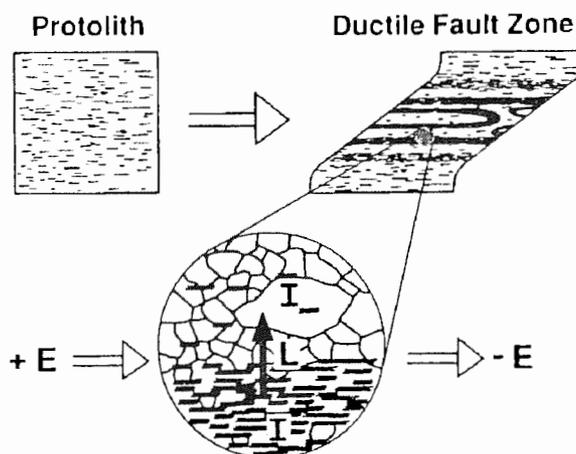


Figure S17.1. The bulk compositions of Rope Ferry (ortho)gneiss (A) and mylonite schists (o) at Stop 17. The common Fe/Mg ratio of all rocks defines a vector that passes through the aluminum oxide axis, and shows that the aluminous schists could be derived from the host gneiss by dealhalization (Dipple *et al.*, 1989).

schists are also concentrated in vein and migmatitic structures. Evidence that quartz and plagioclase were added to the gneissic layers comes from the large poikiloblastic nature of the quartz and plagioclase grains that locally include an earlier biotite foliation. Thus all the evidence supports the hypothesis that the blastomylonitic rocks in these cuts were derived from the host orthogneiss through some structurally-induced metamorphic differentiation process associated with the high strain in this fault zone.

Figure S17.2. Conceptual model for the development of mylonitic schist from Rope Ferry (ortho)gneiss. I, L, and E indicates components with a low, moderate, and high differential molalities, respectively.



Hunts Brook Fault Zone	
I = Mg, Fe, Ti, Mn, V, Zn	+ E = Si, K
L = Si, Al, Ca, Na, Ba, Zr	- E = Al, Ca, Na, Ba

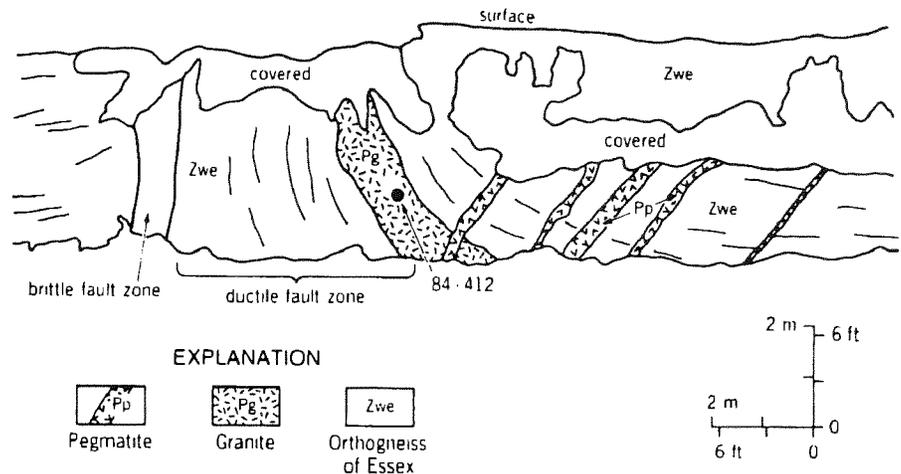
- 28.8 Return to bus. Proceed north onto SR 9.
- 29.4 Cuts on east side of SR9. STOP 18.

## WINTSCH, SUTTER, KUNK, ALENIKOFF AND BOYD

**STOP 18. GRANODIORITIC ORTHOGNEISS, AVALON TERRANE (30 MINUTES) Essex Quad.**

(Lundgren, 1964; Wintsch, 1985; Wintsch and Aleinikoff, 1987). The purpose of this stop is to show how geochronology can be used to indirectly date the deformation in these Avalon terrane rocks. The most important rock here is the plagioclase gneiss of Stop 16, although containing a lower density of amphibolite xenoliths. Foliation dips gently to the northeast, but is cut by a sucrosic textured, unfoliated, weakly layered granite dike, and this in turn is deformed by a vertical dextral shear zone believed to be associated with deformation along the Falls River fault (Fig. 8). Zircons from the gneiss define a discordia whose upper intercept (620 Ma) is interpreted to reflect the crystallization age of the gneiss, and whose lower intercept (270 Ma) is interpreted to reflect the time of secondary event causing overgrowths on the zircons (Wintsch and Aleinikoff, 1987). Zircons in the granite dike (Fig. S18.1) define a

Figure S18.1. Tracing from a photograph of the road cut of Stop 18 (facing east), showing the foliated Rope Ferry (ortho)gneiss intruded by a Permian granite dike (sampled for zircon and monazite, #84-412), folded near a steeply dipping ductile fault zone, and then reintruded by feldspathic veins (hydrothermal pegmatites).



similar discordia, reflecting the source of the zircons as xenocrystic, and not the age of the dike. Monazite in the dike, however, is ~270 Ma, similar to the age of sphene in the plagioclase gneiss. Thus the granite intruded the gneiss when the gneiss was at the closure temperature of sphene (~500-600°C). 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. Because the syenitic pegmatite dikes are undeformed, and younger than the time of sphene closure, they must have formed at a temperature less than the temperature of sphene closure; they cannot be magmatic, they must be hydrothermal.

- 29.4 Return to bus. Proceed north on SR9.
- 30.1 Exit Rt. 9 at interchange 4, turn left (north) on SR 154.
- 31.4 Turn right (east) on Southworth St.
- 31.5 Turn left (north) into parking lot of Town Park. STOP 19.

**STOP 19. AVALON-MERRIMACK TERRANE BOUNDARY HONEY HILL FAULT (30 MINUTES) Deep River Quad.** (Lundgren, 1963; Wintsch, 1985). The purpose of this stop is to show that here the Honey Hill fault juxtaposes Avalon and Merrimack terrane rocks; with no Putnam-Nashoba rocks present. In spite of this, the contact does not appear highly strained. In the trench west of the ball field excavated by Janet Stone for this trip is exposed the contact between blastomylonitic schists of the Avalon terrane, and granular, migmatitic schists of the Merrimack terrane. The contact zone is intruded by several pegmatites, some highly deformed and boudinaged, and others later dikes only weakly deformed. The significance of this exposure is the lack of Putnam-Nashoba rocks, and the lack of classical evidence for high strain. Rocks south of the park along SR 154 are cataclastic, showing reactivation and overprinting of greenschist facies fabrics on the higher grade gneisses of the Merrimack terrane.

## WINTSCH, SUTTER, KUNK, ALEINIKOFF AND BOYD

- 31.5 Return to bus, leave turning right (west) on Southworth St.
- 31.6 Turn right (north) on SR154 (Main Street).
- 32.0 Turn left (west) on SR80 at traffic light.
- 32.6 Turn right (north) on SR9.
- 33.0 Bus will stop on SR9, passengers with hard hats and reflective vests have permission to examine rocks in road cut, east side of highway. STOP 20.

**STOP 20. BONEMILL BROOK FAULT ZONE, BRONSON HILL TERRANE (30 MINUTES)** Deep River Quad. (Lundgren, 1963; Wintsch, 1985). The purpose of this stop is to examine some of the very highly strained rocks in the hanging wall of the Bonemill Brook fault zone. These rocks are a strongly layered quartz-plagioclase-biotite-(anthophyllite) granofels and schist interlayered with strongly boundinaged pegmatites and thin biotite-bearing amphibolites. The granofelsic rocks are locally strongly differentiated at the grain scale (1 mm) into monomineralic layers, but also contain strongly feldspathized regions, where porphyroblasts are interpreted to reflect feldspar growth in the solid state rather than dismembered pegmatites. The most conspicuous structures within the foliation plane are sub-horizontal biotite streaks that can be traced the length of the exposure. Hornblende on the edges of the amphibolite boudins has reacted with  $K^+$  (probably from dissolving K-feldspar) to form biotite +/- epidote. The concentration of biotite on the margins of these strong amphibolite layers produces a very weak schist, which tends to localize some of the strain. The strain, in turn, helps drive the reaction to produce more biotite. This self feeding reaction softening process tends to destroy the amphibolite, and its ability to boudinage; it evolves with reaction progress from being the strongest to the weakest rock. Consequently, some small (10 cm thick) amphibolite boundins are sheathed in biotite schist, and schist layers only 1-2 cm thick may join boudins many meters apart. The nearly mylonitic textures and very strongly developed biotite streaks are interpreted to reflect middle-late Permian dextral strike-slip motion generated as the Avalon terrane moved NW against this Bronson Hill buttress. We speculate that the Bonemill Brook fault was active as trust in the early Alleghanian, but has been folded and overturned by the Chester School anticline (Fig 8); the lower amphibolite facies textures present in this outcrop (cooler, and thus younger than sphene cooling ages of ~ 285) apparently completely overprint this earlier history.

*End Road Log. Return to bus. Follow 9N to Exit 6, exit, and reenter southbound. Follow SR9 south to I95 east, to Boston.*

Chapter I

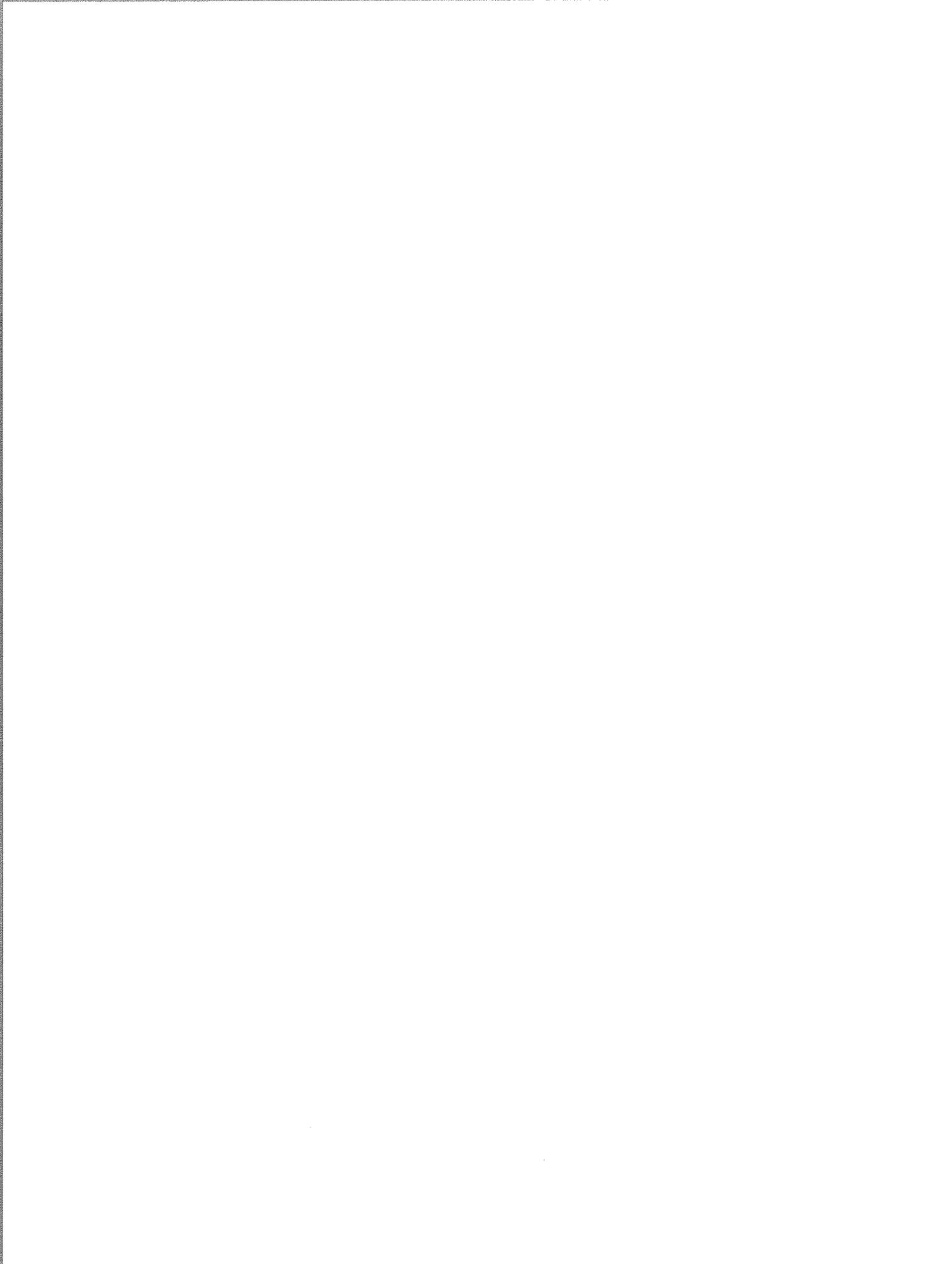
# The Late-Glacial Marine Invasion of Coastal Central New England (Northeastern Massachusetts - Southwestern Maine): Its Ups and Downs

By Thomas K. Weddle, Carl Koteff, Woodrow B. Thompson, Michael J. Retelle, and Cheryl L. Marvinney

Field Trip Guidebook for the Northeastern United States:  
1993 Boston GSA

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**THE LATE-GLACIAL MARINE INVASION OF COASTAL CENTRAL NEW ENGLAND  
(NORTHEASTERN MASSACHUSETTS - SOUTHWESTERN MAINE): ITS UPS AND DOWNS**

by

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**INTRODUCTION**

Retreat of the late Wisconsinan Laurentide Ice Sheet from the Boston area into southwestern Maine between 15,000 to 13,000 yr BP was accompanied by invasion of the late-glacial sea. Glacial and glacial marine sedimentation within the area of marine transgression is recorded by the surficial deposits. Recently, these deposits have been mapped on large-scale 7.5-minute surficial geologic quadrangles, and delta elevation measurements and detailed stratigraphic section logging have been recorded. Timing of late-glacial events and ice retreat in the region has been better constrained by datable materials found in association with ice-marginal positions (Dorion, 1993).

The precisely surveyed altitudes of ice-marginal deltas deposited during systematic ice retreat provide a record of sea level fluctuations during this time, and characterize the nature of postglacial uplift in the region. Eustatic sea level appears to have been relatively unchanged from the uncovering of the Boston area at about 15,000 yr BP until 14,000 yr BP, when the ice margin was in southwestern Maine and adjacent New Hampshire. Then, within 50 to 100 years, sea level appears to have risen sharply between 20 to 30 feet (7 to 10 m) before leveling off. This information relates terrestrial evidence to oceanic studies that suggest little appreciable meltwater was added to the ocean system until after 14,000 yr BP.

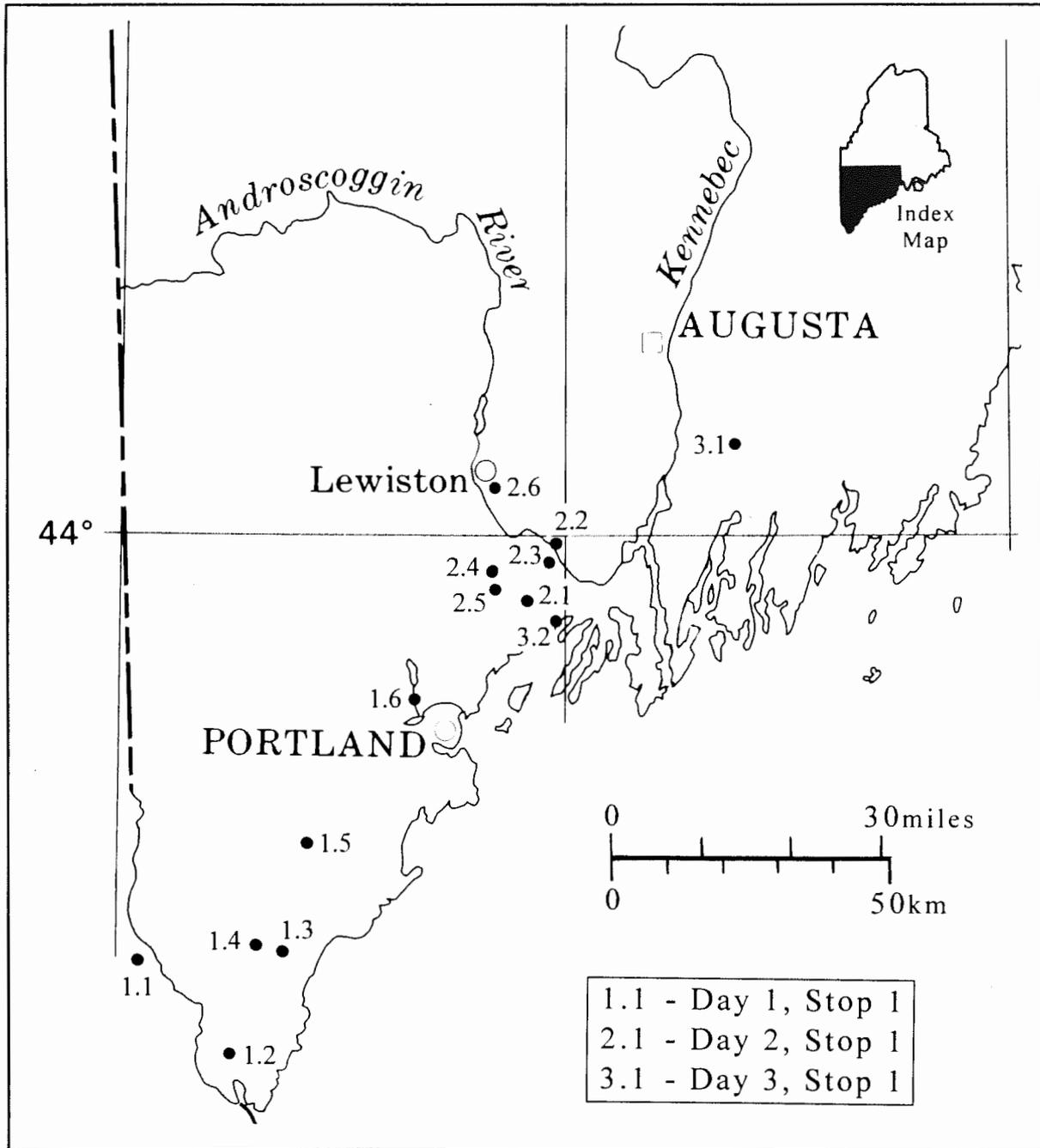
This trip will visit excellent examples of glacial marine deposits including deltas, fans, stratified end moraines, and fossiliferous glacial marine mud, as well as nearshore deposits associated with synglacial sealevels and later marine regression resulting from postglacial uplift (Figure 1; general field trip map). In addition, examination of precisely obtained delta topset/foreset contact altitudes will demonstrate that a model of delayed postglacial uplift derived earlier in central New England (Kotteff and Larsen, 1989) is supported by the coastal evidence in southwestern Maine and New Hampshire, and that this model can be used to show extremely rapid changes in late glacial sea levels.

**DEGLACIATION OF COASTAL CENTRAL NEW ENGLAND**

The nature of retreat of the late Wisconsinan ice sheet from terrestrial southern New England has been summarized by Kotteff and Pessl (1981). They characterized the Laurentide ice margin as active ice, systematically retreating northward with a narrow stagnant ice margin flanking live ice, and termed this mode of deglaciation "stagnation-zone retreat." As the margin retreated, deposits known as morphosequences were laid down by meltwater, and they record a relative chronology of ice recession and deposition (Kotteff, 1974). The width of the stagnant zone varied depending on local topography, and generally appears to have been only a few kilometers. Minor ice readvances have been found, but the pattern of systematic and steady ice retreat is persistent throughout the region. Detailed surficial geologic mapping in southern and central New England permits a timing of events based on a very consistent relative chronology of meltwater deposition to be constrained by the few available radiocarbon dates.

The morphosequence concept was originally applied to glacial lacustrine and glacial fluvial features, but is equally well applied to the glacial marine environment, especially in the coastal central New England area where

the marine-based ice sheet was grounded rather than floating. Although a marginal stagnation zone may not have existed along the marine-based margin, features such as the numerous washboard moraines and ice-shoved proximal portions of deltas and fans attest to nearby active ice throughout the deglacial history of the area. The glacial marine fans and deltas associated with the ice margin were deposited from subglacial tunnels or from glacial streams at their termini. Ice-marginal deglacial-phase glacial marine deposits, in particular in the coastal zone of Maine, have been described recently in detail by Attig (1975), Stemen (1979), Jong (1980), Ackert (1982), Lepage (1982), Smith (1982), Smith, Stemen, and Jong (1982), Thompson (1982), Miller (1986), Smith and Hunter (1989), Retelle and Bither (1989), Thompson and others (1989), and Ashley and others (1991). The ice-marginal deposits to be seen during this field trip are similar in their internal sedimentological character to the deposits described in the cited studies as examples of deposits from a temperate marine grounded ice-sheet.



**Figure 1.** Location map and approximate sites of field trip stops.

## Southwestern Coastal Maine

Because most of the field trip will be in southwestern Maine, the following section summarizes the deglacial history of Maine's coastal lowland in the region between the Kennebec River and the New Hampshire border, and includes a brief description of the major types of glacial and glacial marine deposits that formed during ice retreat. These deposits collectively record the recession of the late Wisconsinan ice sheet in a marine environment, and also provide evidence of the regression of the sea due to postglacial isostatic uplift. The Surficial Geologic Map of Maine (Thompson and Borns, 1985a) published by the Maine Geological Survey provides an overview of the area visited on this trip.

**Previous Work and Current Research.** The late-glacial deposits in southern Maine constitute one of the best-exposed and most accessible glacial marine sequences in the continent, and have been examined by geologists since the 1830's. Pioneering studies were done by Stone (1899) and Leavitt and Perkins (1935). However, there were few detailed investigations until the 1970's, when the Maine Geological Survey began a surficial geologic quadrangle mapping program. This work has provided a foundation for topical studies focused on systematic description and interpretation of glacial marine deposits (cited above), and summarized by Thompson and Borns (1985b). There is also continuing interest in using Maine's glacial marine deltas and shoreline features as indicators of paleosea-level and glacial isostasy (Belknap and others, 1987; Thompson and others, 1989; Kelley and others, 1992; Koteff and others, in press). Research in Maine has been further stimulated by the concurrent surge of interest in glacial marine deposits in other parts of the world (e.g., Dowdeswell and Scourse, 1990; Anderson and Ashley, 1991). The chronology of deglaciation in the region of the marine submergence is better known than in the part of Maine above the marine limit, due to the abundance of shells and other organic materials in the glacial marine mud of the Presumpscot Formation (Stuiver and Borns, 1975; Smith, 1985). Currently, researchers from the University of Maine and the Maine Geological Survey are collaborating in an intensive search for sites where organics are included in ice-marginal deposits and can be collected for AMS-radiocarbon age-dating.

**Deglaciation History.** It has been suggested that the late Wisconsinan glacier margin receded to the position of the present coastline in extreme southwestern Maine by about 13,800 radiocarbon years ago (Smith, 1985). This conclusion is based on a single date from a coastal bluff in Kennebunk, where shells were found in marine sediments between till and an overlying gravelly unit. These shells were dated at  $13,830 \pm 100$  yr BP (QL-192; Smith, 1985). However, recently obtained dates on *Portlandia arctica* shells indicate ice retreat as far north as Scarborough by about 14,800 yr BP, to Freeport by about 14,000 yr BP, and to Lewiston by about 13,000 yr BP (H. W. Borns, Jr., unpub. data). More data are needed to establish a firmer chronology of deglaciation, but new dates in both eastern and western Maine (Dorion, 1993, and unpub. data) suggest that ice retreat occurred at least a few centuries earlier than previously believed. Dates obtained by Anderson and others (1992), and Dorion (1993, and unpub. data), have also verified earlier claims that the ice margin receded to the marine limit in central Maine by about 13,300 yr BP. At this time, drawdown of the Laurentide Ice Sheet in the St. Lawrence Lowland of Quebec had already begun to isolate the northern Appalachian ice mass from the main ice sheet (Dyke and Prest, 1987).

As the ice sheet withdrew into central Maine, glacial isostatic uplift eventually exceeded eustatic sea-level rise, causing regression of the sea. A wide range of radiocarbon dates have been obtained from shallow-water shell assemblages associated with the marine offlap (about 13,000 to 10,000 yr BP; Smith, 1985; Retelle and Bither, 1989). However, Smith and Hunter (1989) suggest that southwestern coastal Maine was completely emerged by 11,500 yr BP. As relative sea level fell, glacial sediments were considerably reworked by wave and current action. This process formed a patchy cover of shoreline and nearshore deposits over older Pleistocene sediments. Large sand plains locally overlie the Presumpscot Formation at elevations below the marine limit. These sand plains may have likewise resulted (at least in part) from nearshore marine erosion, though they tend to occur where rivers were establishing their postglacial courses and redistributing sediments in response to lowering base level (Borns and Hagar, 1965).

**Morphostratigraphy.** The systematic northward retreat of the marine ice sheet is documented by a complex association of sediments and landforms deposited along the former grounding lines of the tidewater glacier

margin throughout the coastal lowland. The most common ice-marginal deposits are composites of linear end moraines, submarine fans and deltas which are products of meltwater discharge into the marine basin at the grounding line of the ice sheet.

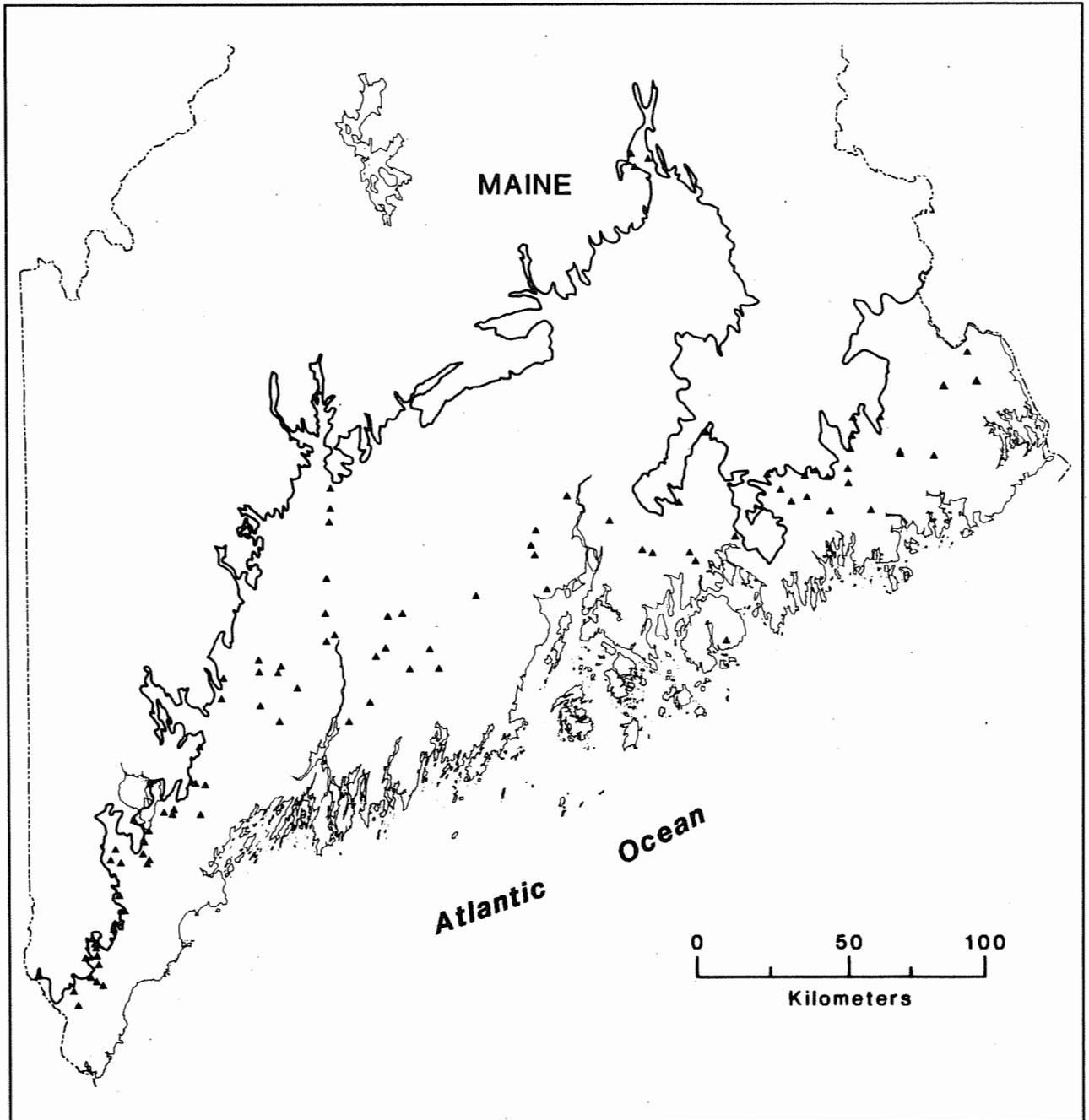
**End moraines.** There are thousands of end moraines in the coastal lowland. They occur in clusters of parallel ridges, many concealed beneath younger ice-marginal deposits or blanketed by fine-grained glaciomarine mud (Smith, 1982; Thompson, 1982). The principal characteristic shared by most moraines is their ridge morphology, or at least a linear alignment along a former ice-marginal position. In general, the moraines are 3 - 45 feet (1 - 15 m) high, and 15 - 45 feet (5 - 15 m) wide, with some larger moraines exceeding 300 feet (100 m) in width. Most in southwestern Maine are minor moraines, referred to as washboard or DeGeer moraines (Smith, 1982; 1985) and are typically a few feet high to several hundred feet long and are regularly spaced around 200 - 250 feet (65 - 70 m). Larger moraines are generally more complex in morphology and composition. Many of the moraines are comprised of diamicton including lodgement till, debris flow deposits and glaciotectonized stratified drift. The larger landforms, commonly referred to as stratified end moraines (Borns, 1973; Ashley and others, 1991) consist of stratified sand and gravel interbedded with minor amounts of diamicton. The stratified moraines usually exhibit evidence of glaciotectonism, with the interstratified materials folded and thrust-faulted by overriding ice or ice-shove. The occurrence of the many moraines and the intimate stratigraphic relationship with glaciomarine sediments indicate that: (1) the grounded ice margin was active during its northward retreat, and (2) ice retreat was accompanied immediately by marine incursion of the coastal lowland as ice was in contact with the sea in a calving bay regime (Smith, 1982; Hughes and others, 1985).

**Eskers, submarine fans, and deltas.** Sediment deposited by meltwater occurs in several genetically connected and related landforms in the ice-marginal zone and include eskers, submarine fans, and deltas. Eskers (Thompson and Borns, 1985a) ice-tunnel deposits (Ashley and others, 1991) or subglacial conduit facies (Sharpe, 1988) are terms used to describe the coarse-grained sediments from a predominantly high-energy fluvial source that delivers sediments to the ice-proximal zones of fans and deltas. These deposits occur as distinct and separate sinuous ridges in valleys above marine limit (cf Thompson and Borns, 1985a), as feeder "tails" on ice proximal sides of deltas (Thompson and others, 1989), and as coarse-grained cores of fans and deltas where the supporting ice has retreated and a delta or fan has prograded basinward over its former conduit.

Submarine fans originate at the mouth of the esker or meltwater conduit and grade distally to the seafloor. The submarine fans, also referred to as submarine outwash (Rust and Romanelli, 1975), subwash fans (Burbidge and Rust, 1988), and grounding line fans (Powell, 1990) are irregular mound, cone or fan-shaped features that commonly drape or flank the moraine ridges. Over time, particularly if the retreat of the grounded ice margin is halted at a pinning point on a subglacial topographic high or a valley constriction, a submarine fan may aggrade vertically and prograde distally and evolve into a more massive and extensive deposit that approaches and may eventually even reach contemporaneous sea level. In this latter and most developed case, the fan may eventually become a delta with a flat subaerially exposed topset plain (Powell, 1990; Slayton, 1993).

The fan sediments represent the transition between fluvially dominated processes associated with the ice tunnel environment and processes of the proglacial marine basin. Consequently, materials in the fans show complex stratigraphic relationships both parallel to the ice margin and distally from the former tunnel mouth. Sediments in the proximal zone include coarse-grained gravelly stratified materials supplied by the meltwater currents and diamicton that may originate from slope failure and downslope movement from the adjacent moraine, or ice thrust (Retelle and Bither, 1989; Ashley and others, 1991). In medial and distal zones of the fans the major facies include rhythmically bedded sand and mud that grade to the muddy seafloor. These sediments, termed cyclopsams and cyclopels (Mackiewicz and others, 1984) were likely deposited by suspension from a highly turbid but buoyant suspended sediment (Cowan and Powell, 1990). Resedimentation of these deposits, particularly in the proximal and medial portions of the fan produces grain flows and debris flows of varying textural composition. Stratigraphic sections in the medial and distal portions of the fans are commonly fining-upward sequences (Retelle and Bither, 1989) that grade transitionally upwards from coarse proximal fan sediments at the base to distal fan sediments. The fining-upward sequence may either represent ice retreat with removal of the ice tunnel sediment source or a lateral switching of a distribution channel on the fan lobe.

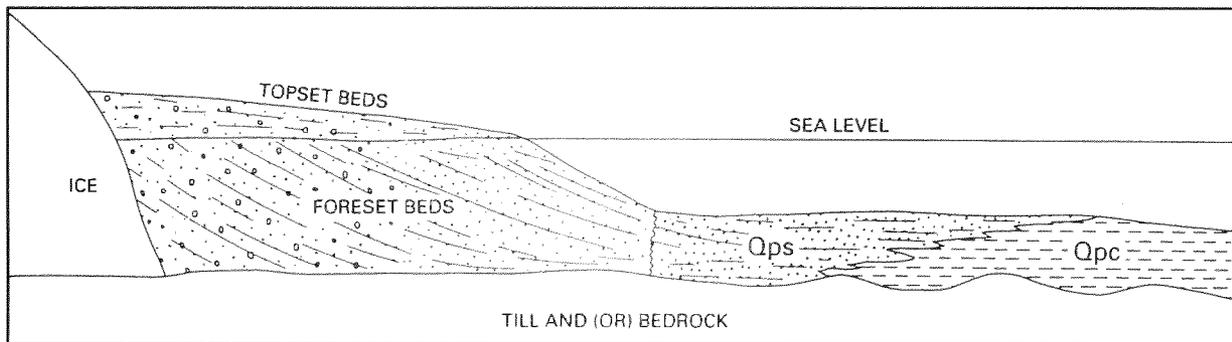
Over 100 glacial marine deltas occur in southern Maine. The origin and stratigraphy of these deltas has been described and categorized by Thompson and others (1989). Although they are widely distributed across the coastal lowland, the largest examples are located at or near the inland marine limit in southwestern and eastern Maine (Figure 2). Many are localized where the ice margin was temporarily pinned against bedrock ridges and other topographic highs. Four categories of glacial marine deltas in Maine include ice-contact deltas, esker-fed ice-contact deltas, leaside deltas, and distal outwash deltas (Thompson and others, 1989). Point-source



**Figure 2.** Marine limit and locations of deltas in Maine (Thompson and others, 1989).

sedimentation for the first two types is indicated by eskers that connect with the ice-proximal side, or by stratigraphic and morphologic features in the proximal part of the delta. Leaside deltas formed in topographically controlled situations where the ice margin was pinned against a ridge that protruded above sea level and separated the ice margin from the open ocean. Meltwater flowed through a gap in the ridge and deposited the delta on the lee side. In most cases there was only a short distance between the ice margin and the delta head, and an esker may lead up to or pass through the gap. Distal outwash deltas, formed where meltwater streams entered the sea some distance from the ice margin, are rare in Maine. Isostatic uplift was in progress across much of the region as the ice withdrew inland from the marine limit. Outwash reaching the sea was distributed as a thin fluvial sand across the Presumpscot Formation as relative sea level dropped in response to uplift.

Many pit exposures in deltas show that they are Gilbert-type deltas, with horizontal fluvial topset beds (delta-plain deposits) overlying inclined foreset beds deposited on the prograding delta front (Figure 3). These deltas are typical of environments where coarse-grained sediments are rapidly deposited in basins of sufficient depth to produce a steeply-sloping delta front (Nemec, 1990). The contact elevation between the topset and foreset beds in Maine's deltas approximates the late-glacial sea level to which the deltas were graded. Thompson and others (1989) surveyed the topset/foreset contact elevations to define the plane of the upper marine limit. Isostatic uplift has tilted the plane to the southeast, and it now has an average slope of 2.82 ft/mi (0.53 m/km) in the Kennebec River Valley region. However, this is only a minimum gradient for postglacial tilt, because any uplift during ice retreat would have lowered relative sea level and reduced the slope of the delta elevation profile. An estimate of postglacial tilt in the lower Androscoggin River Valley using elevations of delta tops from topographic maps and topset/foreset contact measurements from Thompson and others (1989) is between 3.33 and 3.75 ft/mi (0.63 and 0.71 m/km). Koteff and others (in press) have determined the postglacial tilt from



**Figure 3.** Schematic drawing showing relation between glacial marine deltas and the Presumpscot Formation: Qps (sandy facies Presumpscot Formation) is the bottomset beds laid down as extensions of the foreset beds; Qpc (silty-clay Presumpscot Formation) is the distal mud deposited beyond the delta on the submarine plain (from Koteff, 1991).

deltas in coastal New Hampshire and southwestern Maine to be 4.5 ft/mi (0.85 m/km), supporting their proposal for delayed uplift in southern New England (Koteff and Larsen, 1989). The discrepancies between the regions are partly the result of recent topset/foreset contact measurements and re-interpretation of previously measured delta elevations in the Maine - New Hampshire border area (Koteff and others, in press). Moreover, if uplift had begun by the time the ice margin was in the lower Androscoggin River Valley, a less steep postglacial gradient of the water plane would be expected in the field trip area.

As noted elsewhere in this guide, recent work has shown that the deltas in extreme southern Maine were inundated by the rising late-glacial sea (Koteff and others, in press). The deltas have been wave eroded to varying degrees, including partial or even complete removal of the topset beds. Consequently, shoreline or nearshore deposits of sand and gravel may unconformably overlie the eroded delta foresets, producing a

stratigraphic relationship which may at first be confused with topset/foreset contacts. This reworking by marine processes also explains the scarcity of preserved meltwater channels on delta tops in southwestern Maine. Glacial marine deltas elsewhere in the state have distributary channels on the delta plain (Stop 1, Day 3), and local terracing of channels on some of the larger deltas in eastern Maine (e.g, Silsby Plain and Pineo Ridge) indicates that uplift exceeded eustatic sea-level rise as the deltas were forming.

**Presumpscot Formation.** Glacial marine mud has been named the Presumpscot Formation in Maine by Bloom (1960; 1963); the unit has been mapped and extended into New Hampshire by Koteff (1991). Subsurface data and surface exposures show that this unit directly overlies bedrock, till, fans, and end moraines, and is interbedded with medial and distal components of fans and deltas (schematically represented in Figure 3.) It can be massive or layered, containing outsized clasts and ice-rafted debris, and is fossiliferous, with a rich assemblage ranging from foraminifers and diatoms through mollusc shells to marine vertebrates. It has a blue-gray color when fresh, and an olive gray color when weathered. Fractures in the weathered Presumpscot Formation can have iron-manganese(?) stain along them. Trefethen and others (1947) previously considered the weathered mud to be a different unit from the blue-gray mud, deposited by different glacial advances. However, Leavitt and Perkins (1935), Goldthwait (1949, 1951), Caldwell (1959), and Bloom (1960, 1963) attributed the color difference to postglacial oxidation. Kelley (1989) and Mayer (1990) have discussed the mineralogy of the Presumpscot Formation from both offshore and onshore samples.

The Presumpscot Formation was deposited by glacial fluvial discharges directly into the glacial sea, the winnowed fine-grained fraction settling out as glacial marine mud (processes discussed in Retelle and Bither, 1989). A sandy facies of the Presumpscot Formation, found overlying the fine-grained facies, is common in southwestern Maine (Thompson, 1982, 1987). The contact between the facies may be sharp or gradational. The origin of the sandy facies is unclear, however, it appears to be associated with shoaling during the regression of the sea. Stop 2 of Day 3 will be an opportunity to see and discuss the fine-grained and sandy facies of the Presumpscot Formation. Two recent symposia on the Presumpscot Formation provide an excellent overview of its geologic and engineering characteristics (cf. Andrews and others (1987), and Northeastern Section Geological Society of America Abstracts with Programs, 1988, SEPM symposium: "Glaciomarine sedimentation: the Presumpscot Formation of northeastern North America and analogues").

**Shallow Marine Deposits.** Shallow or nearshore marine sediments were deposited after the ice margin and sediment source retreated from a depositional center, either as sea level overstepped the landform, when isostatic sea level rise was greater than isostatic rebound, or as sea level fell around the landform due to isostatic rebound exceeding eustatic sea level rise. The shallow marine facies (Smith, 1985; Retelle and Bither, 1989) contains a range of litho- and biofacies ranging from well-sorted tidal to subtidal sand to coarse bouldery lag deposits or lagoonal mud. The deposits include a variety of morphostratigraphic units such as beaches, spits, tombolos, subtidal sand bodies, and an extensive and nearly ubiquitous veneer of wave-reworked sediments. The deposits also display a wide range of textural maturity reflecting the source landform subjected to wave reworking and the available energy.

#### ACKNOWLEDGEMENTS

The field trip leaders want to thank the landowners and the Maine Department of Transportation who allowed access to the stops. Maine State Geologist Walter Anderson is acknowledged for his support of these activities. Thanks to Robert Johnston and John Poisson (Maine Geological Survey) who drafted figures for this guide. Finally, Dr. George Denton, Dr. Harold Borns, Jr., Chris Dorion (all from the University of Maine), and Gwyneth Jones (formerly Bates College) are thanked for their support and contributions through the University of Maine EPSCoR grant which provided funding for radiocarbon dating of shells found in the region.

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## ROAD LOG

There is no mileage log; however, all field trip stops are located on Figures 4, 6, and 9; the stop locality maps are from USGS 1:250,000 scale topographic maps upon which the field trip route can be traced (Portland, Lewiston, Bangor, Bath quadrangles). Stop 1 (Day 1) is in New Hampshire; all subsequent stops will be in Maine. A location paragraph for travel to each stop contains landmark references; the stop description gives the estimated time to spend at each pit, the 7.5-minute USGS topographic quadrangle map where the stop is located, and the approximate elevation of the stop. NOTE: those visiting sites on their own must obtain landowner permission before venturing on private land. For future reference, locating sites could be facilitated by purchase of The Maine Atlas and Gazetteer and The New Hampshire Atlas and Gazetteer, both published by DeLorme Mapping Company, P.O. Box 298, Freeport, Maine 04032 (207-865-4171).

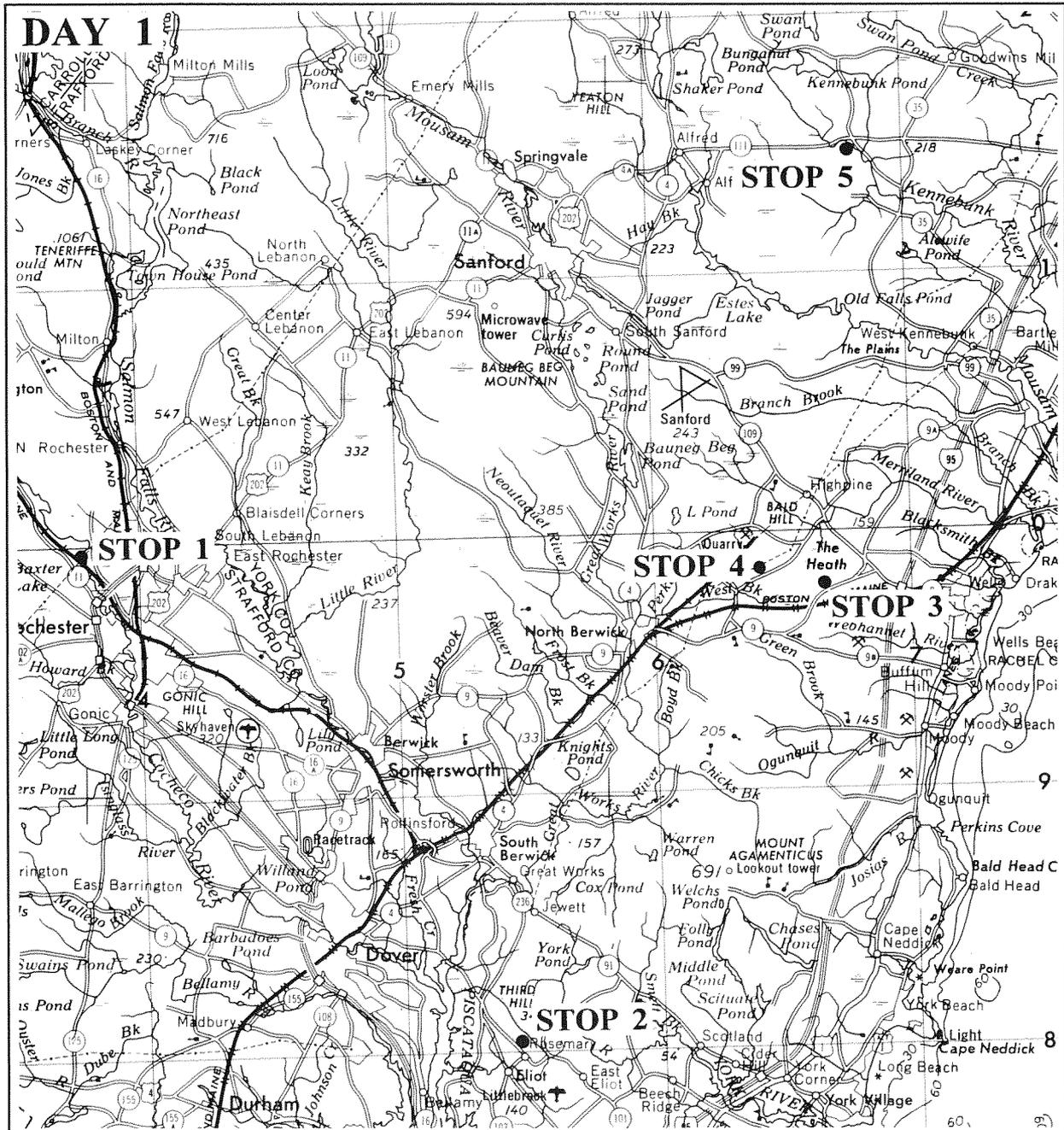


Figure 4. Stop locality map, Day 1 (Portland 1:250,000 sheet).

## ROAD LOG - DAY 1 (10/22/93)

Refer to locality map (Figure 4) for directions and approximate distances of travel. **Lunch for today only will be at some predetermined stop (pay on your own).** We will visit exposures in glacial

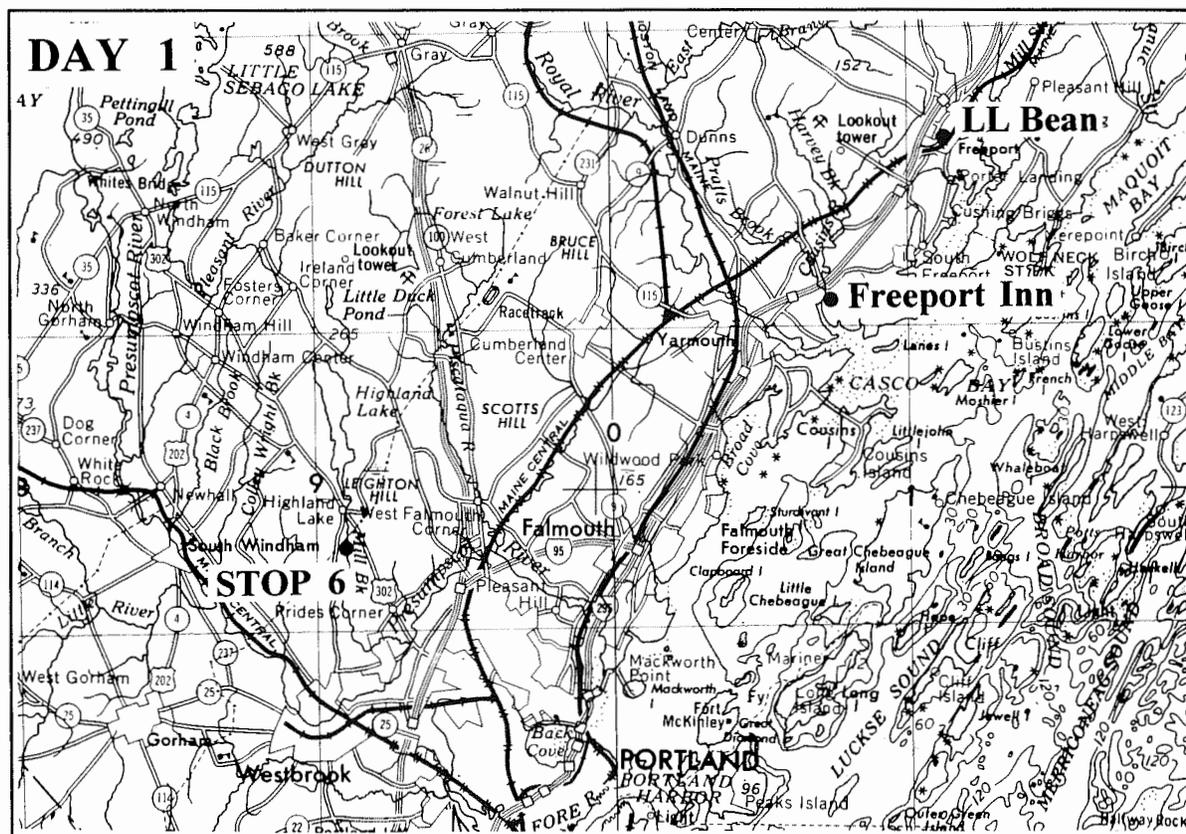


Figure 4 (Continued). Stop locality map, Day 1 (Portland 1:250,000 sheet).

marine fans and deltas deposited prior to uplift and which have been trimmed by synglacial and postglacial sea levels, see evidence for a sharp rise in sea level of 20 - 30 feet (7 - 10 m), a cut in a drumlin, and an end moraine-fan complex where superb striations and meltwater abrasion features on bedrock are present.

*Assemble at field trip staging area in Boston by 7:30AM, October 22, 1993, and leave at 8:00AM by bus. Travel to I-95 North heading to New Hampshire and Maine. Approximate travel time to Stop 1 is 1.5 hours. From I-95 in Portsmouth, take Spaulding Turnpike (Route 16) in New Hampshire for approximately 20 miles; take exit 15 onto Route 11 West. After about 2 miles, turn right on Little Falls Bridge Road and proceed approximately 1 mile to pit entrance on left.*

**STOP 1. TORR PIT** (60 minutes; Baxter Lake quadrangle, elev. ca. 250 feet (76 m); leader Carl Koteff).

This exposure is in the distal part of a glaciomarine delta deposited in the Cocheco River estuary approximately 14,000 yr BP, and its present-day topset/foreset altitude of 247 feet (75.3 m) records the end of the 1000-year stable synglacial sea level that existed during ice retreat from Massachusetts into southeastern New Hampshire and southwestern Maine (Koteff and others, in press). Within the next 4.7 miles (5 km) northwest up the Cocheco River Valley, altitudes from several successive ice-marginal marine deltas show a sharply rising eustatic level of 20 - 30 feet (7 - 10 m). This exposure also represents the northernmost of 13 glaciomarine deltas that provided topset/foreset altitudes from which the uplift profile was derived, which shows uplift to have been 4.5

ft/mi (0.852 m/km) to the N 28.5° W in this region. In the Connecticut Valley to the west, an uplift profile derived from altitudes of 23 glaciolacustrine deltas indicated that the uplift was 4.7 ft/mi (0.889 m/km) to the N 21.5° W (Koteff and Larsen, 1989). Because of the similarity of slopes and closeness of fit of the altitude data, the delayed uplift model proposed for western New England is applicable to the coastal region. Other evidence for this proposal will be seen at later stops.

The Torr Pit delta has added significance because it demonstrates both the precision and accuracy of using topset/foreset altitudes to determine past glaciomarine levels. Wind action in the estuary was minimal because of its protected position, and there appears to have been no erosion of the delta by waves and reorientation of deposits that might alter the original topset/foreset contact. This has been a problem in examining deltas nearer the coast that were exposed to more open water of the Gulf of Maine, where wave-formed deposits, derived chiefly from the delta, mimic the original fluvial topset beds but are at a lower altitude. Also, tidal intertonguing of much less than a few feet (one meter) is indicated. Thin beds of subaqueous fine sand and silt, with landward-directed current structures, are interbedded with fluvial gravel of the topset beds that have seaward-directed current structures. At this locality, the topset beds are less than a meter thick, and the tidal intertongue penetrates only half that thickness. During late-glacial time, the tidal range for the Gulf of Maine was probably small, because at the time, the Gulf was connected to the Atlantic Ocean by a narrow channel at Georges Bank. Scott and Greenberg (1983) have modeled the Gulf of Maine tidal range for the last 7000 years, and have concluded that the range in this area, when Georges Bank was awash, was perhaps 20 - 25 per cent of the present-day range. A modern analogy is the Mediterranean Sea, which has small tidal fluctuation and is connected to the Atlantic by the Strait of Gibraltar.

Rollover beds (fluvial beds that can be traced into foreset beds) are present, indicating that the topset/foreset contact here is very accurate. It is assumed that most of the topset/foreset contact here was formed at the low water mark, as most contacts that formed at high tide were probably destroyed as the tide went out. Nevertheless, the accuracy of the former sea level altitude at this locality is still within a meter.

*Return to bus; all stops after this will be in Maine. Return to Route 11, turn left, return to Spaulding Turnpike, south to I-95. Take exit north to Maine. Exit I-95 at Route 236 West in Maine and proceed for approximately 4 miles to Eliot, Maine. Turn right at high school, proceed to Route 101, turn right and immediately turn left into pit for Great Hill.*

**STOP 2. GREAT HILL FILL AND GRAVEL, INC.** (40 minutes; Dover East quadrangle, top of landform from map elev. ca. 300 feet (90 m); leader Carl Koteff).

Great Hill has been visited by several field excursions (Mayewski and Birch, 1984; Weddle and others, 1990) in the past, attracted by the spectacular till exposure caused by excavation; it is one of the rare localities where a drumlin has been cut nearly in half. Although our trip is less concerned with till stratigraphy, it should be noted that most of the drumlin appears to be composed of the older of two identified tills in New England, probably of Illinoian age. Late Wisconsinan till overlies the older till, although the boundary is not sharply defined here because of the similarity in the texture and the lack of apparent weathering differences between the tills.

Great Hill is an excellent example of wave-modification by the late-glacial sea. The present east-west axial alignment is in contrast to the regional northwest-southeast drumlin trend caused by glaciation. It appears that wave and current action generated by winds from the south have eroded the till, which was an island during the marine invasion, and redeposited the reworked till as nearshore sand and gravel mostly around the south and east sides of the drumlin. Before excavation work, there probably was at least 196,000 yd<sup>3</sup> (150,000 m<sup>3</sup>) of wave-formed sand and gravel present, which is a conservative estimate, and does not include finer grain sizes of silt and clay. The altitude of the beach deposits derived from the drumlin is about 170 feet (52 m), which fits very closely on the uplift profile that defines the stable late-glacial sea level between 15,000 and 14,000 yr BP. The large volume and distribution of the reworked sand and gravel suggest that relative sea level at the time was relatively stable, at least for a period long enough to form the extensive nearshore deposits. Because the altitude of the beach here, as well as many other similar localities in coastal New Hampshire and Maine, is within 6 - 10

feet (2 - 3 m) of the synglacial sea level indicated by glacial marine deltas, it is felt that this represents not only a stable eustatic level, but helps to confirm a delayed postglacial uplift model as proposed here, and as developed for the Connecticut Valley to the west (Koteff and Larsen, 1989).

*Return to bus and return to Route 101, turn right and proceed to Route 236; turn right to South Berwick. Lunch stop will be along this route. Turn right onto Route 4 to North Berwick and turn right onto Route 9 East. Proceed for approximately 5 miles to left entrance into Wells Town Garage property, and into pit area behind buildings.*

**STOP 3. MERRILAND RIDGE, WELLS TOWN PIT** (40 minutes; North Berwick quadrangle; elev. ca. 200 feet (60 m); Carl Koteff leader).

Exposures in this pit show 12 - 15 feet (4 - 5 m) of cobble to boulder gravel with depositional current directions northward unconformably overlying 15 - 20 feet (5 - 7 m) of sand and sand and gravel foreset beds dipping southward. In the past, evidence of ice shove in a southerly direction has been observed on the ice-proximal north side of the ridge that showed marine-bottom sand and silt thrust over the foreset beds. These also are unconformably overlain by the cobble to boulder gravel. Our interpretation is that Merriland Ridge was a glaciomarine delta formed at an east-west ice margin that retreated a short distance northward before the ice readvanced over at least the proximal end of the delta. Subsequent wave action caused by southerly winds in the late-glacial Gulf of Maine after further ice retreat reworked morainal material and any remaining topset beds forming nearshore deposits that were laid down in a northerly direction. The original topset/foreset contact appears to have been completely destroyed, and we do not know whether this delta was deposited in the earlier and lower sea level or in a slightly later and sharply higher level. Only a few km of ice retreat occurred between deposition of deltas at the two distinct late-glacial sea levels in this area, but the overlying wave-formed deposits appear to have been constructed at the higher level. Although the small readvance indicates a slight change from systematic ice retreat, it is believed that interval was short. On the other hand, the extensive development and distribution of the wave-formed deposits suggest that sea level stabilized once more after the sharp rise from the lower level.

*Return to bus and exit from town garage property and turn right onto Route 9; turn immediate right (west) on Swamp John Road for approximately one mile. Turn left onto Bragdon Road and proceed for approximately 0.5 miles to right-turn entrance into Chase Pit.*

**STOP 4. CHASE PIT** (30 minutes; North Berwick quadrangle; elev. ca. 230 feet (70 m); leader Carl Koteff).

This delta is the southernmost delta found in Maine that was graded to the higher synglacial sea level. With a topset/foreset contact altitude of 226 feet (68.9 m), it is about 20 feet (7 m) higher than the projected lower sea level. Contrary to the evidence in the Cocheco River valley in New Hampshire, no glaciomarine deltas constructed at intermediate steps from the low to higher late glacial sea level are known in Maine, although a few are known to have been deposited during the lower level. One of the lower-level deltas (now completely excavated) was located 4.3 miles (7 km) southwest of here, but only about 2.5 - 3 miles (4 - 5 km) south of here in terms of ice retreat distance. The Bragdon Road delta has an esker feeder at its north end and the ice-contact slopes there indicate that the ice margin in this area trended east-west. Topset beds are several m thick at the ice-proximal end, and thin to only a meter or so within a short distance southward where they appear to intertongue with intertidal sand that has northerly depositional current directions. Wave-formed cobble gravel overlaps both topset and foreset beds at the southern edge of the exposure. The Bragdon Road delta thus is interpreted as an ice-marginal feature deposited into the late-glacial sea immediately after a sharp and rapid rise in eustatic level of at least 20 feet (7 m); it then was affected by wave action after further ice retreat but sea level appears to have stabilized. Topset/foreset contacts of several deltas north of this locality indicate that sea level may have risen only a meter or two during ice retreat of approximately 12.4 miles (20 km). During all of this time during ice retreat from the Boston area into southeastern New Hampshire and southwestern Maine, postglacial uplift still was delayed.

Return to bus; turn left from Chase Pit entrance and proceed approximately 2 miles; turn left onto Route 109 and proceed about 9 miles to Route 4. Travel approximately 4 miles to Route 111; turn right onto Route 111 and proceed for about 5 miles to right-turn entrance of Spang Mills Pit.

**STOP 5 (ALTERNATIVE STOP). SPANG MILLS PIT** (30 minutes; Alfred quadrangle; elev ca. 270 feet (82 m); leader Carl Koteff).

Cobble and pebble gravel with interlayered sand beds at the north side of this exposure grade to sand at the south side over a distance of about 300 feet (100 m). Marine clayey silt and silty clay of the Presumpscot Formation occurs to the south, and is presumed to interfinger with the sand, although this is not exposed at present. Most of the material is in foreset beds that dip south to southeast and are slightly convex; nearly every foreset bed can be traced from the approximate upper meter of cobble to pebble gravel, whose surface slopes southward more than 30 feet (10 m) over a distance of about 600 feet (200 m). This feature is interpreted to be a subaqueous fan, deposited from a tunnel at the base of the ice margin into the late-glacial sea. The surface of this fan is very close to the projected higher late-glacial sea level, about 270 feet (82 m) altitude in this area, but no topset beds were observed. An exposure across Route 111 north of this stop appears to be in a delta that does have topset/foreset beds, but it is not clear if the fan and delta are contemporaneous, or that the two features represent successive ice margins.

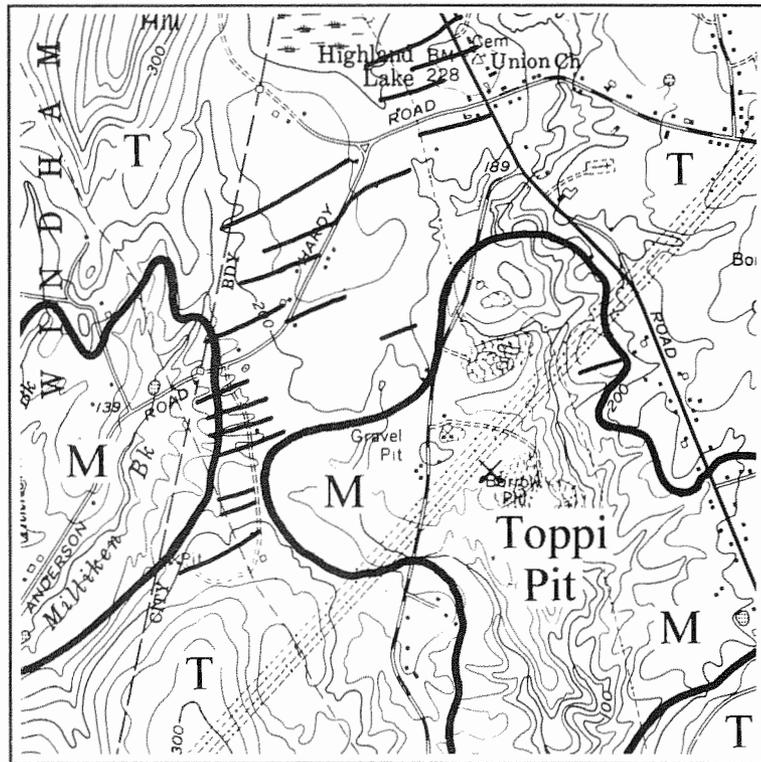
Unfortunately, as of this writing, we have not obtained an altitude of this topset/foreset contact.

Return to bus; leave pit and turn left onto Rout 111 to I-95. After about 5 miles, access I-95 North at Interchange 4 and proceed to Interchange 7; exit west to Route 25 to Westbrook. Approximately 0.75 miles bear right onto Cumberland Street and cross Presumpscot River. Turn right after about 0.5 miles onto Methodist Road; continue north for about 3.0 miles passing under power lines; enter Toppi gravel pit on right.

**STOP 6. TOPPI PIT** (60 minutes; Portland West quadrangle; elev. ca. 200 feet (60 m); leader W. B. Thompson).

This stop is located west of Mill Brook, in a small north-south valley (Figure 5). The valley has been filled with glaciomarine sediments, while the higher ground on either side is underlain by till.

The elevation of the upper marine limit in this area during deglaciation was approximately 285 feet (87 m) (Thompson and others, 1989). Numerous end moraines occur just north and west of the Toppi Pit. Exposures at this stop and other nearby pits indicate that many other moraines are buried beneath the marine sediments along Mill Brook. The dissected upper surface of this valley-fill (the former sea floor) can be seen along the power line as we drive into the Toppi Pit. The features to be examined at this stop include (1) a large expanse of glacially abraded bedrock, and (2) a sequence of glacial and glaciomarine deposits that record the recession of the marine-based late Wisconsinan ice margin from the Mill Brook Valley.



**Figure 5.** Stop 6 (Day 1), Toppi Pit; T (till), M (glacial marine sediments), lines (end moraines); (Portland West 7.5' quadrangle; geology after Thompson, 1976).

Five major units (including bedrock) are exposed in the Toppi Pit. When the site was documented for this guidebook (November, 1992), the eastern part of the pit was being operated on four levels, each exposing 10 - 15 feet (3 - 5 m) of section. The working faces trended north to east-northeast across the pit, and showed oblique cross-sections through two end moraines. These moraines were situated on the striated bedrock surface, and were overlain by a sequence of submarine fan deposits and glaciomarine mud. The exposures will have changed greatly by the time of this trip, but the basic stratigraphy is the same, and is typical of end moraine/submarine fan associations formed during the deglaciation of coastal Maine. The units in the Toppi Pit are described below, from oldest to youngest.

**Unit 1 - bedrock.** The bedrock exposed on the pit floor is chiefly granite and granite pegmatite. The outcrops consist of a series of rock knobs, on which the stoss sides are prominently striated and polished, while the lee sides commonly show a smooth, scalloped surface that presumably resulted from subglacial meltwater abrasion. The striations vary only slightly in orientation across the pit floor, and have an average trend of 167°. This trend is consistent with the east-northeast orientation of local end moraines. Crescentic marks are associated with the striations. Some of these fractures are convex in the direction of glacial flow, while others are concave. An unusual "channel" trends southwest across the bedrock surface near the pit entrance ramp. The origin of this feature is uncertain and may have been a composite of several processes. Some parts of the channel walls are meltwater-abraded, while others are striated or appear to have been plucked along bedrock joints.

**Unit 2 - end moraine deposits.** The two buried moraines exposed by pit operations in 1992 trend E-W to ENE-WSW across the bedrock surface. (Additional moraine deposits have since been exposed elsewhere in the pit complex). The southern moraine varies in composition from massive glacial diamict (till) with sand laminae, to a complexly sheared and interlayered mixture of till, sand, and gravel. The diamict lithofacies is olive-gray, sandy, stony, non-fissile, and moderately compact. It may include both flow till and lodgement till. The till contains stones up to 10 feet (3 m) in diameter, some of which are faceted and striated. The layered parts of the moraine resulted from minor glacial readvance and mixing of till with proximal submarine fan deposits, in the manner described by Smith and Hunter (1989). This tectonic layering dips toward the proximal side of the moraine, and together with thrust faults and recumbent folds, indicates ice shove from the north. The northern moraine consists chiefly of sand and gravel with a few diamict lenses. These sediments were deposited at the ice margin as proximal submarine fans, and are further described under Unit 3 below. The trough between the two moraine crests is conformably filled by laminated silt and clay of Unit 5.

**Unit 3 - proximal submarine fan deposits.** There is overlap in the depositional environments of Units 2 and 3, and the distinction between them is not always clear in the field. Proximal fan deposits form a major component of some moraines in the area, but they also can occur independently. These fan deposits were emplaced where high-energy meltwater streams emerged from the mouths of subglacial tunnels at the glacier margin. In the Toppi Pit, they are composed chiefly of massive to well stratified pebble-cobble gravel and sand. Bedding dips at shallow to moderate angles, and locally exhibits folding and reverse faults due to ice-shove. The coarsest fan gravel in the northern moraine (mentioned above) seems to occur as mound-shaped concentrations in the lower part of the section. These gravels may be tunnel-mouth deposits. The gravel fraction in this moraine is poorly sorted, massive to weakly stratified, mostly angular to subangular, and clast-supported. The proximal fan sediments higher in the moraine are stratified to a greater degree, with lenses of planar-bedded sand. Contacts between Units 3 and 4 (distal fan deposits) range from gradational to sharp and unconformable.

**Unit 4 - distal submarine fan deposits.** This unit is generally quite thin, less than 10 feet (3 m), and is not present in all parts of the pit area. It marks a transition between Units 3 and 5 as ice retreat resulted in lower-energy sedimentation from density underflows and settling of mud from overflow plumes. Distal fan deposits in the Toppi Pit consist of well-stratified, interbedded sand, silt, and clay. They show sub-horizontal to gently-dipping planar beds, locally offset by normal faults and convoluted by water-escape structures. Channel-shaped unconformities occur at the base of this unit, within the unit itself, and even along the overlying contact with Unit 5. The channel-fills within the basal and internal channels locally include rip-up clasts of clay derived from erosion of the channel walls. Some of the sand beds in the channels are normally graded; others appear massive. These channel features are inferred to have resulted from submarine slumps and scouring by turbidity currents. Contacts between Units 4 and 5 range from gradational to sharp.

**Unit 5 - glaciomarine mud.** This unit consists of laminated silt, clay, and minor very fine sand. It was deposited in a quiet-water environment on the sea floor, and is part of the regionally extensive Presumpscot Formation. The observed thickness along the north wall of the Toppi Pit in 1992 was about 20 feet (6 m). *Portlandia arctica* shells recently collected in this part of the pit have a radiocarbon age of  $10,375 \pm 80$  yr BP (AA-10159). This date is surprisingly young, since the *Portlandia* shells are typical of ice-marginal environments in coastal Maine and often yield dates older than 13,000 yr BP. However, other data from the radiocarbon lab suggest that the date from this locality is erroneous, and it is planned to re-date the shells.

*Return to bus and backtrack to Interchange 7; access Maine Turnpike North to I-95 (Exit 9). Continue on I-95 North to Exit 17 and access Route 1 North. Go approximately 0.5 miles, cross Cousins River (Muddy Rudder Restaurant on right) and turn right into Freeport Inn parking lot. End of Day 1 excursion.*

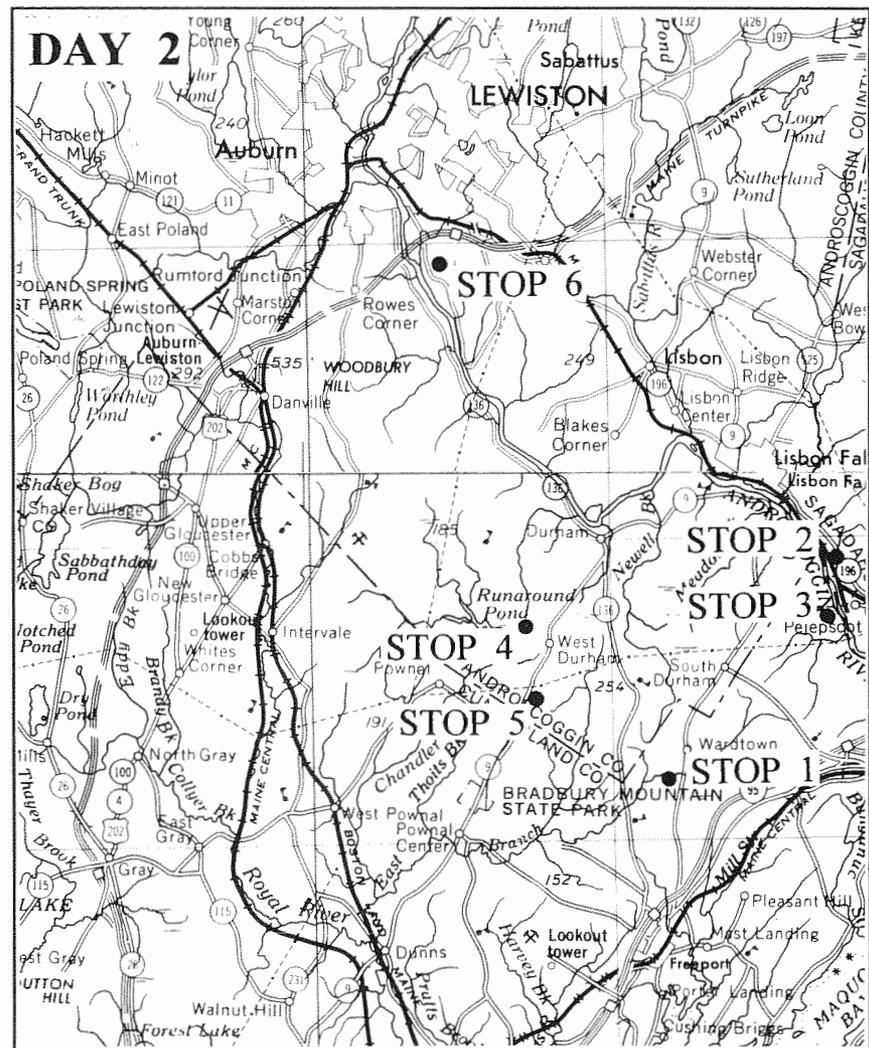
### ROAD LOG - DAY 2 (10/23/93)

Refer to locality map (Figure 6) for directions and approximate distances of travel. We will examine glacial marine fans and deltas, stratified end moraines, and nearshore deposits associated with postglacial uplift in the lower Androscoggin River valley.

Assemble in parking lot of Freeport Inn for 8:00AM departure. Access I-95 N and proceed to Exit 20 (Freeport/Durham; west on Routes 125 and 136); turn right at end of ramp, cross over I-95, turn right at stop sign. Approximately 0.75 miles turn right at fork onto Route 125, remain on main road approximately 2.0 miles; turn left onto Curtis Road just past Florida Lake Campground. Turn right after about 0.25 miles into gravel pit and drive to north end of pit.

**STOP 1. SEYMOUR PIT** (Park at north end of pit; 40 minutes; Lisbon Falls South quadrangle; elev. ca. 200 feet (60 m); leader T. K. Weddle).

This stop is in an ice-marginal moraine-fan complex, in which evidence for ice-shove and ice-contact collapse is well exposed. At the north end of the pit, ice-contact deposits are overlain by southward-dipping planar-

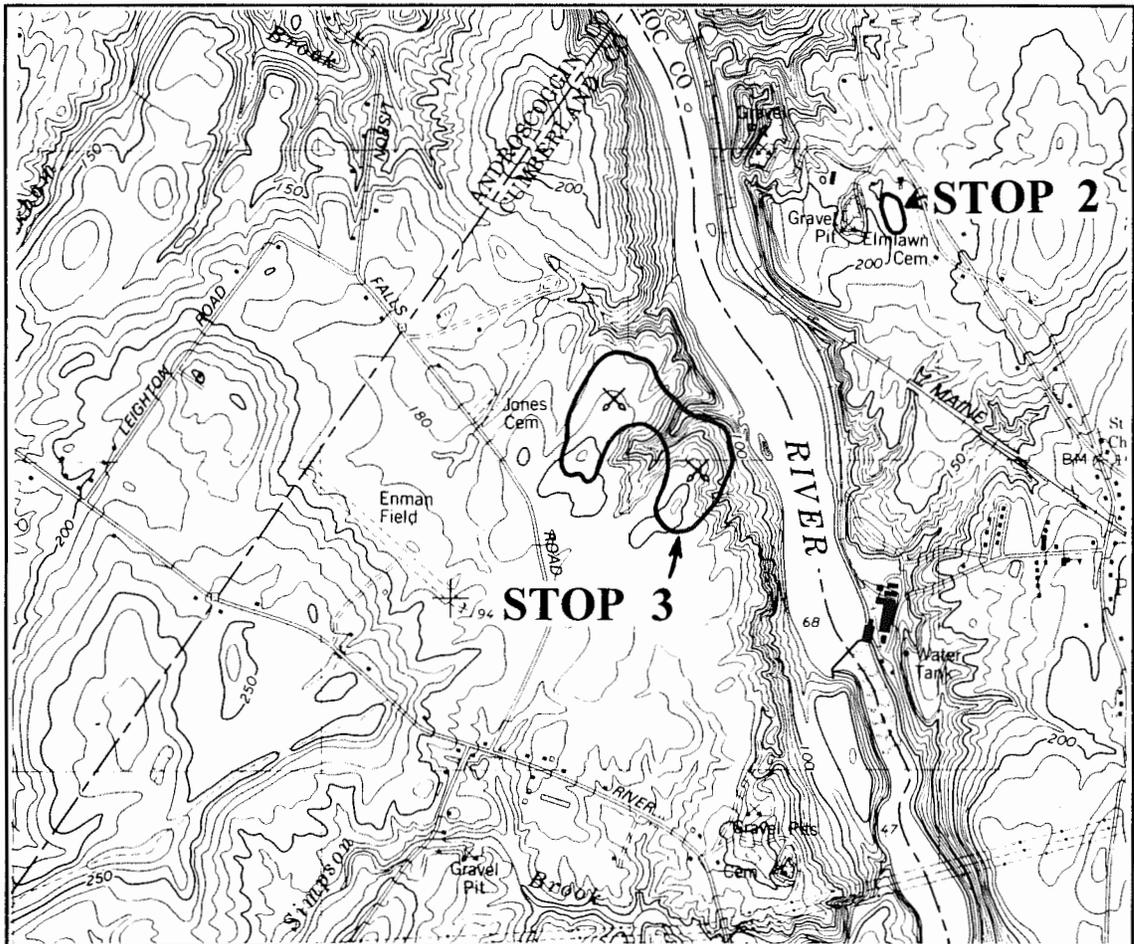


**Figure 6.** Stop locality map, Day 2 (Portland and Lewiston 1:250,000 sheets).

and cross-bedded medium- to fine-grained sand, and glacial marine mud; ripple-drift cross-laminated sand also is present. Steeply-dipping stratified sandy diamicton in the ice-contact deposits marks the ice margin on the north side of the east-west trending ridge. Articulated *Portlandia arctica* from slumped glacial marine mud over ice-contact fan deposits in the north end of the pit have dated at  $14,045 \pm 95$  yr BP (AA10164). The *Portlandia* are the only shells found here, and are very small ( $< 1.0$  cm). As the ice margin stood at this position, it readvanced several times in a series of minor pulses, deforming and stacking the fan deposits that make up the Curtis Road ridge. In the central part of the pit, ice-shove deformed fan deposits are overlain by undeformed fan deposits; the thrust faults trend near  $90^\circ$  and dip northwest, and are associated with northwest-dipping diamicton bodies, apparently deposited during minor readvance of the ice. Striations ( $175^\circ$ ) nearly perpendicular to the thrust faults are present on exposed bedrock in the south end of the pit.

Return to bus, retrace route, and access I-95 N and proceed to Exit 24 (Topsham/Lewiston); turn left at end of exit ramp (west) on Route 196 to Lewiston. Approximately 1.5 miles cross under power lines; continue ca. 1.5 miles, observe small cemetery on left (Elmlawn Cemetery) and turn left at dirt road immediately after cemetery.

**STOP 2. MAINE DEPARTMENT OF TRANSPORTATION PIT** (turn left after entrance and follow road to end of pit exposure on right; 30 minutes; Lisbon Falls South quadrangle; elev. ca. 200 feet (60 m); leader T. K. Weddle) (Figure 7).



**Figure 7.** Locations of stops 2 and 3 (Day 2); northeast trending moraine visible in lower left corner of map is on trend with fans at stops 2 and 3 (Lisbon Falls South 7.5 quadrangle).

The purpose of this stop is to see a Pleistocene nearshore deposit associated with regression of the glacial sea after uplift was underway. The deposit overlies glacial marine mud which rests on distal fan deposits of interbedded silt and fine sand. The larger exposure in the core of the fan immediately adjacent to the north shows the coarse nature of the proximal fan deposits associated with the ice tunnel from which the fan deposits originated. The flanks of the fan are mantled by massive glacial marine mud from which the pit operator claims fossil shells were found. Bloom (1960) reported a varied assemblage of fossil shells from this pit including *Hiatella arctica*, *Macoma calcarea*, *Musculus substriata*, *Mya arenaria*, *Mytilus edulis*, *Nuculana jacksoni*, *Serripes groenlandicus*, *Natica clausa*, *Neptunea decemcostata*, and *Balanus*, and represent mixed intertidal and deeper-water affinities, reflective of the depositional environments associated with the sediments in the pit. Molds of pelecypods have been found in the mud under the nearshore deposit but fossils have not been recently observed elsewhere in the pit.

*Return to bus, return to Route 196 and turn left. Continue west on Route 196 to Lisbon Falls and turn left at lights (Routes 125 and 9). Follow road to left, cross Androscoggin River, and turn left at end of bridge. Proceed on road about 1.0 miles and turn left at Shiloh Road. Follow road about 2.0 miles (past Town of Brunswick Graham Road Landfill) and turn left at Labbe gravel pit entrance; proceed into pit.*

**STOP 3. LABBE PIT** (follow road to lower levels; bear left and park at bottom of pit; walk to top of landform; 30 minutes; Lisbon Falls South quadrangle; elev. ca. 260 feet (78 m); leader T. K. Weddle).

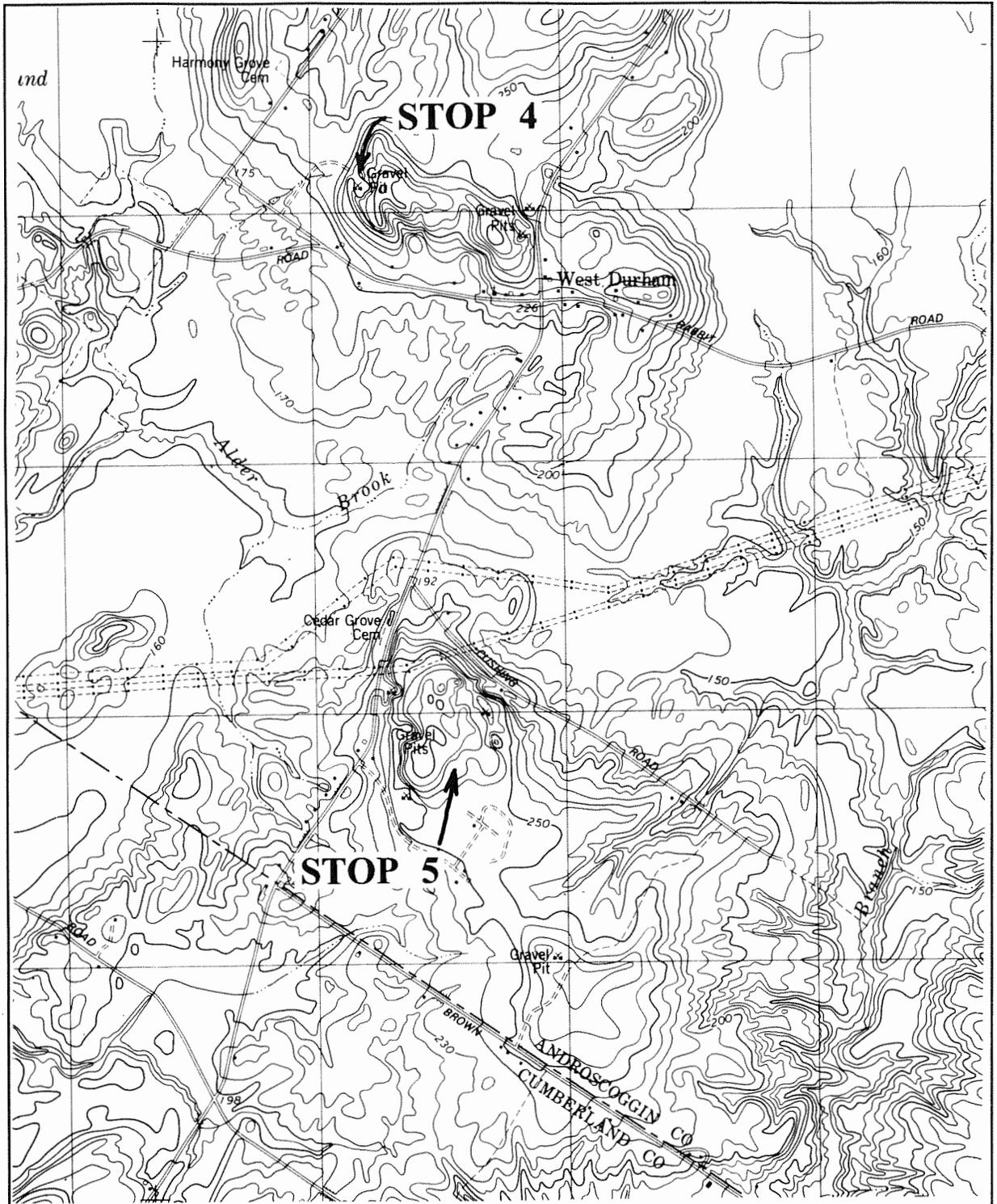
The Labbe pit is excavated into a large fan that is anchored to the east end of a large grounding-line moraine (Figure 7). The fan deposits have been deformed by ice-shove and contain low-angle thrust faults and drag folds. Sandy till is interbedded with stratified drift and is associated with the deformation, however, the till is found in tabular and wedge-shaped bodies. Some of the till may be lodgement till, whereas in places it may represent slumped material or flow till. The lower exposure where the vehicles are parked shows that the fan deposits have been truncated, either by marine regressive processes or by the early post-glacial Androscoggin River. Undated marine fossils have been found in the mostly reclaimed pit to the north, including *Hiatella arctica*, *Nucula expansa*, and *Nuculana*.

*Return to bus and out to pit entrance; turn left, proceed to end of road. Turn right and immediate left across main road; proceed to end of road and turn right with caution. Proceed to intersection of Route 125 (about 0.75 miles), cross Route 125 with caution, and continue along road under power lines. Bear left at fork approximately 0.5 miles from power lines and proceed to intersection of Route 136; cross Route 136 with caution. Proceed to intersection of Route 9; cross Route 9 with caution and continue west past gravel pits. Bear right at fork after pits; stop about 0.1 mile and park at Scott Dugas P&S pit sign.*

**STOP 4. DUGAS, HANNA, AND LARABEE PITS** (walk into pit along the left up onto landform; **lunch stop** and walk to overview of different pits (75 minutes); North Pownal quadrangle; elev. asl ca. 300 feet (90 m); leader C. L. Marvinney) (Figure 8).

The above gravel pits are excavated into coalescing glacial marine fans which form the western edge of a large grounding-line moraine. This moraine, although not continuous, can be traced 4.5 miles (7.2 km) eastward to the Labbe gravel pit of Stop 3. The Dugas pit, through which we enter, is excavated into southward-dipping foreset beds of a fan. Limited exposure of the foreset beds occurs in the eastern wall of the pit. The foreset material is fine to coarse grained sand, with pebble to cobble sized clasts and occasional layering. Interbedded silt, clay, and flow till are present in the western portion of the pit. Unconformably overlying all material is a massive clast-rich unit 3 - 4 feet (1 m) in thickness. Clasts which are sub-rounded to rounded range in size from large cobbles to small boulders. This unit represents a reworked deposit formed during the marine regression. In places, there appears to be fluvial-like deposits below the reworked sediments, suggestive of topset bedding. The top of the landform is just over 300 feet (90 m), close to the marine limit in the area, and there is a small depression (kettle?) on the landform surface. However, the evidence that this feature is a delta is inconclusive.

The Hanna pit is in a fan just east of the Dugas pit. Here, too, uniformly dipping foreset beds are well exposed. As in the adjacent fan, bed grain-size ranges from fine- to coarse-grained sand with local pebbles and



**Figure 8.** Stops 4 and 5 (Day 2); three large landforms with gravel pits in them are glacial marine fans constructed at successive northward-retreating ice-marginal positions. The two southerly fans are connected by a spit which formed during the marine regression; the northerly fan complex (Stop 4) is anchored to the west end of the grounding-line moraine in Figure 7 (North Pownal 7.5' quadrangle).

pebble gravels. Marine mud is found in the central portion of the pit. Here the deposits have been deformed by ice shove. Large open folds are seen in the upper clay and sand interface. As in the Dugas pit, a 3 - 4 feet (1 m) thick clast-rich layer associated with the marine regression unconformably overlies all units. In places, a fine uniform sand forms the uppermost regressive unit.

The Larabee pit is excavated into the proximal side of the Hanna fan complex. A portion of the pit contains the same features discussed above. However, this pit also contains exposures of ice-proximal features, slumping, flow tills, faults, and drastic changes in stratigraphy.

The exposures represent a stillstand of the ice margin that can be traced on the topographic maps from Stop 3 to this stop. At the top of the landform and looking southward, an earlier ice-marginal deposit is visible, built when the ice margin was to the south. The next stop will be at that landform, which is another fan built to near the synglacial sea level for this area. Although these two exposures are not conclusively interpreted as deltas, there are deltas approximately 10 miles (16 km) to the north which do not have reworked sediments on their tops. This is supportive evidence that uplift was underway when those deltas were deposited, and possibly had begun by the time when the fans seen at Stops 4 and 5 were being deposited.

*Return to bus; retrace route back to Route 9 and turn right (south). After 0.75 miles pass under power lines, go past graveyard on right, and pass under second set of power lines; gravel pit visible on left. Access to pit on left; proceed just beyond on Route 9 to sign for campground on left and enter here.*

**STOP 5. BLACKSTONE PIT** (turn left at campground sign; proceed to end of access road to working face; 40 minutes; North Pownal quadrangle; elev. ca. 300 feet (90 m); leader C. L. Marvinney).

The Blackstone gravel pit is located within a large glacial marine fan. Although the elevation of the top of the feature is just over 300 feet (90 m) (Figure 8), very close to the marine limit in the area, no deltaic topset/foreset contact is visible in the exposure. The deposit was formed during the retreat of the late Wisconsinan ice sheet at the ice margin. Foreset beds dominate the exposure, best seen in the central portion of the pit where they are uniformly dipping to the southwest. Presumpscot Formation mud interfingers with sand beds along the southern, oceanward portion of the pit. On cleared faces, small-scale syndepositional faults and dewatering structures are present. Proximal and intermediate to distal facies of the glacial marine fan are represented by the coarse and fine sand and the interbedded fine sand and mud, respectively. In the northern end of the pit, a reworked sandy deposit overlies the sandy foreset beds, and represents wave reworking of the fan at the marine limit. A clast-rich unit of massive character unconformably overlies the fan and reworked beds, and is itself a reworked deposit formed during the marine regression. This unit is traceable down the topographic slope of the fan and consists of small to medium sized boulders within a matrix of fine sand to gravel. Although no apparent bedding is visible within the coarse-grained material, pockets of fine-grained material do contain some cross-bedding. A massive, uniformly bedded sand also associated with marine regression caps the deposit.

*Return to the bus and exit pit; turn right (north) on Route 9 and continue approximately 2 miles to intersection with Route 136. Turn left and continue on Route 136 as it parallels the west bank of the Androscoggin River. After approximately 8 miles (and after passing under the Maine Turnpike) approach Auburn city limits. Bear right (roller skating rink on right) and at T-intersection turn right and onto bridge over Androscoggin River; take second right turn after bridge onto River Road. After approximately 1.8 miles take left forl (Goddard Road) and continue up hill. Dube gravel pit entrance located approximately 0.5 miles on right.*

**STOP 6. DUBE, PIKE, AND LEWISTON PITS** (75 minutes; Lewiston quadrangle; elev. ca. 330 feet (100 m); leader M. J. Retelle)

At this stop, we will look at exposures in several active pits located in a complex landform that occupies an area of approximately 0.4 mi<sup>2</sup> (1.0 km<sup>2</sup>), east of the Androscoggin River in Lewiston. Because of extensive excavations, the original morphology of this feature has been greatly altered, however, examination of the 1908 U. S. Geological Survey 15-minute topographic map shows this feature originally had an extensive flat upper surface, the remnants of which form the current border of the Dube pit with the City of Lewiston pit. The

elevation of the upper flat surface is approximately 330 feet (100 m), close to what is interpreted as marine limit for this area (Thompson and others, 1989). Based on original morphology, the landform is interpreted as a glacial marine delta, however, prior to discovery of the 1908 map, the landform was interpreted as a glacial marine fan complex (Slayton, 1993). We now interpret this landform as an ice-contact glacial marine delta that grew from a fan, which aggraded to sea level and prograded southward into the marine basin. The purpose of visiting this site is to examine sequences that illustrate the development of the landform from early ice-contact facies through fan and delta facies to eventual shallow marine sedimentation during the postglacial marine regression. We will start in the Dube pit (north) and walk to the southern exposures and be picked up by the bus at the southern end of the traverse.

The Dube pit is the northernmost operating gravel pit in the landform and exposes mostly ice-proximal facies in the deposit. The pit is bordered by a rock quarry to the north; the west wall of the pit is a steep ice-contact face. The landform narrows to the north probably reflecting a wave-modified and smoothed ice-tunnel deposit (Ashley and others, 1991). The north wall of the pit exposes a cross-section through the ice tunnel or conduit facies that fed the fan complex. Coarse clast-supported boulder gravel is exposed in several faces and is interpreted as bedload in the tunnel. Overlying and adjacent fine sandy gravel and sand foresets may be early stages of the submarine fan development. A distinctive unconformity separates the gravelly ice-contact facies from the fine-grained marine mud facies that draped the landform after ice retreated from the feeder tunnel position. Above the unconformity, several meters of laminated and massive fine-grained marine mud is exposed, which in places is fossiliferous, containing numerous paired and disarticulated mollusc shells including *Hiatella arctica*, *Portlandia arctica*, *Mya truncata*, and *Mytilus edulis*, as well as *Balanus* plates. One sample of *Portlandia* shells recently yielded an age of  $12,890 \pm 85$  yr BP (AA10165), and may be interpreted as dating the ice margin at this site. There is, however, evidence of reworking in the deposits as the fossils at this site are found in several contexts, including both *in situ* in the massive mud, but more commonly in a coarse-grained diamicton that is exposed between the laminated mud and the east wall of the pit. The diamicton is interpreted as a debris flow deposited down the dipping foreset beds of the delta. The marine mud is capped by sandy and muddy offlap sediments whose structures mirror the underlying topography of the landform.

There are several exposures in the central and southern part of the landform (Lewiston and Pike pits) that nicely illustrate the sedimentary regimes in the medial and distal portions of the complex. A north-south oriented wall in the central area of the pit exposes thin south-dipping sandy foreset beds, comprised predominantly of normally graded and ungraded beds with a few 4 - 8 inch (10 - 20 cm) thick beds of ripple-drift cross laminated sand. The graded sandy beds are interpreted as grain flow deposits originating upslope on the delta foresets. Some sandy beds contain rounded clasts of diamicton (till balls) up to 2 inches (5 cm) in diameter. Syndepositional extensional faulting is also seen in translocations of the sandy strata subparallel to bedding.

The distal fan facies is well exposed near the southwestern corner of the landform near the top of a fining-upward sequence with rhythmically laminated sand at the base, overlain by sand-mud interlayers, laminated mud, and massive sand. At several locations in the section, the fine-grained marine mud deposits are interrupted by thin layers of matrix-supported diamicton with a sandy, muddy matrix and stream-rounded gravelly clasts. These units are interpreted as debris flow deposits that originated high on the foreset beds of the landform and incorporated sand and mud as the flows travelled basinward.

*Return to bus and leave pit to River Road and turn left (south); proceed about 3 miles and take right (Cotton Road) and continue about 2 miles; turn left onto Pine Woods Road. Go approximately 2.5 miles to Lisbon center; junction with Route 196 (at lights) and turn right (south); continue on Route 196 about 8.5 miles to I-95. Access I-95 south (Interchange 24) and return to Freeport Inn (about 13 miles). Take Exit 17 and follow signs to access Route 1 north; cross Cousins River and turn into Freeport Inn parking lot. End of Day 2 excursion.*

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## ROAD LOG - DAY 3 (10/24/93).

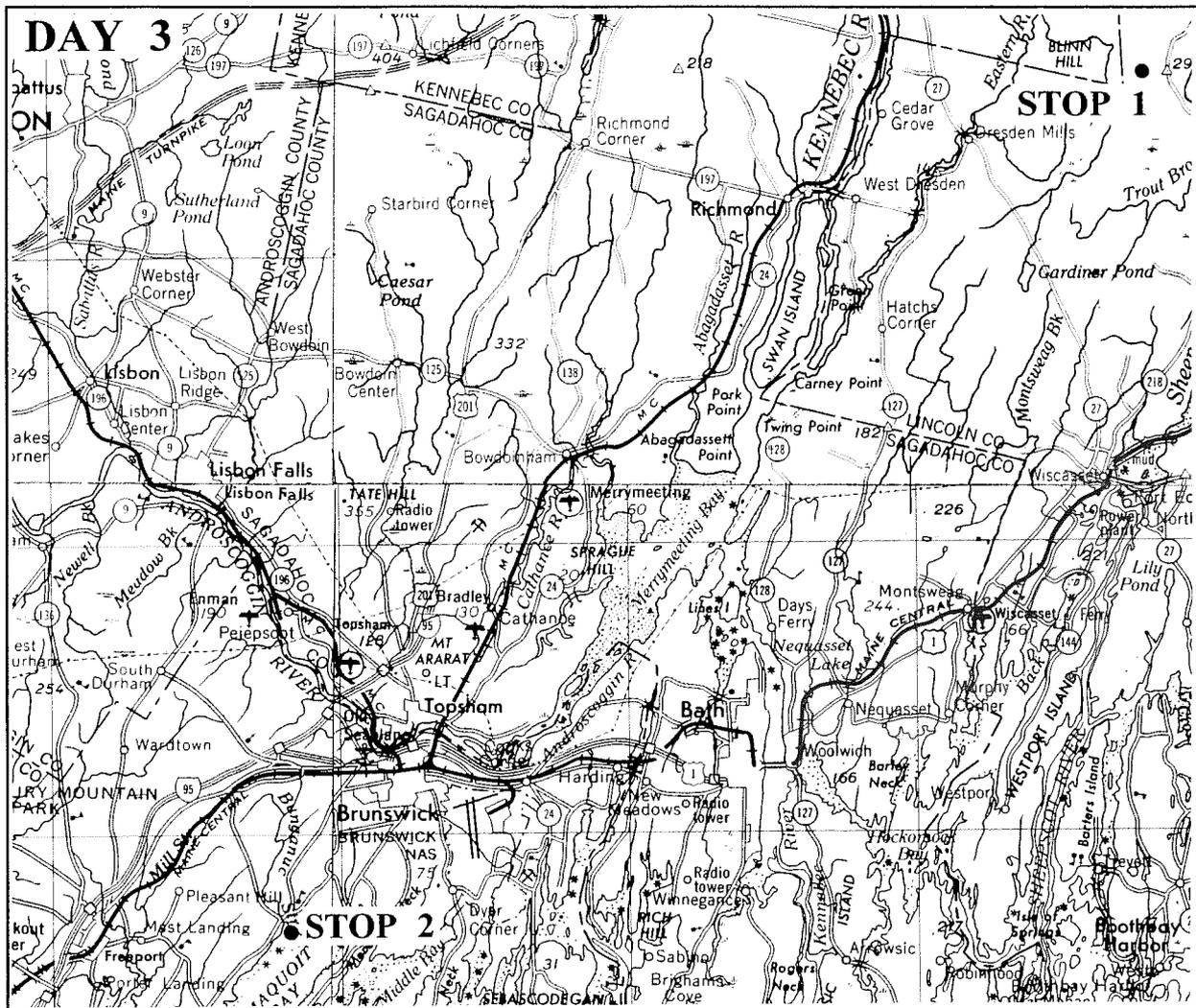


Figure 9. Stop locality map, Day 3 (Augusta, Bath, Lewiston, Portland 1:250,000 sheets).

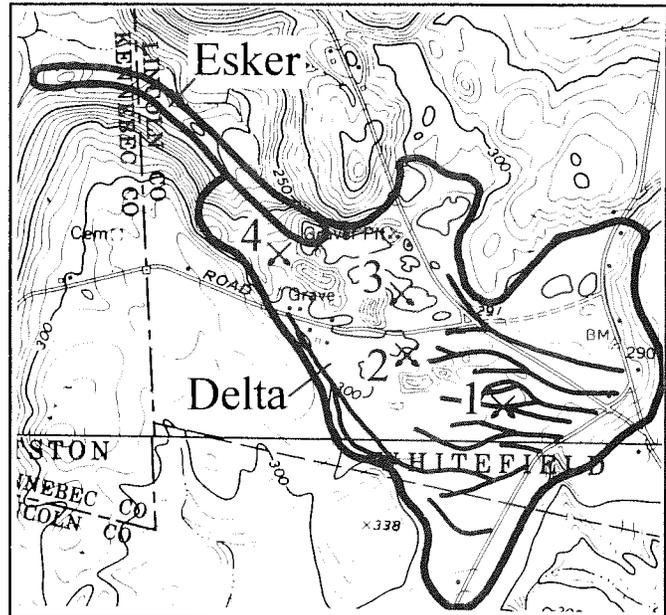
Refer to locality map (Figure 9) for directions and approximate distances of travel. In the morning, we will visit a glacial marine delta north of Freeport deposited after uplift was underway. Returning south, a final stop (and lunch) to see a thick sequence of glacial marine mud will complete the trip. The bus will then return to Boston arriving no later than 4:30PM.

Assemble in parking lot of Freeport Inn for 8:00AM departure. Access I-95 North and proceed to Exit 22 (Brunswick/Bath) to access Route 1; continue east through Brunswick to Bath. Cross Carlton Bridge over Kennebec River in Bath (Bath Iron Works Shipyard on right). Continue on Route 1 east to Wiscasset; in Wiscasset center turn left onto Route 218 (immediately before crossing bridge over Sheepscot River). Continue north on Route 218 about 8.5 miles to the village of Head Tide; about 1 mile north of Head Tide on Route 218, turn left onto Thayer Road (unpaved) and continue straight 0.5 miles where gravel pit is on left; proceed through gate.

**STOP 1. PALMER HILL DELTA** (East Pittston and Wiscasset quadrangles; elev. ca. 300 feet (90 m); leader W. B. Thompson).

The Palmer Hill Delta in Whitefield is an excellent example of an esker-fed ice-contact glaciomarine delta. It is one of the most southerly deltas in mid-coastal Maine, and thus important for defining late-glacial sea level in this part of the state. It also shows a complete assemblage of the morphologic and stratigraphic features that are typical of glaciomarine deltas throughout southern Maine (Thompson and others, 1989). We will visit one or more of the following gravel pits in this delta, depending on available time and exposures: Crooker Pit, Hanley Pit, and Rice Pits (Figure 10). Together,

these pits show a series of exposures from the collapsed ice-contact head of the delta to the distal part of the delta plain. The Palmer Hill Delta is a typical Gilbert-type delta, with topset and foreset beds. It formed approximately 13,500 years ago, during recession of the late Wisconsinan ice margin. A striated pavement outcrop recently exposed on the east base of the delta front indicates an ice-flow direction of  $150^{\circ}$  in this area. Thompson and others (1989) surveyed the elevation of the topset-foreset contact in the Potter Pit, and found that the position of former sea level recorded by the delta is presently at 291 ft (89 m). The delta was deposited where the tidewater glacier margin was pinned against Palmer Hill to the northeast and the unnamed hill to the southwest. It is inferred from bedrock outcrops and subsurface data that the water depth during deposition was generally less than 100 ft (30 m). The Sheepscot River valley in front of the delta is occupied by stratified end moraines, which are overlain by glaciomarine silt and clay of the Presumpscot Formation. The subglacial stream that deposited the adjacent esker (Figure 10) supplied all or most of the sediment in the delta. The crest of the preserved part of the esker ridge (in woods northwest of Rice Pit) is 100 ft (30 m) lower than the head of the delta plain. This suggests that englacial hydrostatic pressure elevated the esker conduit to the delta head, where meltwater and sediment discharged from one or more fountains. An unusual ball of red-brown clay, 3 ft (1 m) in diameter, was found in the topset beds in the Crooker Pit. It consists of kaolinitic saprolite that is thought to have been subglacially eroded from a valley northwest of the delta and washed through the esker system to the delta surface.



**Figure 10.** Palmer Hill delta and associated esker (Stop 1, Day 3); 1-Crooker pit, 2-Hanley pit, 3-Potter pit, 4-Rice pit. Lines indicate meltwater channels on delta plain, flow to east (East Pittston and Wiscasset 7.5' quadrangles).

**Crooker Pit.** The Crooker Pit is located in the distal part of the Palmer Hill delta and provides the best exposures of the topset and foreset beds. The topsets consist of up to 10 ft (3 m) of poorly sorted pebble to cobble gravel with minor boulder gravel and sand lenses. This gravel is massive to moderately well stratified, and ranges from matrix-supported to clast-supported. It shows sub-horizontal bedding and local fluvial cross-bedding in channels. The underlying foreset beds are composed of well-stratified, variably sorted sand and gravel. They dip in various directions, but usually between northeast and south, at angles as steep as  $33^{\circ}$ . The foresets are planar-bedded, with individual beds ranging in thickness from a few inches to 1.5 feet (0.5 m) or more. They may be either clast-supported or matrix-supported. Deposition of the foresets occurred through a

combination of grain flows and sediment gravity flows. Slump structures resulting from avalanching on the delta front are occasionally exposed in the pit face. Meltwater channels 3 - 10 ft (1 - 3 m) deep can be seen in the blueberry fields next to the Crooker Pit. On the seaward edge of the delta plain, the channels are truncated by a low beach ridge. The preservation of the channels and associated topset beds suggest that crustal uplift was in progress when the Palmer Hill Delta was deposited, preventing the marine inundation and erosion that affected the deltas seen on Day 1 of this trip.

**Hanley Pit.** The Hanley Pit is located in the central part of the delta, where there are several kettleholes. The meltwater deposits in this pit are proximal topset and foreset beds. They are coarser grained than the sediments in the Crooker Pit, and contain abundant cobble-boulder gravel. The section was not well exposed during the preparation of this guidebook, but a face in the southwest part of the pit showed the deltaic sediments overlying till. It is not known whether the till is part of the basic ground moraine or a buried end moraine. Mounds of till were formerly exposed just north of here in the floor of the Potter Pit, and some of this till was interlayered with deformed sand and gravel. The latter deposits were probably end moraines, which commonly occur beneath deltas and submarine fans in coastal Maine.

**Rice Pit.** The Rice Pit is located in the ice-contact head of the delta, where it was fed by the esker. The sediments here include coarse gravel, and locally show ice-contact collapse structures. Unfortunately, many of the interesting parts of this deposit have been removed, and it is now difficult to reconstruct sedimentary environments at the ice margin. A series of exposures at different levels in the pit revealed lenses of compact, silty diamicton interlayered with sand and gravel. One such lens, high in the section, was up to 1.6 ft (0.5 m) thick and could be traced laterally for at least 45 ft (14 m), with a 10° dip to the northwest. The diamicts are glacial till, but it is not known whether they are the product of debris flows off the glacier margin (flowtills) or lodgement tills deposited by active ice overriding the back of the delta. In either case, they indicate a zone of debris-rich ice that reached at least to the top of the delta. It is rare to find such a large quantity of till intertonguing with the heads of glaciomarine deltas in Maine. Till is also exposed in the lowest level of the Rice Pit, below the presumed base of the esker-delta. The latter till unit is dark gray to olive-gray (5Y-4/1 to 4/2), silty, compact, massive, and contains striated stones up to 6 ft (2 m) in diameter.

*Return to Thayer Road and turn left; continue to junction with Route 194 and turn left. Follow Route 194 to junction with Route 27 and turn right and proceed to Randolph. A little over 1 mile turn left at lights and proceed over bridge (Routes 24/9/201) over Kennebec River. Turn left at end of bridge at lights to Route 24 and follow Route 24 south parallel to Kennebec River through the towns of Richmond, Bowdoinham, and Topsham. Fine views of the Kennebec River and Merrymeeting Bay (estuary where Kennebec and Androscoggin Rivers join). Turn left at lights (junction of Route 196) in Topsham and proceed south across bridge to Brunswick over Androscoggin River. Turn right onto Route 1 after crossing bridge and continue on Route 1 South. Turn left at third set of lights onto Church Road and proceed 3.75 miles to intersection with Bunganuc Road on left. Immediately after intersection turn left just before old farmhouse into Collete driveway and proceed to end of drive.*

**STOP 2. BUNGANUC BLUFF (Lunch stop;** after lunch walk down steps at left of house to salt marsh, turn left and follow marsh fringe out to open bay to bedrock outcrop; Freeport quadrangle; elev.ca sea level; leader T. K. Weddle).

**This stop can be very muddy!** Several large slump exposures along the coast reveal a section of glacial, glacial marine, and nearshore deposits that represent changing environments over time. A subglacial(?) and an ice marginal(?) component is present at the west end of the bluff, overlain by a thick, muddy medial to distal glacial marine fan or submarine plain component in the central and eastern end of the section, in turn overlain by nearshore deposits and possibly an eolian cap. Zink (1953), Amos and Sandford (1987), and Hay (1988) have published previous reports on the exposure, and Devin and Sandford (1990) recorded a test boring drilled to 70 feet depth without refusal at the site. Kelley and Hay (1986) and Kelley and Kelley (1988) have described the section in other field guides. Belknap and others (1989) and Belknap and Shipp (1991) described the section as an example of glacial marine mud deposited in part under an ice shelf or at least near the ice-grounding line. The record of graded, rhythmically-bedded fine sand, silt, and clay exposed here and in the

subsurface boring data concur with the latter rather than the former interpretation, at least for the lower part of the section, and represents either distal turbidite deposition or sedimentation from an overflow plume with cyclical sorting of fine sediment, attributed to an ice-margin source probably about 2 miles to the north, where there is surficial topographic expression of moraines. The lower deposits most likely record a glacial marine sedimentary source, whereas the upper deposits are more likely representative of marine and nearshore influenced sedimentation. One question to consider at the stop is the influence of cyclical process on discharge, such as daily, seasonal, and tidal processes as described by Cowan and Powell (1990) and Phillips and others (1991). A better understanding of the depositional environment at this exposure needs to be determined so that comparisons with modern glacial marine and subglacial marine environments can be made (cf. Anderson and Domack, 1991; Anderson and others, 1991).

Beginning at the west end of this exposure, striated bedrock (170°) is overlain by a very poorly exposed section consisting of a thin veneer of sandy till (best represented by numerous boulders at the base of the bluff), a massive sand, and glacial marine mud. Eastward in newly exposed slides, near the top of the section is 6 to 8 feet (2 - 3 m) of medium- to fine-grained sand, which contains rip-up mud clasts. The sand is found in erosional channels into underlying syndepositionally deformed stratified sand, silt, and clay. In gradational contact with the upper dominantly sandy portion of the bluff, muddy sediments are best exposed in slumps in the eastern end of the section. These lower deposits are thin, well-stratified, and laterally continuous silt and clay beds with silty fine sand laminations, grading upward into more massive mud. Dropstones are extremely rare. At the eastern end of the section, striated bedrock (195° to 201°) is exposed at the waterline. Southwest-oriented striae are found along many of the peninsulas and islands in this part of Casco Bay, and reflect the strong topographic control on the Late Wisconsinan ice by the northeast trend of the bedrock structure in the area. Hussey (1981) maps the trend of the Flying Point fault along Maquoit Bay just to the east. Seismic surveys by Kelley and others (1987) show that depth to bedrock in the bay troughs east of the bluff is over 300 feet deep in places. These data suggest that ice may have preferentially flowed along the northeast-trending troughs, in which less resistant, brittle-faulted rock is found. At some localities, the southwest-oriented striae are younger than the dominant southeast set. However, this relationship has not been seen consistently enough to warrant regional merit, and in other places in the area the striae have the reverse age-sense.

Kelley and others (1988) have described the Holocene history of salt marshes along coastal Maine, including the Bunganuc bluffs area. The slumping of the bluffs provides mud to the clam flats in Maquoit Bay, sustaining a major industry and resource. However, the homes built above the bluffs have experienced significant loss of property. Engineering and geotechnical aspects of landslides in the Presumpscot Formation are discussed in Amos and Sandford (1987), Andrews and others (1987), and Devin and Sandford (1990). Human activity over the past 250 years has affected the coast during its evolution. In particular, deforestation allowed erosion, and subsequently provided more sediment to the coast causing an increase in tidal flat extent and marsh development. Later, vegetation was reestablished and sediment supply to the coast was diminished, resulting in erosion of the marshes. A small, eroding salt marsh is present in the central area separating the two main areas of the bluff. The marsh has eroded enough to expose horizontally oriented logs beneath it. Modern trees at the bluffs come down with recent slumps, and the buried trees in the peat have their recent counterparts resting at present at the toe of the slumping bluffs.

*Return to bus and to Church Road, back to Route 1 and turn left at lights. Follow Route 1 South to I-95 South to Portland (stay on I-295 through Portland) to its junction with the Maine Turnpike in South Portland. Continue south following signs to Boston. Travel time from Brunswick to Boston is approximately 2.5 hours. End of Day 3 excursion and end of trip.*

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#### **APPENDIX A: CHARACTERISTICS AND FORMATION OF WAVE DEPOSITS**

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Wave deposits are recognized based upon three types of observations: the geometry of sediment units, the presence of sedimentary structures characteristic of wave deposits, and the sorting of sedimentary deposits. Sediment erosion, transport, and deposition by waves is fundamentally different from that of unidirectional flow, although there is some overlap in behavior. Waves are produced by wind stress on the water surface. In the simplest case, waves comprise a sinusoidal surface that induces a series of circular turbulent eddies in the water called orbitals (Fig. 1). The diameter of these orbitals and their orbital velocity decrease with depth below the surface wave. The depth to which there is significant movement of water is roughly one-half the length of distance between two wave crests (i.e. one-half the wave length). The magnitude of the orbital velocity depends upon the wave height and period (amount of time for two consecutive wave crests to pass a point) which are controlled by the amount of wind stress on the surface. As waves propagate into shallower water the orbitals exert a stress on the sediment surface and move loose particles. This stress deforms the orbitals into ellipses that are progressively flatter near the sediment - water interface where they are completely flattened into a bimodal linear vector. When water depth is shallower than one-sixth the wave length, the drag of the wave on the bottom deforms the shape of the wave from a sinusoidal surface to one with broad troughs between narrow crests, and

results in an asymmetry in the bottom movement vector toward the direction of propagation. Bottom drag of waves also slows their rate of propagation resulting in refraction of the crestlines (Fig. 2). On low-slope surfaces, the refraction makes wave crests bend parallel to the shoreline. Angular incidence of waves induces a current parallel to the shoreline called a longshore current. Where waves intersect steep surfaces projecting out of the water, the waves have greater erosional force and they are diffracted around the object causing crests to bend toward the leeside of the object (Fig. 2). The refraction and diffraction of waves as they intersect shallow water results in several distinct geometries. Refraction and longshore drift makes ridges of sediment (bars and spits) that are characteristically long and straight parallel to shore or slightly curved. Diffraction of waves around projections in the water produces linear zones of sediment accumulation on the leeside called tombolos. Wave-cut benches form where waves are not significantly slowed by a bottom slope before intersecting a surface.

Waves produce a variety of sedimentary structures that could be misinterpreted as fluvial or deltaic in origin. The primary wave structure is an oscillatory ripple (Fig. 3). Oscillatory ripples are characterized by straight to slightly sinuous crests, flat to scalloped bases, sinusoidal internal foresets, and may have opposing orientations. Unlike ripples formed under unidirectional flow, oscillatory ripples can form in average grain sizes greater than 0.6 mm in diameter. Also, unlike ripples under unidirectional flow, oscillatory ripples vary in size from a millimeter thick to over a meter thick with wavelengths varying from a centimeter to several meters. In areas with large long-period waves, dune-scale features with curved crests may form. These structures produce trough cross-bedding very similar to that formed by unidirectional flow. The important features that may help to recognize the wave-formed trough crossbeds is the sorting of the internal strata (see below) and the association with oscillatory ripple structures. Wave formed bars have tabular bedding sets, commonly oriented shoreward or parallel to shore reflecting storm wave propagation or longshore drift (Fig. 4). Wave-formed tabular sets are highly variable in thickness (from a decimeter to several meters in thickness); they may have intercalated oscillatory ripple structures; they may have opposing orientation; and commonly each foreset layer is comprised of moderately well-sorted grain sizes, but adjacent layers may be comprised of highly variable grain sizes. This latter trait reflects the fact that wave strength is dependent on wind stress which is highly variable. Breaking waves may form beaches comprised of planar lamination that forms low-angle sets that are inclined seaward. These planar laminae may be intercalated with tabular foresets of bars.

The characteristic motion of wave stress has a tendency to segregate grains by size and shape. This grain segregation defines the layering in ripples, bars, and beach deposits. In coarse-grained deposits the grain size and shape segregation is more obvious and diagnostic of wave sorting. Coarse pebbles and cobbles that have been transported by waves are separated into beds comprised of open framework clasts of the same size and shape (Fig. 5A). Platy clasts may be imbricated (sometimes in different orientations) and spaces between clasts may be filled with finer grains. The relation of waves to wind stress results in abrupt changes in grain sizes within a deposit. Where waves have reworked a bouldery deposit, the sorting is not as obvious. The largest clasts may be too big for waves to move during most conditions, so they are left as lags as the finer material is reworked. Waves are diffracted around the boulders which disrupts the energy resulting in an irregular sorting of the finer grained material. These poorly sorted deposits are identifiable by the open framework sorting of the matrix into patches of the same size and shape clasts (Fig. 5B).

In the field area, bar and tombolo deposits, and wave-cut benches in moraines and other hills such as drumlins are present. Cross-sections of bars and tombolos consist of tabular crossbeds that have been misinterpreted as deltaic deposits. In addition to their gross geometry, these deposits can be distinguished by the absence of unidirectional ripple crosslaminae, the size sorting within layers, and the orientation of the foresets. In some places, evidence of tidal overprints include clay drapes, bioturbation breaks within layering, and erosional reactivation surfaces within foreset sequences. Oscillatory ripple cross-lamination has been observed in a few places. Wave-sorted conglomerates are commonly associated with wave-cut benches into tills. The sorting is typically patchy around the lags of large boulders. End-member varieties of wave deposits are recognizable, but there are still lots of deposits where the criteria are not obvious. In these latter cases, the interpretation is dependent upon the large-scale context of the occurrence as much as the internal textures.

Raaf, J. F. M. De., Boersma, J. R., and Van Gelder, A., 1977, Wave-generated structures and sequences from a shallow marine succession, Lower Carboniferous, County Cork, Ireland: *Sedimentology*, v. 24, p. 451-483.

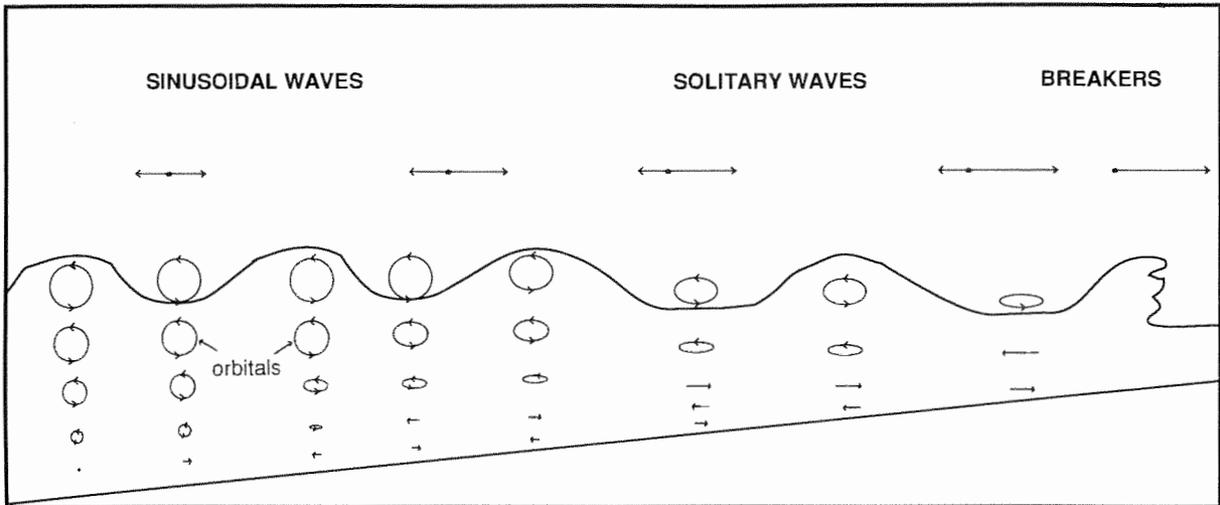


Figure 1. Shape of orbitals in waves from deeper water to shallower water. Vectors of net sediment transport are shown on top.

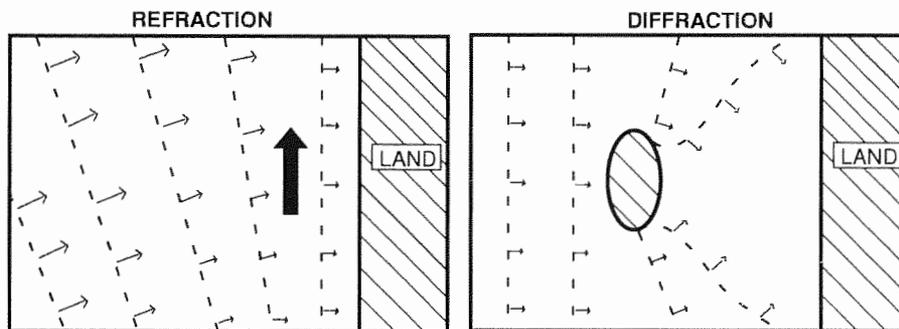


Figure 2. Refraction and diffraction of wave crests (dashed). Small arrows indicate vectors of wave propagation. Large arrow is longshore drift.

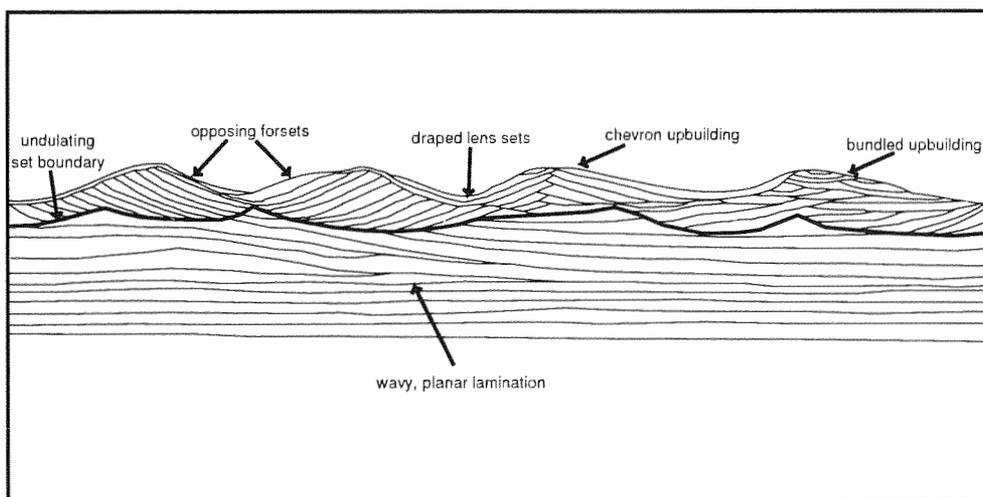


Figure 3. Characteristic features of oscillatory ripple cross-lamination. From Raaf and others (1977)

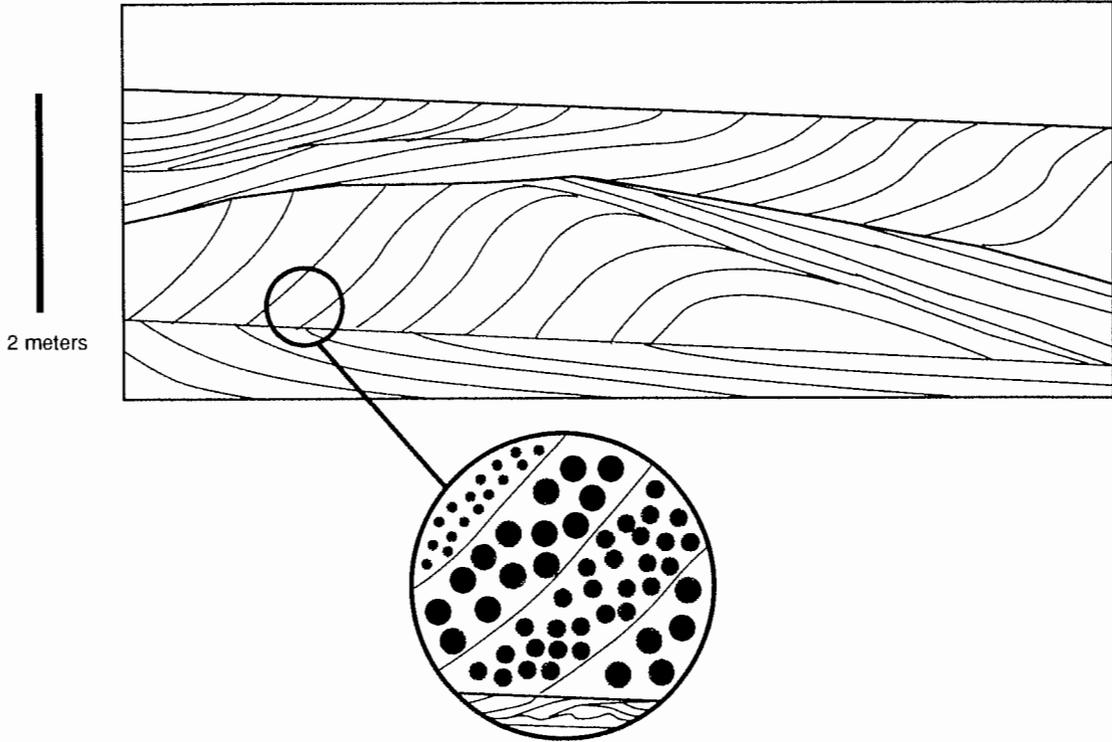


Figure 4. Crossbedding in a wave-formed bar showing opposing sets, grain sorting in layers, and oscillatory ripples in partings.

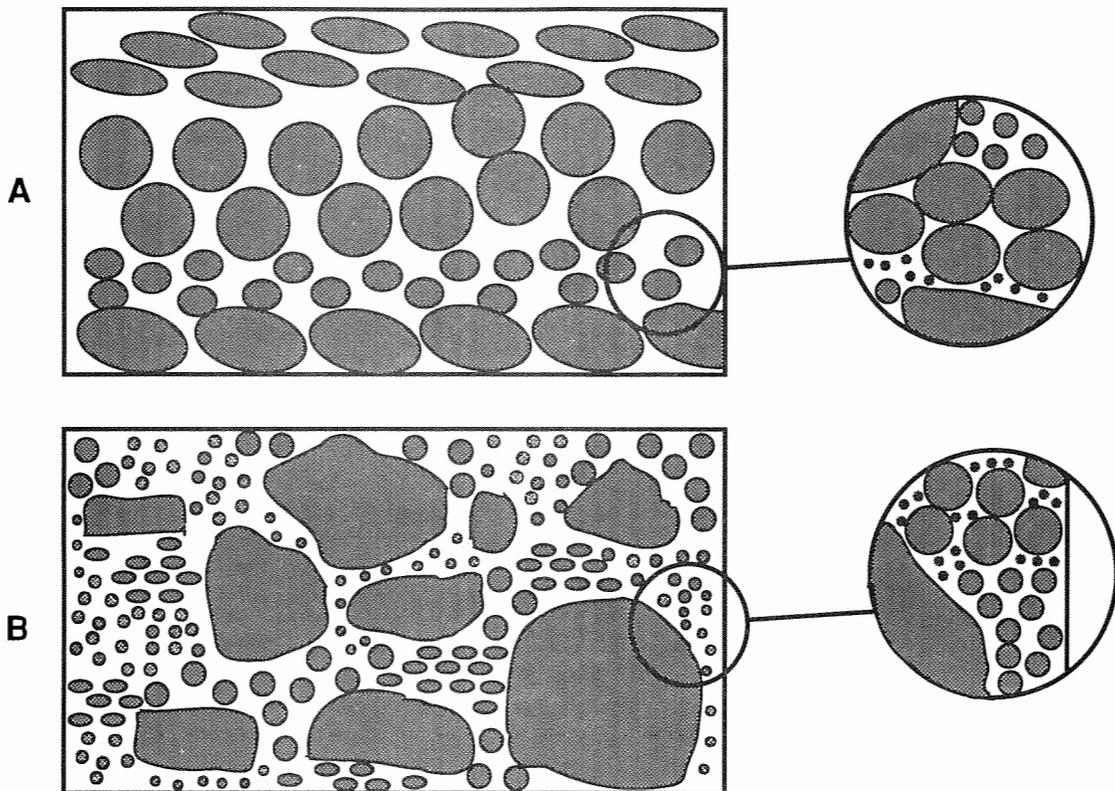


Figure 5. Wave sorting of conglomerates. A) Clasts sorted by size and shape. B) Large clasts are not moved by waves and matrix has patchy sorting. Note sorting of intergranular material in insets.



Chapter J

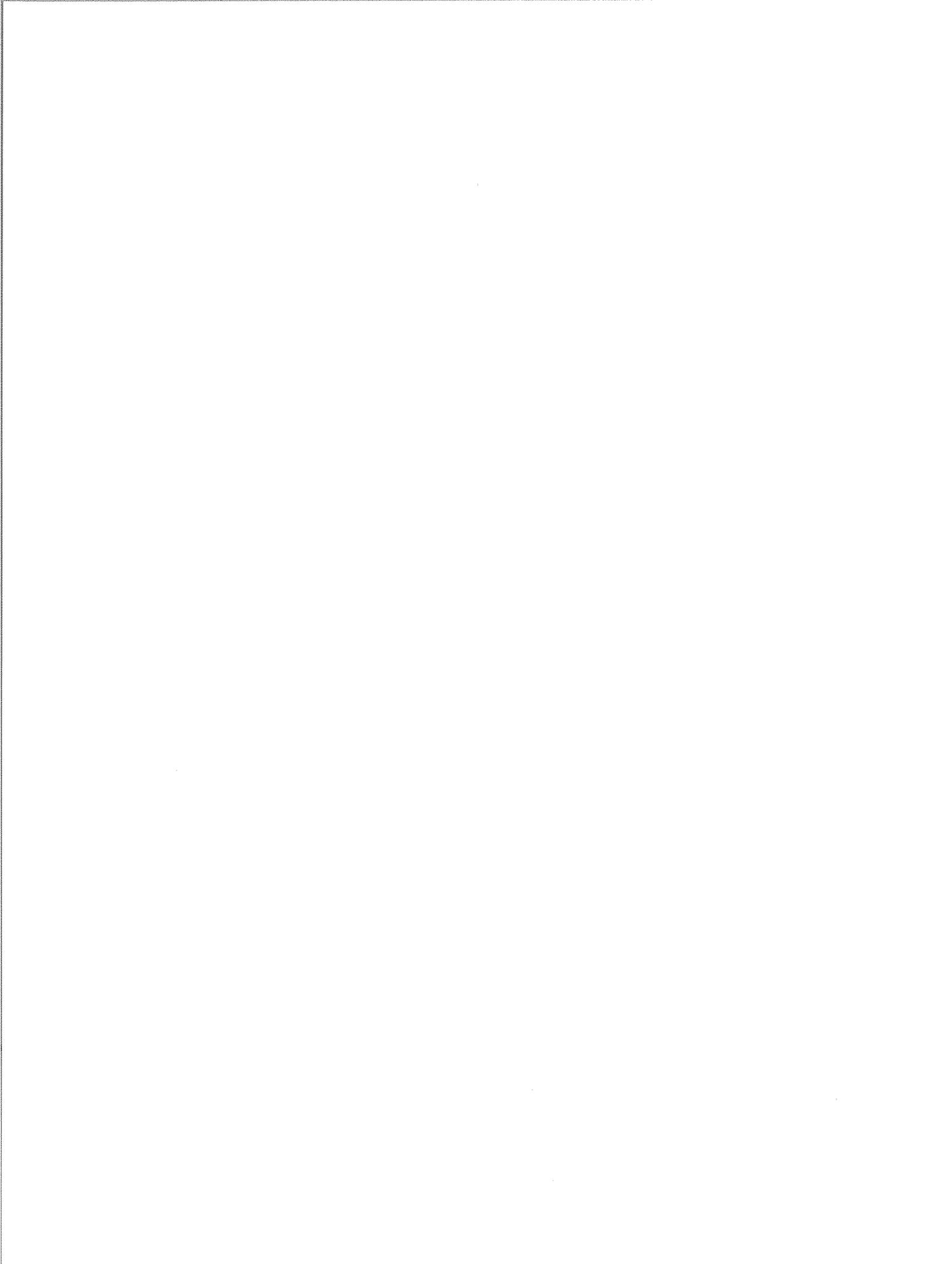
# Geochronology and Petrogenesis of Adirondack Igneous and Metamorphic Rocks

By James M. McLelland

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## GEOCHRONOLOGY AND PETROGENESIS OF ADIRONDACK IGNEOUS AND METAMORPHIC ROCKS

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### INTRODUCTION AND GEOCHRONOLOGY

The Adirondacks form a southwestern extension of the Grenville Province (figure 1) and are physiographically divided into the Adirondack Highlands (granulite facies) and Lowlands (amphibolite facies) by a broad zone of high strain referred to as the Carthage-Colton Mylonite Zone (figs. 2,3) which is continuous with the Chibougamau-Gatineau line, or Labelle shear zone, in Canada (AB on figure 1). Together these two zones separate the Grenville Province into two major blocks with the Central Granulite Terrane (CGT) lying east of AB and the Central Metasedimentary Belt (CMB) and Central Gneiss Belt (CGB) lying to the west. Within the southwestern portion of the Grenville Province further subdivisions exist and are shown in figure 3.

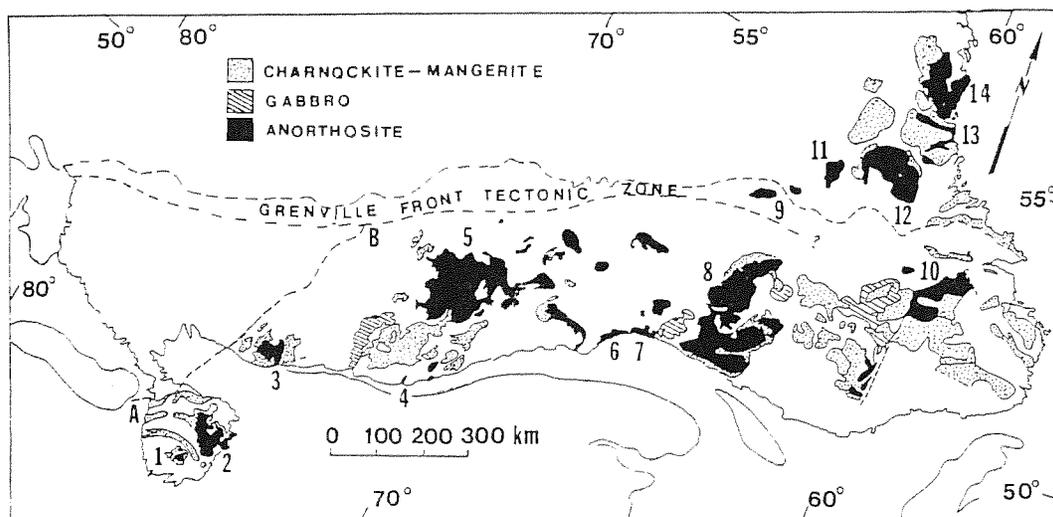


Figure 1. Generalized map of anorthositic massifs within the Grenville Province and adjacent Labrador. The dashed line, AB, separates terranes with anorthosite massifs on the east from ones lacking them on the west and corresponds to the Carthage-Colton-Gatineau-Chibougamau Line. 1-Snowy Mt. and Oregon domes (ca. 1130 Ma); 2-Marcy massif (ca. 1135 Ma); 3-Morin anorthosite and Lac Croche complex ( $1160 \pm 7$  Ma); 4-St. Urbain anorthosite (ca. 1070 Ma); 5-Lac St. Jean complex ( $1148 \pm 4$  Ma); 6-Sept Isles ( $1646 \pm 2$  Ma); 7-8-Havre St. Pierre complex ( $1126 \pm 7$  Ma) including the Pentecote ( $1365 \pm 7$  Ma) anorthosite; 9-Shabagamo intrusives; 10-Mealy Mts. anorthosite ( $1646 \pm 2$  Ma); 11-12-Harp Lake anorthosite (ca. 1450 Ma); 13-Flowers River complex (ca. 1260 Ma); 14-Nain complex (1295 Ma) including Kiglapait intrusive ( $1305 \pm 5$  Ma). From McLelland (1989).

As demonstrated by recent U-Pb zircon and Sm-Nd geochronology summarized (table 1) by McLelland and Chiarenzelli (1990) and McLelland et al. (1993), the Adirondack-CMB sector of the Grenville Province contains large volumes of metaigneous rocks that represent recent (i.e., ca. 1400-1200 Ma) additions of juvenile continental crust. These results (figure 4) indicate that the Adirondack-CMB region experienced wide-spread calcalkaline magmatism ca. 1300-1220 Ma. Associated high grade metamorphism has been fixed at  $1226 \pm 10$  Ma by Aleinikoff (pers. comm.) who dated dust air abraded from metamorphic rims on 1300 Ma zircons. Identical rocks, with identical ages, have been described from the Green Mts. of Vermont by Ratcliffe et al. (1991), in northern Ireland by Menuge and Daly (1991), in the Llano uplift of Texas (Walker, 1993) and in the Texas-Mexico belt of Grenville rocks by Patchett and Ruiz (1990). It appears, therefore, that a major collisional-magmatic belt was operative along the present southern flank of the Grenville Province during the interval 1300-1220 Ma and may have been related to the assembly of a supercontinent. More locally, this magmatism and associated metamorphism, represents the Elzevirian Orogeny of the Grenville Orogenic Cycle, as defined by Moore and Thompson (1980). Within the Adirondacks, Elzevirian rocks are represented by 1300-1220 Ma tonalites and alaskites whose distribution is shown in figure 5. The apparent absence of this suite from the central Highlands is believed to be the combined result of later magmatic intrusion and recent doming along a NNE axis. Within the Frontenac-Adirondack region, the Elzevirian Orogeny was followed by 40-50 Ma of quiescence terminated at 1170-1130 Ma by voluminous anorogenic (figure 4) magmatism referred to as the

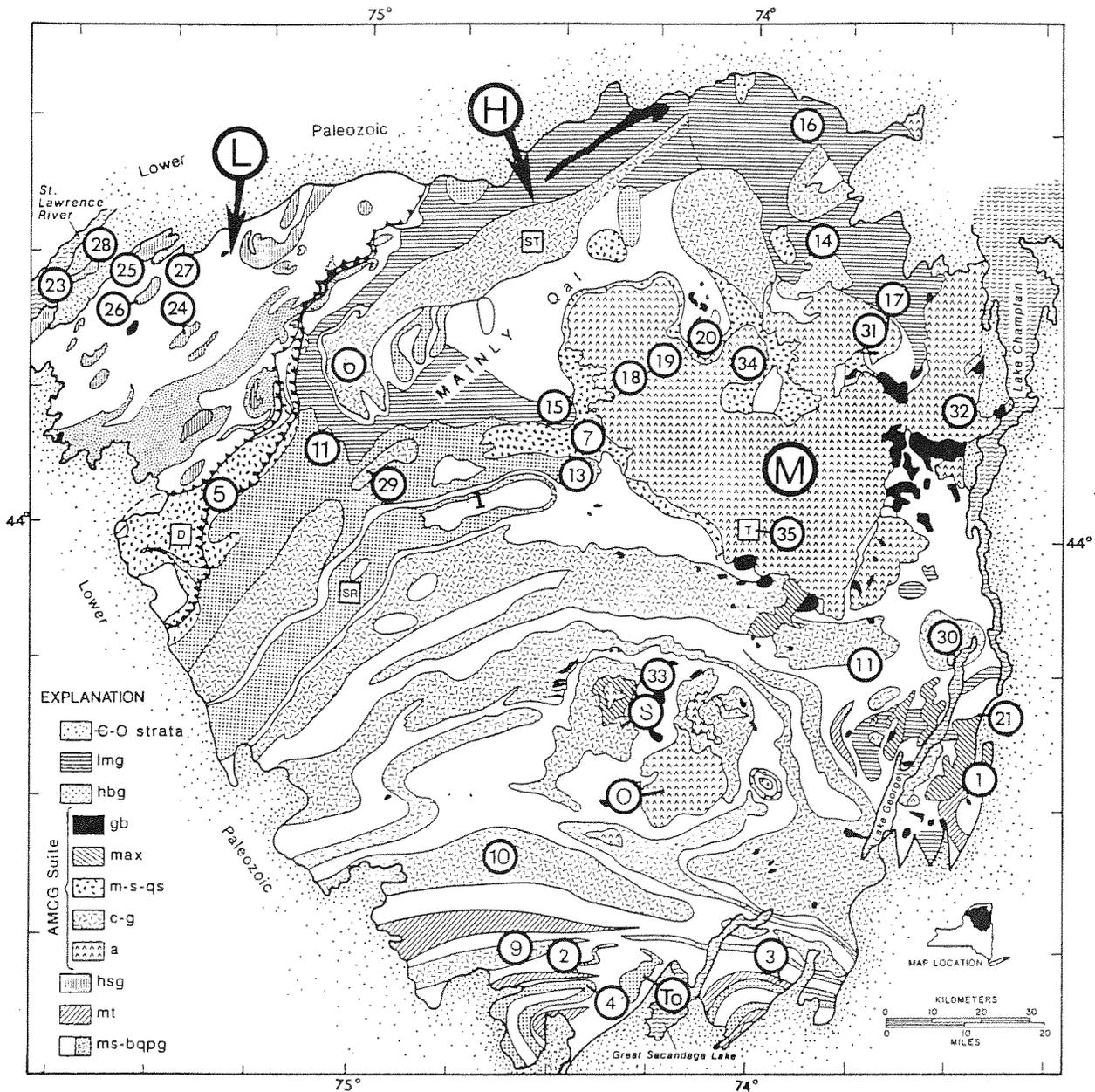


Figure 2. Generalized geologic map of the Adirondack Highlands (H) and Lowlands (L). The Carthage-Colton Mylonite Zone (CCMZ) is shown with sawteeth indicating directions of dip. **Numbers refer to samples listed in Tables 1 and 2.** Map symbols: lmg=Lyon Mt. Gneiss, hbg=hornblende-biotite granitic gneiss, gb=olivine metagabbro, max=mangerite with andesine xenocrysts, a=metanorthosite, m-s-qs=mangeritic-syenitic-quartz-syenitic gneiss, ms=metasediments, bqpg=biotite=quartz-plagioclase gneiss, hsg=Hyde School Gneiss, mt=metatonalitic gneiss. Locality symbols: A=Arab Mt. anticline, C=Carthage anorthosite, D=Diana complex, O=Oregon dome, S=Snowy Mt. dome, ST=Stark complex, SR=Stillwater Reservoir, T=Tahawus, To=Tomantown pluton. From McLelland and Chiarenzelli (1990) and Daly and McLelland (1991).

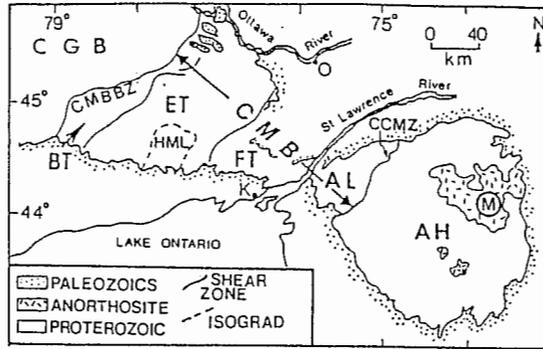


Figure 3. Southwestern Grenville Province. CMB=Central Metasedimentary Belt, CGB=Central Gneiss Belt, BT=Bancroft Terrane, ET=Elzevir Terrane, FT=Frontenac Terrane, AL=Adirondack Lowlands, HL=Adirondack Highlands, HML=Hastings metamorphic low, K=Kingston, O=Ottawa, CCMZ=Carthage-Colton Mylonite Zone, M=Marcy massif.

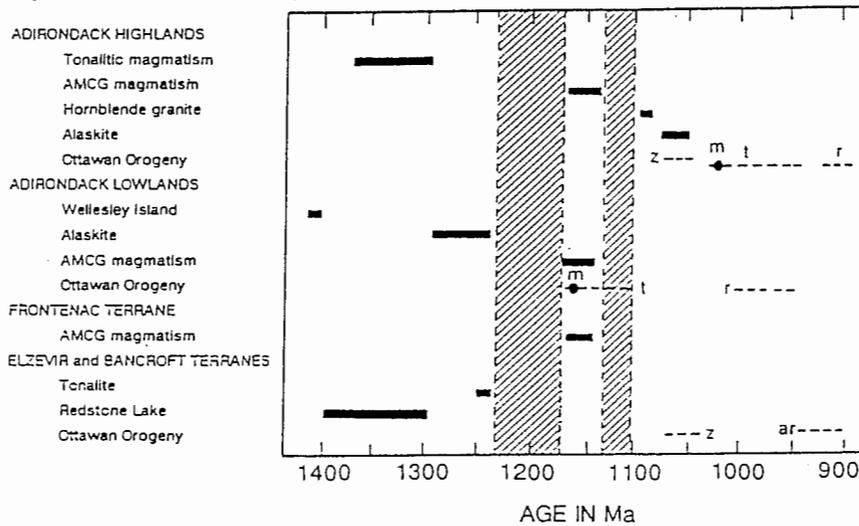


Figure 4. Chronology of major geological events in the southwestern Grenville Province. z=zircon, t=titanite, m=monazite, r=rutile, ar=Ar/Ar. Diagonal ruling=quiescence. From McLelland and Chiarenzelli (1991).

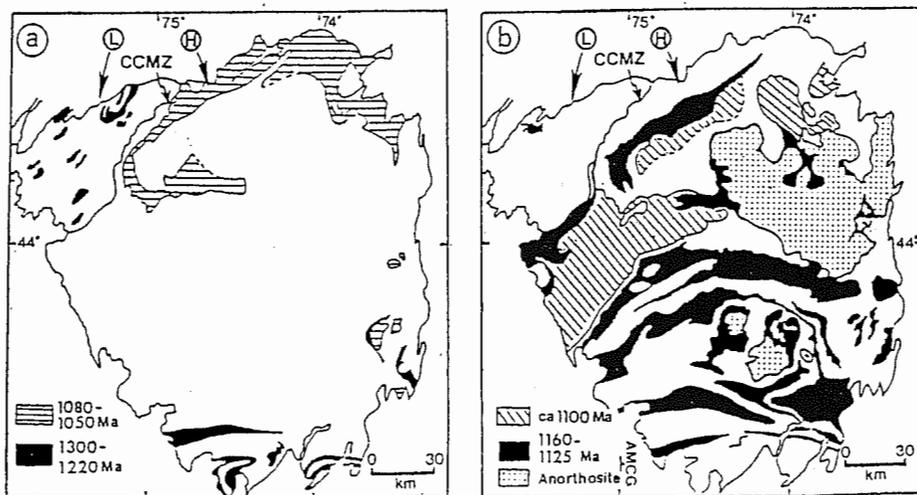


Figure 5. Chronological designation of Adirondack units. L=Adirondack Lowlands, H=Adirondack Highlands, CCMZ=Carthage Colton Mylonite Zone. From Chiarenzelli and McLelland (1991).

anorthosite-mangerite-charnockite-granite (AMCG) suite. The older ages are characteristic of AMCG magmatism in the Frontenac Terrane (including the Lowlands) while the Highlands commonly exhibit ages of 1150-1130 Ma (figure 5). The large Marcy anorthosite massif (figure 2) and its associated granitoid envelope were emplaced at ca. 1135 Ma (McLelland and Chiarenzelli, 1990). These ages are similar to those determined (Emslie and Hunt, 1990) for the Morin, Lac St. Jean, and several other large massifs farther northeast in the Grenville Province (figure 1). Rocks of similar age and chemistry (i.e., Storm King Granite) have been described within the Hudson Highlands (Grauch and Aleinikoff, 1985). The extremely large dimensions of the AMCG magmatic terrane emphasize its global-scale nature corresponding, perhaps, to supercontinent rifting with the rifting axis located farther to the east. Valley et al. (1990) and McLelland et al. (1991) have provided evidence that contact, and perhaps also regional, metamorphism accompanied emplacement of hot ( $\sim 1100^\circ\text{C}$ ), hypersolvus AMCG magmas. Wollastonite and monticellite occurrences related to thermal pulses from AMCG intrusions occur in proximity to AMCG intrusions (Valley et al., 1990). In the Lowlands and the Canadian sector of the Frontenac Terrane, monazite (table 1., no. 28), titanite, and garnet ages (Mezger et al., 1992) all indicate high temperatures ( $\sim 600\text{-}800^\circ\text{C}$ ) at ca. 1150 Ma. Rutile ages and Rb/Sr whole rock isochron ages document temperatures not exceeding  $\sim 400^\circ\text{C}$  at ca. 1050-1000 Ma.

Following approximately 30 Ma of quiescence (figure 4), the Adirondacks, along with the entire Grenville Province, experienced the onset of the Ottawa Orogeny of the Grenville Orogenic Cycle. Initially the Ottawa Orogeny was represented by 1090-1100 Ma hornblende granites in the northwest Highlands. These rather sparse granites were followed by deformation, high grade metamorphism, and the emplacement of trondhjemitic to alaskitic magnetite-rich rocks (Lyon Mt. Gneiss of Whitney and Olmsted, 1988) in the northern and eastern Adirondacks. The zircon ages of these rocks fall into an interval of 1050-1080 Ma (table 1) which corresponds to the peak of granulite facies metamorphism when crust, currently at the surface, was at  $\sim 25$  km. Accordingly, the alaskitic to trondhjemitic rocks are interpreted as synorogenic to late-orogenic intrusives. They were followed by the emplacement of small bodies of fayalite granite (ca. 1050 Ma) at Wanakena and Ausable Forks (figure 2).

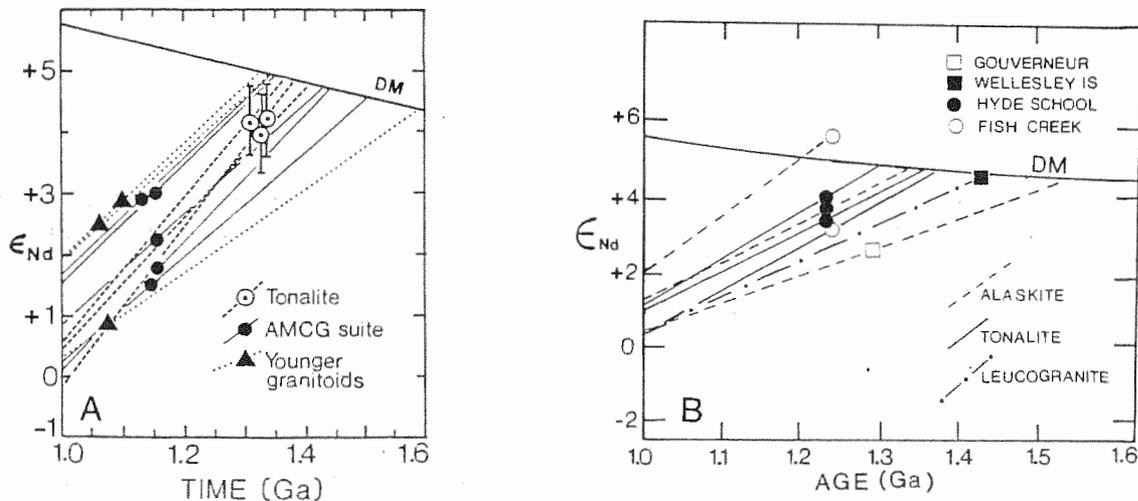


Figure 6. Epsilon-Nd diagrams for orthogneisses of the Adirondack Highlands (A) and Lowlands (B). Symbols fixed by zircon ages.

Sm-Nd analysis (Daly and McLelland, 1991) demonstrates that the emplacement ages of the ca. 1300 Ma tonalitic rocks of the Highlands correspond closely to their neodymium model ages (table 1 and figure 6) indicating that these represent juvenile crustal additions. As seen in figure 6,  $\epsilon_{\text{Nd}}$  evolution curves for AMCG and younger granite suites pass within error of the tonalitic rocks and suggest that the tonalites, together with coeval granitoids, served as source rocks for succeeding magmatic pulses. Remarkably, none of these igneous

suites gives evidence for any pre-1600 Ma crust in the Adirondack region and the entire terrain appears to have come into existence in the Middle to Late Proterozoic. Significantly, Sm-Nd analysis for the ca. 1230-1300 Ma tonalitic to alaskitic Hyde School Gneiss at the Lowlands (table 1, figure 6) demonstrates that it has model neodymium ages and  $\epsilon_{Nd}$  values similar to Highland tonalites (McLelland et al., 1993). These results are interpreted to reflect the contiguity of the Highlands and Lowlands at ca. 1300 Ma. Given this, the Carthage-Colton Mylonite Zone is interpreted as a west-dipping extensional normal fault that formed during the Ottawaan Orogeny in response to crustal thickening by thrust stacking (McLelland et al., 1993). East dipping extensional faults of this sort and age have been described by van der Pluijm and Carlson (1989) in the Central Metasedimentary Belt. Motion of this sort along the Carthage-Colton Mylonite Zone would help to explain the juxtaposition of amphibolite and granulite facies assemblages across the zone. A downward displacement of 3-4 km would satisfactorily explain the somewhat lower grade of the Lowlands terrane.

### PETROLOGIC CHARACTERISTICS OF THE PRINCIPAL ROCK TYPES IN THE ADIRONDACKS

The following discussion is divided into igneous and metasedimentary sections. Whole rock analyses for granitoids are given in table 2 while those for anorthositic and gabbroic rock appear in tables 3 and 4.

#### Igneous Rocks

**Tonalites and related granitoids.** Typical whole rock chemistries for these rocks are shown in figures 7-9. Figure 8 shows the normative anorthite (An)-albite (Ab)-orthoclase (Or) data for these rocks and compares them to similar rocks in the Lowlands. AFM plots are given in figure 8 and calc-alkali index versus silica plots in figure 9; both figures illustrate the strongly calcalkaline nature of the Highland tonalite to granitoid suite. The tonalitic rocks, which will be visited at Stop 1, outcrop in several belts within the southern and eastern Adirondacks. In the field they can be distinguished from, otherwise similar, charnockitic rocks by the white alteration of their weathered surfaces and the bluish grey on fresh surfaces. A distinctive characteristic is the almost ubiquitous presence of discontinuous mafic sheets. These have been interpreted as disrupted mafic dikes coeval with emplacement of the tonalites. Associated with the tonalitic rocks are granodioritic to granitic rocks containing variable concentrations of orthopyroxene.

**AMCG Suite.** Within the Adirondack Highlands AMCG rocks are widely developed and abundantly represented in the Marcy massif as well as the Oregon and Snowy Mt. Domes. The chemistry of granitoid (mangeritic to charnockitic) facies of these rocks is given in Table 2 and figures 9 and 10, both for the older as well as the younger anorogenic plutonic rocks. As shown in figure 9, the AMCG rocks have calcalkali-silica trends that are distinctly different than those shown by the tonalitic suites. McLelland and Whitney (1991) have shown that the AMCG rocks exhibit anorogenic geochemical characteristics and constitute bimodal magmatic complexes in which anorthositic to gabbroic cores are coeval with, but not related via fractional crystallization to, the mangeritic-charnockitic envelopes of the AMCG massifs (i.e., Marcy massif, figure 2). Bimodality is best demonstrated by the divergent differentiation trends (figure 11) of the granitoid members on the one hand and the anorthositic-gabbroic rocks on the other (Buddington, 1972). Eiler and Valley (pers. comm.) report that  $\delta^{18}O$  values for AMCG granitoids are magmatic in origin and demonstrate that these rocks are related by fractional crystallization and were metamorphosed under vapor-absent conditions. Anorthositic members of the AMCG suite have distinctly different  $\delta^{18}O$  values. The extreme low-SiO<sub>2</sub>, high-iron end members of the anorthosite-gabbro family will be seen at Stop 10 and are believed to represent late liquids developed by plagioclase fractionation under conditions of low oxygen fugacities (i.e., dry, Fenner-type trends). Associated with these are large magnetite-ilmenite deposits which will be visited at Sanford Lake.

**Younger Hornblende Granitic Rocks.** The distribution of these rocks is shown in figure 5a, and their ages are given in table 1. An example of these rocks will be visited at Stop 14. In the field these rocks consist of medium grained, pink, streaky granitic rocks containing hornblende and minor biotite. They are difficult to distinguish from the granitic facies of the AMCG suite. As pointed out by Chiarenzelli and McLelland (1991), their restriction to the northwestern Highlands is intriguing but not yet understood.

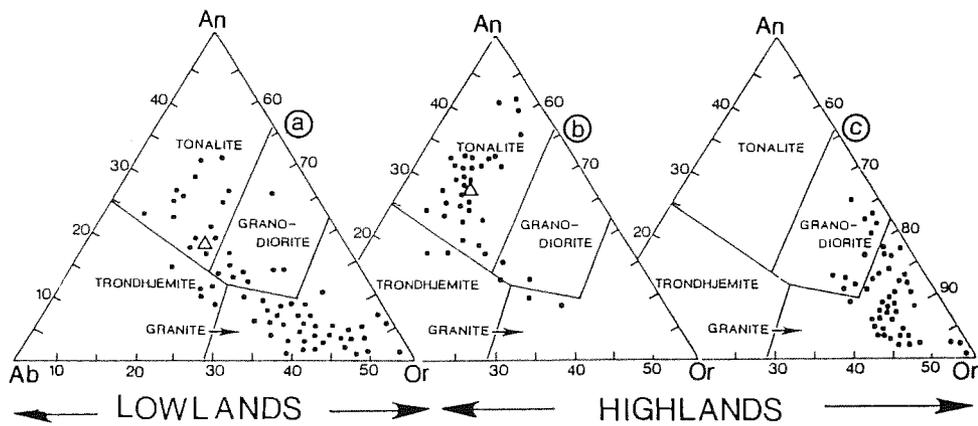


Figure 7. Plots of normative albite (Ab)-anorthite (An)-orthoclase (Or) for (a) Hyde School Gneiss, (b) Highlands tonalites, and (c) Tomantown pluton. Open triangles give average values for tonalitic samples. Definition of fields due to Barker (1979).

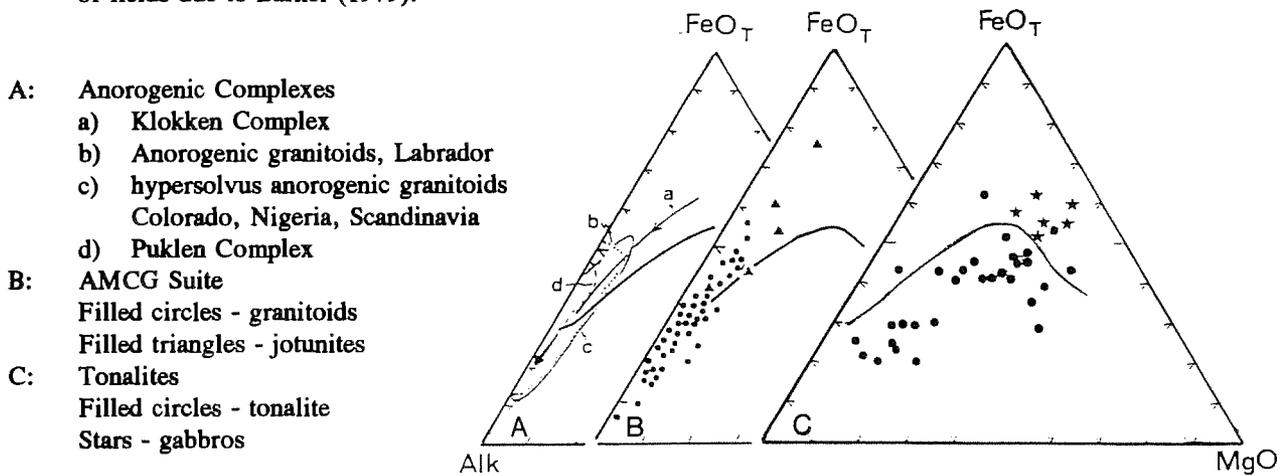


Figure 8. AFM variation diagrams for A) anorogenic complexes, B) AMCG suite, C) Highlands tonalites and associated gabbro (see McLelland 1991 for sources).

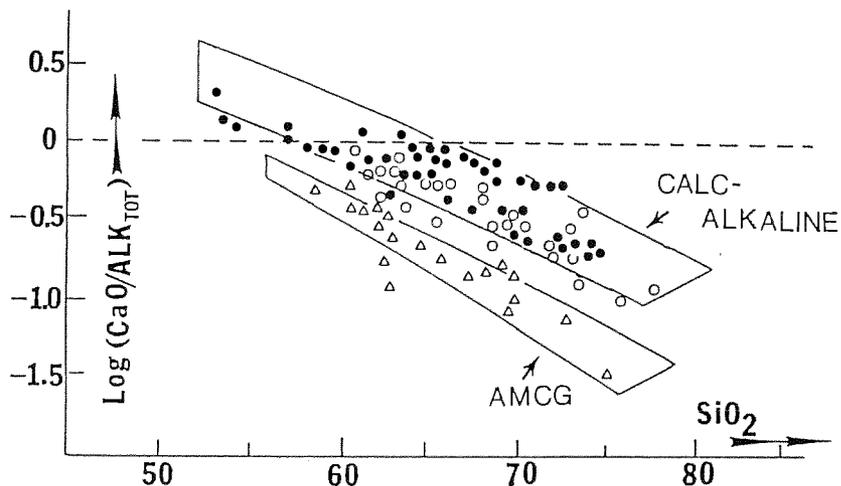


Figure 9. Calcalkali ratio vs. weight percent for AMCG granitoids (triangles), tonalites (closed circles) and Tomantown pluton granitoids (open circles).

TABLE 2

EARLY CALCALKALINE ROCKS				OLDER ANOROGENIC PLUTONIC ROCKS							
	AM-87-13	TOE	CL-6	AM-86-17	AM-86-1	AC-85-1	AM-86-9	AM-86-15	AC-85-2		
SiO <sub>2</sub>	65.00	65.68	74.63	68.90	71.88	73.72	69.14	67.47	75.17		
TiO <sub>2</sub>	.75	1.16	.37	.59	.36	.04	.89	.72	.20		
Al <sub>2</sub> O <sub>3</sub>	15.10	14.97	14.22	14.50	14.82	13.54	13.78	15.12	12.63		
FeO	nd	nd	nd	2.16	1.27	.87	2.83	3.34	1.11		
Fe <sub>2</sub> O <sub>3</sub>	6.03	6.79	1.53	1.1	.96	.11	1.82	1.59	1.07		
MnO	.10	.08	.04	.02	.03	.01	.04	.10	.02		
MgO	.46	1.15	.55	.84	.43	.20	.45	.51	.19		
CaO	2.71	2.56	1.66	2.3	1.87	.85	2.26	2.57	.88		
Na <sub>2</sub> O	4.10	2.80	3.56	3.06	3.93	5.71	3.07	3.41	2.99		
K <sub>2</sub> O	5.13	4.52	4.26	4.18	3.99	4.42	4.91	5.18	5.49		
P <sub>2</sub> O <sub>5</sub>	.10	.51	.10	.24	.09	.01	.23	.19	.04		
LOI	.39	.74	...	.40	.27	.17	.19	.39	.17		
Σ	99.87	100.96	99.42	99.63	99.65	99.61	100.59	99.96	99.96		
Ba(ppm)	1230	710	510	680	660	1100	736	810	442		
Rb (ppm)	100	97	170	160	200	406	81	128	230		
Sr (ppm)	260	230	260	200	130	26	211	215	99		
Y (ppm)	70	70	37	40	110	321	60	71	77		
Nb (ppm)	20	19	17	30	30	15	19	21	15		
Zr (ppm)	790	345	160	270	670	118	538	546	284		
YOUNGER ANOROGENIC PLUTONIC ROCKS											
	AC-85-6	AC-85-10	AM-86-7	WPG	SLC	AM-87-9	AM-86-8	AM-87-10			
SiO <sub>2</sub>	62.15	54.94	58.50	69.20	6.64	61.05	6.94	62.14			
TiO <sub>2</sub>	.88	1.55	.68	.51	1.14	.78	1.39	.36			
Al <sub>2</sub> O <sub>3</sub>	16.40	14.87	2.32	13.90	15.27	15.98	15.91	12.35			
FeO	3.96	1.25	2.81	3.1	9.28	4.60	6.51	1.32			
Fe <sub>2</sub> O <sub>3</sub>	1.49	2.80	.43	1.34	1.77	2.10	1.02	1.7			
MnO	.09	.24	.01	.05	.19	.05	.14	.01			
MgO	1.06	.96	1.47	.52	.74	1.64	1.70	.83			
CaO	3.27	5.52	6.16	2.03	3.97	3.63	4.53	3.65			
Na <sub>2</sub> O	4.81	3.45	5.02	3.02	3.34	3.41	3.55	6.05			
K <sub>2</sub> O	5.13	3.83	3.35	5.48	3.70	4.76	3.86	1.26			
P <sub>2</sub> O <sub>5</sub>	.30	.65	.32	.11	.31	.42	.46	.09			
LOI	.41	.37	.43	.39	.01	.91	.37	.67			
Σ	99.95	99.43	99.50	99.65	100.46	99.63	100.38	99.24			
Ba(ppm)	850	625	nd	1279	823	nd	1100	nd			
Rb (ppm)	106	47	nd	124	87	nd	83	29			
Sr (ppm)	335	367	nd	184	215	nd	410	180			
Y (ppm)	60	55	nd	48	121	nd	110	63			
Nb (ppm)	21	14	nd	14	38	nd	25	79			
Zr (ppm)	464	431	nd	382	647	nd	1200	309			
YOUNGER GRANITIC ROCKS						LATE LEUCOGRANITIC ROCKS					
	AM-86-3	AM-86-6	NO-Fo1	AM-86-13	AM-87-6	GHA	AM-86-11	AM-87-7	AM-86-4	AM-86-10	AM-86-14
SiO <sub>2</sub>	68.62	68.05	71.75	76.30	69.00	73.2	69.98	67.80	7.01	69.05	72.39
TiO <sub>2</sub>	.48	.55	.41	.18	1.43	.35	.46	.42	.69	.57	.38
Al <sub>2</sub> O <sub>3</sub>	14.37	14.67	13.49	11.64	12.12	13.1	12.37	15.76	12.43	13.06	12.63
FeO	3.01	3.51	2.39	1.13	4.94	.84	5.13	1.2	4.23	3.50	1.6
Fe <sub>2</sub> O <sub>3</sub>	.93	1.18	1.12	.61	1.16	2.1	1.11	2.8	1.5	1.42	4.13
MnO	.06	.07	.05	.01	.02	.02	.14	.03	.03	.01	.03
MgO	.49	.45	.11	.01	.67	.33	.08	.56	.01	.17	.29
CaO	1.99	1.81	1.43	.45	.75	1.55	1.25	2.35	.94	.35	1.07
Na <sub>2</sub> O	3.71	3.81	2.99	3.32	2.79	4.16	3.99	3.71	1.92	1.08	6.63
K <sub>2</sub> O	5.67	5.61	5.79	5.22	6.50	4.01	4.91	4.91	8.34	9.64	.52
P <sub>2</sub> O <sub>5</sub>	.13	.12	.07	.03	.17	.07	.04	.13	.17	.12	.70
LOI	.40	.58	.5	.30	.25	.39	.21	.31	.23	.41	.10
Σ	99.86	100.41	100.00	99.20	99.80	100.3	99.67	99.98	100.50	99.38	100.47
Ba(ppm)	861	715	692	nd	1014	1249	160	nd	840	290	98
Rb (ppm)	161	148	182	188	214	178	190	nd	315	330	16
Sr (ppm)	209	240	115	132	303	211	20	nd	73	60	42
Y (ppm)	65	62	72	157	35	86	120	nd	66	75	117
Nb (ppm)	20	20	25	29	17	19	30	nd	18	23	27
Zr (ppm)	394	542	595	392	507	338	1230	nd	414	600	786

**Alaskitic and Leucogranitic Rocks.** The distribution of these distinctive rocks is shown in figure 5a, and their geochronology is summarized in table 1. An example of these rocks will be visited at Stop 12. They consist principally of pink quartz-mesoperthite gneiss commonly with magnetite as the only dark phase. A less voluminous, but important, trondhjemitic facies is also common and is commonly associated with low-Ti magnetite deposits in the unit. Granitic facies also occur within this group which, together, constitutes the Lyon Mt. Gneiss (Whitney and Olmsted, 1989). U-Pb zircon ages of 1080-1050 Ma for the alaskites are interpreted as dating emplacement, and, since this time interval corresponds to granulite facies metamorphism at ~25 km, the Lyon Mt. Gneiss is interpreted as intrusive (Chiarenzelli and McLelland, 1991). This is in contrast to Whitney and Olmsted (1989) who have interpreted the Lyon Mt. Gneiss as a metamorphosed series of altered acidic volcanics. This issue is discussed in detail in the text for Stop 12.

**Olivine Metagabbro.** Numerous bodies of tholeiitic metagabbros are scattered throughout the Adirondacks and are especially abundant near the anorthosite massifs with which they are coeval. Most of these bodies are coronites whose evolution has been discussed by McLelland and Whitney (1980) and which will be visited at Stops 2 and 17.

**Metasedimentary Rocks.** Within the southern and eastern Adirondacks the metasedimentary sequence is dominated by quartzites and metapelites with marbles being virtually absent. The major quartzite of the southern region is exceptionally pure and comprises an ~1000 m-thick unit. Of even greater extent, as well as thickness, are garnet-biotite-quartz-oligoclase  $\pm$  sillimanite gneisses (referred to as kinzigite) together with sheets, pods, and stringers of white, garnetiferous anatectite (Stop 2). The common presence of hercynitic spinel supports an anatectic origin for the leucosomes. Kinzigitic rocks grade into khondalites as sillimanite and garnet increase at the expense of biotite.

In contrast to the southern and eastern Adirondacks, the central and northern Adirondacks contain only sparse kinzigite, and metasediments are principally represented by synclinal keels of marble and calcisilicate with minor quartzite (Stop 15). It is possible that the change from carbonate to pelitic metasediments corresponds to an original shelf to deep water transition, now largely removed by intrusion, doming, and erosion.

## STRUCTURAL GEOLOGY

The Adirondacks region is one of intense ductile strain, essentially all of which must postdate the ca. 1150 Ma AMCG suite which is involved in each of the major phases of deformation, i.e., the regional strain is associated with the Ottawa Orogeny. Earlier Elzevirian fabric may be present and rotated into parallelism. Here we describe the structure of the southern Adirondacks which is best known and representative of the entire region. A complete set of references is given in McLelland and Isachsen (1986).

As shown in figures 2 and 11, the southern Adirondacks are underlain by very large folds. Four major phases of folding can be identified and their intersections produce the characteristic fold interference outcrop patterns of the region (figure 11). The earliest recognizable map-scale folds ( $F_1$ ) are exceptionally large isoclinal recumbent structures characterized by the Canada Lake, Little Moose Mt., and Wakely Mt. isoclines, whose axes trend E-W and plunge 10-15° about the horizontal. The Little Moose Mt. isocline is synformal and the other two are antiformal, and suspected to be anticlinal, but the lack of any facing criteria precludes any age assignments. All of these structures fold an earlier tectonic foliation consisting of flattened mineral grains of unknown age and origin. An axial planar cleavage is well developed in the Canada Lake isocline, particularly in the metapelitic rocks.

$F_2$ -folds of exceptionally large dimensions trend E-W across the region and have upright axial planes (figure 11). They are coaxial with the  $F_1$  folds suggesting that the earlier fold axes have been rotated into parallelism with  $F_2$  and that the current configurations of both fold sets may be the result of a common set of forces. An intense ribbon lineation defined by quartz and feldspar rods parallels the  $F_2$ -axes along the Piseco anticline, Gloversville syncline, and Glens Falls syncline and documents the high temperatures, ductile deformation and mylonitization that accompanied the formation of these folds. Large NNE trending upright folds ( $F_3$ ) define the

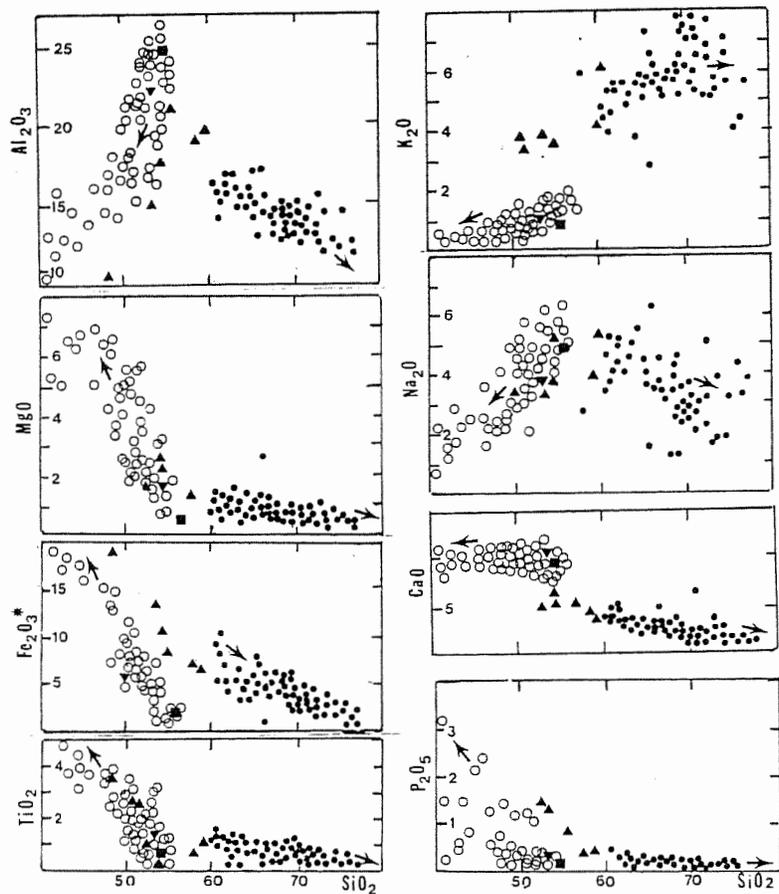


Figure 10. Harker variation diagrams for AMCG-rocks of the Marcy massif. Open circles = anorthositic suite, filled circles = granitoid suite, upright triangles = mixed rocks, inverted triangles = Whiteface facies, square = Marcy facies. Arrows indicate differentiation trends.

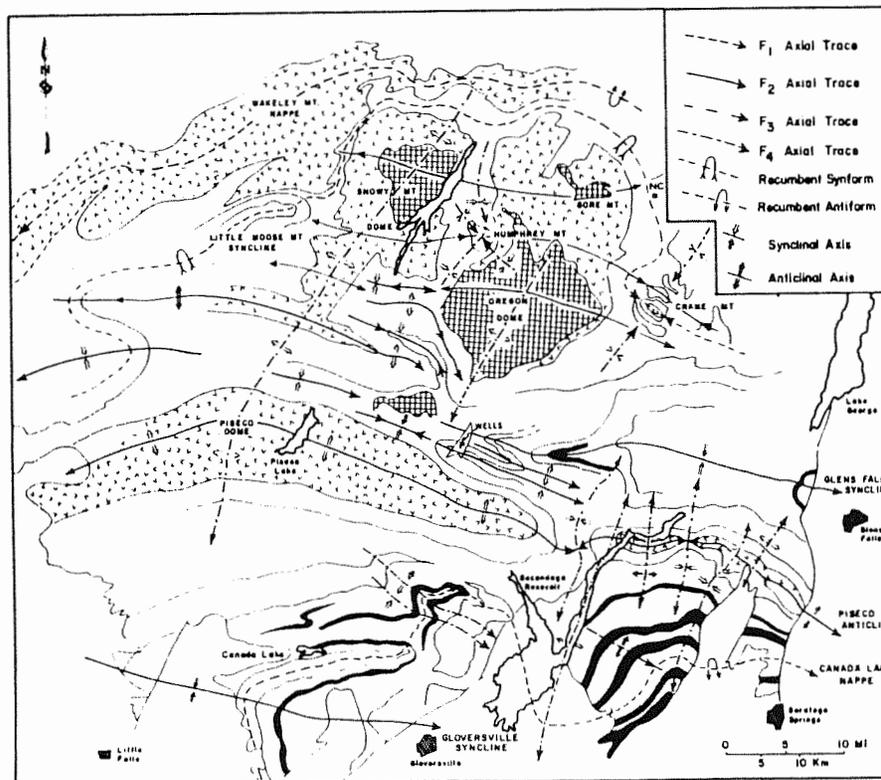


Figure 11. Fold axes within the southern and central Adirondacks. Designation of folds as synclines and anticlines is provisional, since facing directions are not yet known.

Snowy Mt. and Oregon domes (figure 11). Where the  $F_3$  folds intersect  $F_2$  axes structural domes (i.e., Piseco dome) and intervening saddles result. A late NW-trending fold set results in a few  $F_4$  folds between Canada Lake and Sacandaga Reservoir (figure 11).

Kinematic indicators (mostly feldspar tails) in the area suggest that the dominant displacement in the region involved motion in which the east side moved up and to the west (McLelland, 1984). In most instances this implies thrusting motion, however, displacement in the opposite sense has also been documented. This suggests that relative displacement may have taken place in both senses during formation of the indicators.

### METAMORPHISM

Figure 12 shows the well known pattern of paleoisotherms reported by Bohlen et al. (1985). paleotemperatures have been established largely on the basis of two-feldspar geothermometry but (Fe, Ti)-oxide methods have also been used and, locally, temperature-restrictive mineral assemblages have been employed (Valley et al., 1990). The bull's eye pattern of paleoisotherms, centered on the Marcy massif, is believed to result from late doming of the massif. Paleopressures show a similar bull's eye configuration with pressures of 7-8 kbar decreasing outward to 6-7 kbar away from the massif and reaching 5-6 kbar in the Lowlands (Bohlen et al., 1985). The P,T pattern of figure 12 is interpreted as reflecting peak metamorphic conditions, although microtextures suggest that some retrogression exists. Generally, the P,T conditions of the Adirondack are those of granulite facies metamorphism, and most commonly correspond to the hornblende-clinopyroxene-almandine subfacies of the high-pressure range of the granulite facies. These conditions must have been imposed during the Ottawaan Orogeny since they affect rocks as young as 1050 Ma. The identification of ca. 1050-1060 Ma metamorphic zircons by McLelland and Chiarenzelli (1990) fixes the time of peak metamorphic conditions and corresponds well with garnet and titanite U-Pb ages of ca. 1050-1000 Ma in the Highlands (Mezger et al., 1992). Rb-Sr whole rock isochron ages of ca. 1100-1000 Ma also reflect Ottawaan temperatures and fluids. Despite the high-grade, regional character of the Ottawaan Orogeny, the preservation of foliated garnet-sillimanite xenoliths in an  $1147 \pm 4$  Ma metagabbro (McLelland et al., 1987), and the report of several 1150 Ma U-Pb garnet ages (Mezger et al., 1992), reveals that earlier assemblages from the Elzevirian and AMCG metamorphic pulses managed to survive locally. The dehydrating effects of these high temperature events, as well as the anhydrous nature of the AMCG rocks themselves, are thought to be responsible for creating a water-poor terrane throughout the Adirondack Highlands prior to the Ottawaan Orogeny which appears to have proceeded under generally vapor-absent conditions (Valley et al., 1990).

The present depth to the Moho beneath the Adirondack Highlands is ~35 km (Katz, 1955; Hughes and Luetgert, 1992). Since metamorphic pressures of 7-8 kbar correspond to ~20-25 km depth of burial, it follows that during Ottawaan metamorphism the Adirondack region consisted of a double thickness of continental crust, and this portion of the Grenville orogen may have corresponded to a Himalayan-type collisional margin at 1050-1080 Ma. Bohlen et al. (1985) proposed a counterclockwise P-T-t path for the Ottawaan Orogeny, including an almost isobaric cooling history. If this is correct, the necessary magmatic component of heat may have been derived from 1130-1150 Ma gabbroic magmas that were ponded at the base of the lithosphere during the AMCG magmatism. Upward transfer of this heat by conduction would require ~80 Ma to reach the surface (Emslie and Hunt, 1990) and would, therefore, have been present in the crust during the height of the Ottawaan Orogeny. Granitic rocks of the ca. 1050 Ma Lyon Mt. Gneiss may have helped to transport this thermal energy.

- ILMENITE-SILLIMANITE-QUARTZ-GARNET-RUTILE
- FERROSILITE-FAYALITE-QUARTZ
- ◻ SPHALERITE-PYRRHOTITE-PYRITE
- FAYALITE-ANORTHITE-GARNET
- FERROSILITE-ANORTHITE-GARNET-QUARTZ
- ◐ ANORTHITE-GARNET-SILLIMANITE-QUARTZ
- ▲ KYANITE-SILLIMANITE
- △ AKERMANITE

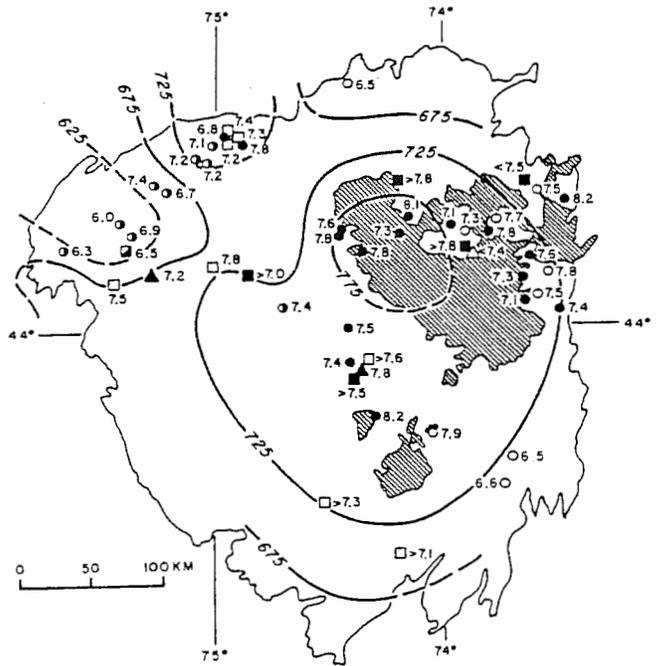


Figure 12. Peak metamorphic P,T conditions for the Adirondack MB. Isotherms are continued from feldspar and Fe-Ti thermometry and Pressures are estimated from the geobarometers given above and indicated on the map. (From Bohlen et al., 1985)

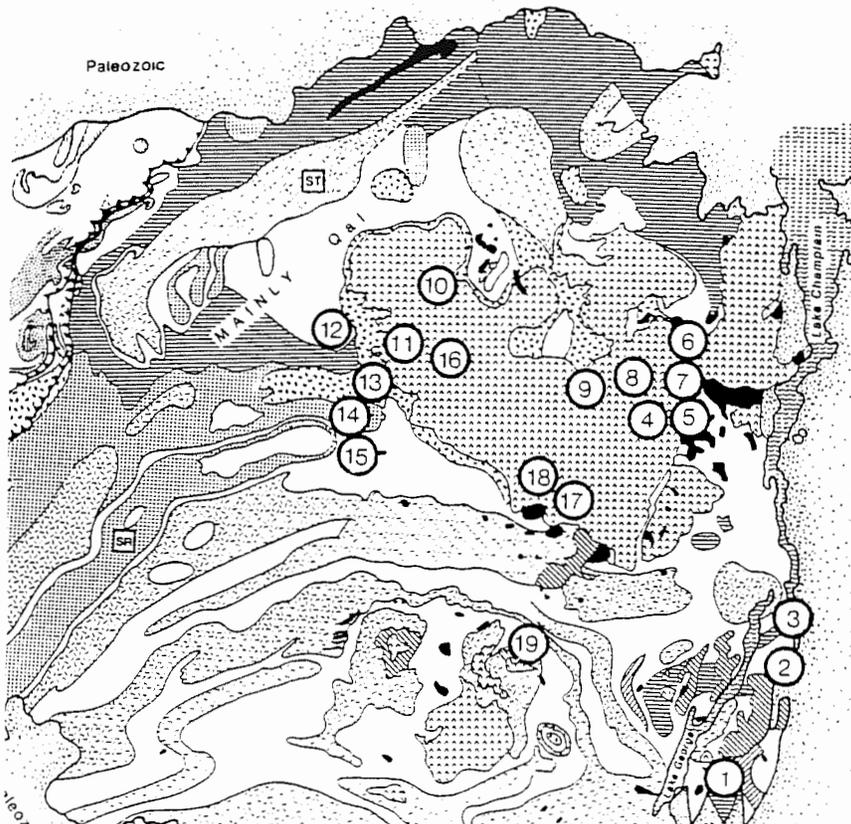


Figure 13. Locations of stops in the Adirondack Highlands. Symbols as in figure 2.

## ROAD LOG

## DAY 1

## Mileage

- 0.0 Intersection of Rt. 4 with NY Rt. 22 in Whitehall, NY. Head north on Rt. 22.
- 1.4 Enter long roadcut of calcsilicates, marble, and quartz-feldspar gneiss. Swede Mt. quartzite on hill to east.
- 2.5 Bridge over South Bay (Lake Champlain).
- 3.7 Park on shoulder of Rt. 22 near its northern junction with Blue Goose Tavern Rd.

**STOP 1. ROADCUTS OF CA. 1300 MA TONALITES (20 MINUTES)**

The grey, xenolith-bearing tonalitic gneiss in these roadcuts are typical of the ca. 1300 Ma tonalites that outcrop throughout the southern and eastern Highlands. Modally the rock consists of 20-25% quartz and 60-70% andesine plagioclase (AN<sub>35</sub>-AN<sub>40</sub>). Pyroxene and hornblende account for 5-10% of the mode. Invariably the tonalites exhibit irregular layering defined by cm-scale amphibolitic layers. Also present, and well exposed on the top of the eastern roadcut, are disrupted amphibolite layers which are interpreted as coeval mafic dikes. Zircon dating of this outcrop yields an age of 1329 ± 37 Ma which is similar to zircon ages for tonalites from the Green Mts., the southern Adirondacks, and the Elzevir Terrane (table 1).

Continue north on Rt. 22

- 6.9 Road to Hulett's Landing. Inequigranular charnockite in roadcuts. Continue north on Rt. 22.
- 9.1 Large roadcut of ferrogabbro with marble dike at the north end.
- 10.0 Southern intersection with Dresden Station Road. Continue straight (north) on Rt. 22.
- 10.2 Large roadcut of ferrogabbro containing xenocrysts and xenoliths of andesine and anorthosite.
- 10.6 Park on shoulder of highway alongside large roadcuts.

**STOP 2. METASEDIMENTS, METAGABBROS, AND FOLIATED XENOLITHS (30 MINUTES)**

At the southwest end of the roadcut is a late ferrogabbroic differentiate of the anorthosite suite containing green (epidotized) to chalk white grains (xenocrysts and xenoliths) of andesine plagioclase. Northward the ferrogabbro becomes increasingly deformed and passes into a mylonite, which is difficult to distinguish from mylonitized quartzitic rocks. North of the contact several isoclinal folds are developed in marble. The southeast side of the roadcut exposes an iron-rich olivine metagabbro which contains garnet coronas between olivine and plagioclase, as well as plagioclase clouded by green spinel. North of the olivine metagabbro is a sequence of coarse sillimanite-garnet-quartz-feldspar rocks with variable amounts of biotite and graphite. These have whole-rock chemical compositions similar to Proterozoic shales. They are interlayered with marbles containing rotated, angular fragments of quartz-feldspar-diopside assemblages.

The metagabbro exhibits a chill margin which survived metamorphic recrystallization, probably due to an absence of fluids. In places the metagabbro-khondalite contact truncates foliation in the khondalite indicating that the metagabbro was emplaced subsequent to fabric formation. On top of the roadcut a large xenolith of foliated garnet-sillimanite gneiss is clearly crosscut by the metagabbro (McLelland et al. 1987), thereby documenting a pre-gabbro (Elzevirian?) age for the fabric. Zircons from the gabbros yield an age of 1144 ± 7 Ma which also provides a minimum age for the foliated garnet-sillimanite gneiss. Note that additional fabric within the metagabbro is probably Ottawan in age.

Near the north end of the cut are exposures of a rusty sulfidic rock known as the Dixon Schist. Mineralogically, this rock consists of garnet + biotite + quartz + feldspar ± sillimanite. In addition it contains abundant pyrrhotite and graphite, the latter accounting for the schistose character of the unit. The Dixon Schist was the major source of graphite mined in the region during the early part of the century.

Continue north on Rt. 22

- 16.5 Overlook - steeply dipping, highly folded gneisses in roadcut to west.
- 17.1 Park on shoulder of road between long roadcuts.

**STOP 3. ANGULAR UNCONFORMITY BETWEEN PROTEROZOIC GRENVILLE GNEISS AND BASAL CONGLOMERATES OF THE UPPER CAMBRIAN POTSDAM SANDSTONE (10 MINUTES)**

The deeply weathered gneisses have nearly vertical dips. The age of the weathering is uncertain but appears to post-date the unconformity and may be due to alteration by groundwater.

- Continue north on Rt. 22.
- 30.3 Intersection of Rt. 22 and Rt. 74 north of Ticonderoga. Proceed west on Rt. 74 to Schroon Lake.
- 49.8 Intersection of Rt. 22 with Rt. 87(Northway). Proceed north on Rt. 87.
- 63.8 Exit onto Rt. 73 west and proceed through Keene, Keene Valley, and Cascade Lakes.
- 92.6 Junction with Heart Lake Rd. Turn south to Adirondack Loj.

**END OF DAY 1**

**DAY 2**

**Mileage**

- 0.0 Junction Heart Lake Rd. and Rt. 73. Turn right (east) on Rt. 73.
- 6.2 Cascade Lakes. Cascade slide directly across lakes and above high waterfall.
- 12.2 Junction of Rts. 73 and 9N in Keene. Continue south (right) towards Keene Valley on Rt. 73.
- 13.6 Junction with 9N to Elizabethtown. Continue east on Rt. 73.
- 21.2 Park at trailhead of Roaring Brook Trail on Giant Mt.

**STOP 4. ROARING BROOK ON GIANT MT. (4 HOURS)**

The valley of Roaring Brook provides some of the finest exposures of AMCG rocks in the Adirondack region. The nature of the geology exposed in Roaring Brook suggests that the events that took place here were uncommon, i.e., dikes of several different compositions and/or generations have intruded parallel to the stream valley and a variety of AMCG rock types are represented. A plausible interpretation of the association is that Roaring Brook represents a zone of weakness that has served repeatedly as a magma conduit. From 1400' to 2000' the brook is underlain by a variety of anorthositic and gabbroic rocks. Generally the gabbroic facies contains two pyroxenes but both augite gabbros and norites are present. Near the lip of the high waterfall (note diabase dike in stream bed) a broad (~3m) somewhat irregular, monzonite dike occupies the northwest side of the brook. The monzonite contains augite and blue-grey micropertite and closely resembles the anorthosite. The dike is enriched in pyroxene along its margins and crosscuts a N20-40W, 60-80N foliation in the anorthosite. Inspection of the anorthosite reveals that subophitic pyroxenes have not been deformed, and therefore, the foliation was imposed prior to complete solidification of the magma.

Proceeding upstream from the lip of the high falls, a 3-10 meter-high cliff crosses the brook and results in a second waterfall. At the base of the waterfall there is exposed a dark, meter-wide, eroded dike of pyroxenite. Downstream this dike splays and pinches out, but upstream it defines a pronounced erosional channel in the cliff and then continues upstream for another 30-40 m until it is lost beneath cover. In fact, the dike is discontinuous and is intermittently exposed for almost a kilometer. The chemical composition of the dike is given in McLelland and Chiarenzelli (1990). The dike is clearly intrusive and was emplaced after the anorthositic rocks acquired their foliation. In several instances xenoliths of anorthosite occur within the dike. Conversely, as noted by deWaard (1979, p. 2072), the anorthositic rocks crosscut the dike at several places, and soft contacts between the dike and country rock are not uncommon, indicating that the rocks are coeval. Because of the dike's composition, it seems unlikely that it was intruded as a liquid. This is consistent with the presence of cumulate and adcumulate textures. A possible mechanism would be the downward draining of a cumulate layer into an underlying fracture developed in cooling anorthosite. The very high Mg-numbers of the pyroxenes in the dike (Opx-65, Cpx-75) suggest that the cumulate formed early in the fractionation history. The smooth outcrop surfaces surrounding the pyroxenite dike are dominated by gabbroic anorthosite transitional to gabbro and provide excellent examples of the composite nature of the anorthositic suite. The oldest anorthosite facies recognizable are coarse grained rafts of blue-grey andesine anorthosite corresponding to the Marcy facies. They occur as xenoliths within a subophitic, medium-grained two-pyroxene gabbro, or anorthositic gabbro, which

TABLE 3

## Major Element Analyses of Roaring Brook Samples

	1	2	3	4	5	6	7	8	9	10
Sample	Marcy-type Anorthosite <sup>a</sup>	Whitface Anorthosite <sup>a</sup>	Anorthosite	Leuconorite	Fine Grained Gabbroic Anorthosite		Gabbro	Small Mafic Inclusions	Orthopyroxenite Dike	Jotunite
			7-19-89-27	7-19-89-20	7-19-89-21	7-19-89-22	7-19-89-23	7-19-89-25		
SiO <sub>2</sub>	54.54	53.54	54.97	52.55	53.75	53.70	53.19	53.29	50.82	47.16
TiO <sub>2</sub>	0.67	0.72	.30	.49	.23	.21	.43	.30	1.27	2.20
Al <sub>2</sub> O <sub>3</sub>	25.61	22.50	25.09	21.48	25.23	26.54	21.66	23.24	4.7	17.23
Fe <sub>2</sub> O <sub>3</sub>	1.00	1.26	1.05	1.13	2.18	1.05	4.07	3.31	2.3	2.75
FeO	1.26	4.14	1.19	3.01	---	---	---	---	14.2	9.24
MnO	0.02	0.07	.02	.06	.02	.01	.06	.04	0.29	0.15
MgO	1.03	2.21	.68	3.96	1.84	.71	3.9	2.76	16.52	2.71
CaO	9.92	10.12	9.95	11.79	10.93	10.40	11.43	11.82	8.44	9.04
Na <sub>2</sub> O	4.53	3.70	4.7	3.84	4.26	4.63	3.78	3.80	0.71	6.61
K <sub>2</sub> O	1.01	1.19	1.46	.66	.75	.80	.65	.71	0.20	2.27
P <sub>2</sub> O <sub>5</sub>	0.09	0.13	.07	.09	.06	.06	.09	.04	0.14	0.59
H <sub>2</sub> O	<u>0.55</u>	<u>0.12</u>	<u>.57</u>	<u>.48</u>	<u>.67</u>	<u>.49</u>	<u>.38</u>	<u>.53</u>	<u>0</u>	<u>0</u>
Total	100.17	100.00	99.86	99.95	99.95	98.57	99.63	99.82	99.59	99.70

	11	12	13	14	15	16	17	18	19	20
Sample	Mafic Mangerite	Mangerite	Charnockite	Garnet-Plagioclase Xenolith	Enclaves					
					Cumulates			Other		
				7-18-89-3	7-18-89-2	7-18-89-7	7-18-89-21A	7-18-89-21B	7-18-89-4	7-18-89-6
SiO <sub>2</sub>	50.05	56.45	62.70	49.23	55.74	52.90	52.75	53.83	57.62	55.28
TiO <sub>2</sub>	1.47	1.50	0.44	2.63	.17	.70	.65	.84	.73	.68
Al <sub>2</sub> O <sub>3</sub>	16.08	15.88	18.41	16.76	2.87	5.50	10.17	12.96	13.54	12.54
Fe <sub>2</sub> O <sub>3</sub>	2.53	2.60	0.63	2.73	5.22	13.37	8.78	11.40	6.76	7.73
FeO	9.06	8.13	2.37	12.68	---	---	---	---	---	---
MnO	0.20	0.17	0.05	.43	.83	.38	.24	.25	.11	.13
MgO	3.21	1.06	0.40	2.73	14.39	8.18	7.61	4.83	5.10	5.56
CaO	8.20	4.39	2.49	7.01	18.51	15.21	14.38	10.66	8.80	10.06
Na <sub>2</sub> O	4.3	3.56	3.85	2.54	.62	1.05	3.00	3.37	2.87	3.03
K <sub>2</sub> O	4.33	5.50	7.59	2.51	.9	1.2	.88	1.34	4.16	3.01
P <sub>2</sub> O <sub>5</sub>	0.57	0.38	0.15	.77	.05	.17	.15	.23	.16	.17
H <sub>2</sub> O	<u>0</u>	<u>Nd</u>	<u>Nd</u>	<u>.01</u>	<u>.10</u>	<u>.10</u>	<u>.38</u>	<u>.30</u>	<u>---</u>	<u>.70</u>
Total	100.09	99.62	00.08	100.02	100.21	98.82	98.99	100.01	99.85	98.89

<sup>a</sup>Buddington (1939)

locally, grades into a noritic facies. This, in turn, is crosscut by a fine grained gabbroic anorthosite similar to the Whiteface facies. Elsewhere on the outcrop, the time sequence is partially reversed and the fine grained anorthosite is crosscut by gabbroic to noritic facies; however, in all instances, the rafts of coarse, blue-grey anorthosite appear to be oldest rock. An apparently older, fine grained gabbro occurs as xenoliths within fine-grained gabbroic anorthosite near the upper edge of the cliff that defines this level of Roaring Brook.

Returning to the trail and proceeding uphill, we cross Roaring Brook and ascend the summit trail to the 2260' (689m) level. Here we leave the trail and descend to water-smoothed pavement outcrops in the brook valley. The outcrops expose a spectacular intrusion breccia consisting of rounded and angular blocks (10-30 cm on average) which include coarse, white anorthosite but consist mostly of grey to black, medium to fine-grained, granular pyroxene-feldspar assemblages. These are set in a medium grained groundmass ranging in composition from gabbroic anorthosite to garnetiferous mangerite and ferrogabbro. These matrix rock types are highly mingled and difficult to separate without the aid of outcrop staining. Although the intrusion breccia is dominated by dark enclaves, white inclusions of anorthosite are easily recognizable. Less easy to recognize, however, are pink, somewhat glassy, blocks that might be mistaken for garnetiferous quartzite but turn out to be fine-grained, oxide-rich garnetiferous anorthosite or gabbro. Xenocrysts of blue-grey andesine occur within each type of enclave and attest to their igneous origin.

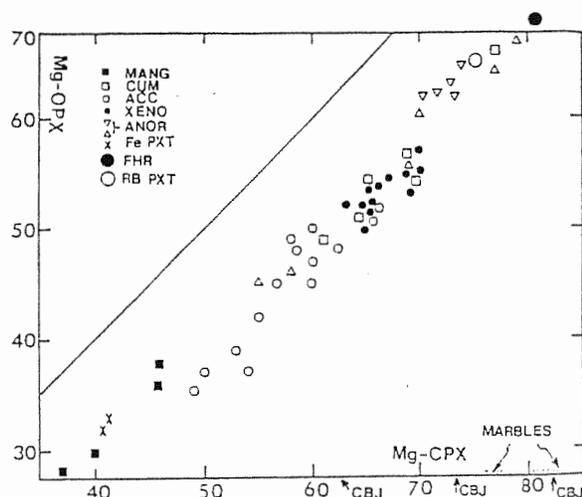


Figure 14. Plot of mole fraction of Mg ( $Mg/Mg + Fe$ ) for opx vs. that for cpx for anorthositic (ANOR) and mangeritic (MANG) rocks for the Marcy massif. Also: CUM=cumulate enclaves, Roaring Brook; ACC=pyroxene pairs with acicular opx, Roaring Brook; RB PXT=Roaring Brook pyroxenite dike; Fe PXT=iron-rich pyroxenite, Roaring Brook; XENO=dioritic enclaves, Roaring Brook; FAR=Forest Home Road; CBJ=pyroxenite dikes in Ausable River, Jay.

A number of dark inclusions exhibit narrow ( $\sim .5$  cm), light colored layers. These have traditionally been interpreted as primary in origin and led to the assignment of a metasedimentary origin to the rocks (see McLelland et al., 1987 for a review). However, staining of slabs indicate that the layers consist essentially of mesoperthitic syenites and mangerite. The mafic pyroxene-plagioclase layers are essentially gabbroic to dioritic in composition with subequal amounts of pyroxene ( $X_{Mg}^{OPX} = .35-.55$ ,  $X_{Mg}^{CPX} = .50-.70$ ) and plagioclase ( $AN_{25-40}$ ). Igneous textures, including acicular, comb layered orthopyroxene, are common within these layers. On the basis of these textures and compositions, we interpret these layered inclusions as igneous enclaves mixed with sub-parallel veins of syenite and mangerite and incorporated into the mixtures of country rock magmas now constituting the breccia groundmass. The acicular orthopyroxenes are identical to comb textured pyroxenes in demonstrably igneous rocks such as orbicular granites and provide compelling evidence for the igneous nature of these rocks. Note that the commonly "slumped" configuration of the layering in these rocks is also consistent with their evolution from magmas. The layering is interpreted as the result of commingling between viscous magmas (Ottino, 1989) in a magma conduit.

The non-layered inclusions in the Roaring Brook intrusion breccia are also interpreted as igneous. The most compelling evidence for this is the presence of acicular orthopyroxenes within these enclaves, but other observations include rock and mineral compositions which do not correspond closely to possible sedimentary precursors such as calcisilicates, but in many cases are similar to dioritic rocks. Note, for example, that in

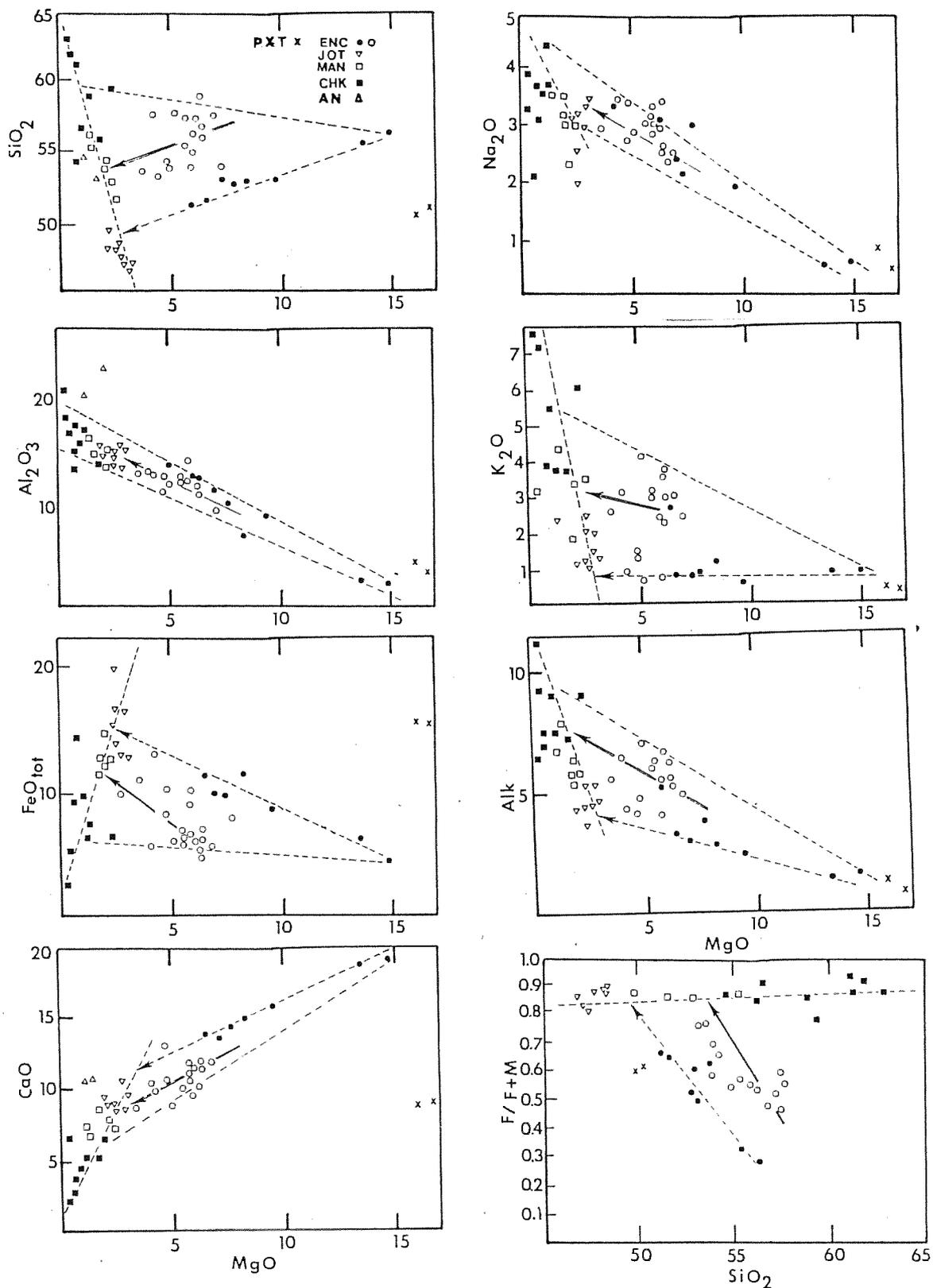


Figure 15. Binary variation diagrams for enclaves and matrix rocks in Roaring Brook intrusion breccia. PXT=pyroxenite, ENC=enclaves (closed for cumulates), JOT=jotunites, MAN=mangerite, CHK=charnockite, AN=anorthosite.

Adirondack calcisilicates the pyroxenes consist almost solely of clinopyroxene and that these are generally more Mg-rich than the clinopyroxene in the Roaring Brook enclaves. In addition, the enclaves never contain calcite, quartz, wollastonite, garnet, graphite, or phlogopite; all of which are common in calcisilicates. Finally, we note that "soft" and lobate contacts between the enclaves and country rocks similar to those typical of coeval, commingled magmas such as proposed here. Indeed there exist several examples in the brook where dark enclaves can be seen in the process of formation as the result of disruption of masses of dioritic material by felsic country rock. The breccia also contains clinopyroxene-rich cumulates interpreted as xenoliths caught up and disrupted by ascending AMCG magmas. Consistent with this interpretation is the 130 ppb platinum concentration of 7-18-89-2. This, and other possible cumulate enclaves, are designated as filled circles on figures 14 and 15. Note that these enclaves do not contain acicular pyroxene and consistently exhibit large, interlocking pyroxene grains with minor interstitial plagioclase.

In order to better constrain the evolution of the enclaves, we have plotted their chemistry in binary variation diagrams with MgO chosen as a common variable (figure 15). The most important conclusion emerging from these plots is that magma-mixing appears to have taken place - both with regard to the enclaves and with regard to the breccia matrix. Several of the variation diagrams, and especially SiO<sub>2</sub> vs. MgO, show a pronounced linear plot of country rock jotunite, mangerite, and charnockite. We interpret this to mean that the intermediate rock types represent mixtures of the other magmas. Here it must be stressed that the jotunites are not the most primitive of the felsic rocks; they are likely to be the result, despite their position in these plots, of mixing of ferrodiorite and mangeritic magma, but the former plot outside the range of the diagrams.

- From parking area turn left (southeast) onto Rt. 73.
- 27.0 Junction with Rt. 9, turn left (north) on Rt. 9 towards Elizabethtown.
- 33.4 Park on shoulder of Rt. 9 across from large roadcuts

#### **STOP 5. LONG OUTCROP OF ISOCLINALLY FOLDED ANORTHOSITIC AND GRANITIC ROCKS SOUTH OF ELIZABETHTOWN ON RT. 9. (20 MINUTES)**

Although most rocks in this outcrop are highly altered, it affords the opportunity to see the effect of Ottawa deformation on AMCG rocks. Most of the outcrop consists of somewhat gneissic anorthositic gabbro (table 4).

Note the parallel compositional layering in the outcrop. Locally amphibolite layers are crosscut and disrupted by anorthosite, indicating that the anorthosite intruded into the amphibolite probably in sheets and parallelism has been further accentuated by strain. Light colored granitic layers must also have been intruded in parallel sheets. None of this layering is primary.

- 34.7 Enter Elizabethtown. Junction Rts. 9 and 9N. Continue north on Rt. 9N.
- 39.7 Village of Lewis. Turn west on Wells Road.
- 40.3 Park at Lewis Wollastonite Mine

#### **STOP 6. LEWIS WOLLASTONITE MINE (30 MINUTES)**

The Lewis wollastonite mine is about 15 km SW of the Willsboro mine, in what appears to be the same belt of metasedimentary rocks. Exposure in the vicinity of the mine is poor, so the relationship of the ore to the surrounding rocks is not as well known as at Willsboro. The mine's ores, however, provide an excellent opportunity to examine the wollastonite unit itself. On the west side of the pit, the ore is overlain by charnockitic gneiss, while to the north and east metanorthosite, contaminated by a variety of inclusions, is exposed above the ore. Based on temporary exposures within the mine, gabbroic anorthosite gneiss appears to underlie the ore body. Complete references may be found in Valley et al. (1985).

The mineralogy of the ore is identical to that of Willsboro deposit, with modes of 60-70% wollastonite, 8-12% clinopyroxene, and 15-30% garnet. Pyroxene compositions range from Wo<sub>33</sub>En<sub>47</sub>Fs<sub>20</sub> to Wo<sub>33</sub>En<sub>27</sub>Fs<sub>40</sub>; the garnets are grandites ranging from Gr<sub>12</sub> to Gr<sub>72</sub>. The garnet in this setting may be the result of reaction between plagioclase and wollastonite during the granulite facies metamorphism. Values of δ<sup>18</sup>O in the wollastonite are as low as -1.2 permil, similar to the Willsboro deposit.

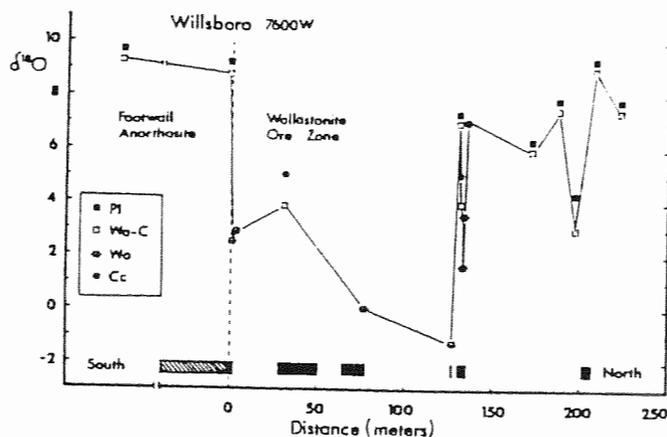


Figure 16. Oxygen isotopic composition (in permil SMOW) of minerals along a traverse perpendicular to the ore zone at the Willsboro wollastonite mine. Note the steep gradients at contacts. (After Valley and O'Neil, 1982.)

Valley et al. (1985) have reported anomalously low  $\delta^{18}\text{O}$  (-1.3 to 3.1; up to 20 permil lower than typical Adirondack marbles) in the wollastonite ore within 125 m of the anorthosite contacts, as well as extremely sharp  $\delta^{18}\text{O}$  gradients between the wollastonite and the surrounding rocks (figure 42). The low  $\delta^{18}\text{O}$  values cannot be explained solely by devolatilization reactions (Valley et al., 1985) and probably result from deep circulation of heated meteoric waters along fractures at the time of anorthosite intrusion. Because such fluids would be at hydrostatic pressure, they could not penetrate a ductile metamorphic environment where fluids are at lithostatic pressure. This suggests that skarn formation occurred at shallow depths (<10 km) relative to granulite facies metamorphism. The origin of the sharp gradients is enigmatic, but their preservation through granulite facies metamorphism indicates that there was no significant fluid movement across strike during the metamorphism.

Turn around and return to Rt. 9 via Wells Road.

40.9 Junction Rt. 9. Turn right (south).

45.9 Junction Rts. 9 and 9N, Elizabethtown. Turn right (west) on Rt. 9.

47.0 Park in pull-off on right side of Rt. 9N

#### STOP 7. WOOLEN MILL GABBRO (30 MINUTES)

Park on the right side of the road opposite high roadcut on left. The cut shows metanorthosite intruded by a dark, fine-grained rock, referred as the "Woolen Mill Gabbro". It is a clinopyroxene-garnet-oligoclase granulite with considerable opaque oxides and apatite, and minor K feldspar and quartz. It contains a few large, uncrushed andesine xenocrysts, probably derived from the host anorthosite. The texture is that of a granulite, but the xenocrysts have apparently escaped recrystallization or grain-size reduction, even along their margins. This peculiar situation may be explained by static recrystallization of an initially fine-grained intrusive rock. The composition of rock (table 7) is that of a somewhat  $\text{K}_2\text{O}$  rich (1.20 wt%) ferrogabbro of the type common in the Adirondack Highlands, especially near magnetite-ilmenite concentrations. It also is found associated with anorthosite at stops 7 and 8 and is commonly present as disrupting material in block structure. Woolen Mill gabbro may represent gabbroic anorthosite magma enriched in mafic components by separation of cumulus plagioclase as suggested by mixing calculations (Ashwal, 1978). This is the type locality for deWaard's (1965) clinopyroxene-almandine subfacies of the granulite facies.

Cross the road and examine the outcrops in the stream bed. At the west end of the stream exposures, Woolen Mill gabbro clearly crosscuts anorthosite, and veins and dikes of the gabbro extend into the anorthosite. Within the anorthosite there is well-developed "block structure" where several types of anorthosite have undergone brittle fracture before being intruded by thin dikes or veins of mafic as well as felsic material. Some of these veins are identical to the mafic granulite in the roadcut (and at the west end of the stream exposure) and are part of the anorthosite suite. Some of the disrupting material is anorthositic gabbro more commonly associated with the anorthosite as on Giant Mt. or Lake Clear. In addition, a variable amount of granitic

material is present in many of the veins as revealed by staining. The relationships here suggest formation of a plagioclase-rich cumulate, which was then fractured and intruded by a later mafic differentiate. This apparently brittle behavior suggests a relatively shallow depth of intrusion. Notice also the very large (up to 10 cm) giant orthopyroxenes that occur in the anorthosite, especially near the contact with Woolen Mill gabbro.

Continue west on Rt. 9N.

56.0 Turn left (east) at intersection of Rts. 9N and 73.

56.3 Turn right onto Hulls Falls Road.

57.3 Park on west end of bridge at Hulls Falls.

#### STOP 8. KEENE GNEISS AT HULLS FALLS (20 MINUTES)

Here the East Branch of the Ausable River has exposed a typical section of hybrid anorthosite-mangerite-charnockite gneiss known as Keene Gneiss (Buddington, 1939). The water smoothed outcrops consist of irregular, garnetiferous interlayers of plagioclase-rich and micropertthite-rich gneisses with little actual mixing between them. The whole rock chemistry of several of the granitic fractions is given below. Within the granitic facies of Keene Gneiss, blue-grey xenocrysts of andesine ( $An_{48}$ ) are readily visible and commonly exhibit lighter-colored reaction rims of more sodic plagioclase similar in composition to that in the mangeritic host ( $AN_{30}$ ).

TABLE 12

	(1)	(2)	(3)
SiO <sub>2</sub>	51.63	55.02	58.90
TiO <sub>2</sub>	3.1	1.6	1.66
Al <sub>2</sub> O <sub>3</sub>	14.23	13.66	14.27
Fe <sub>2</sub> O <sub>3</sub>	2.1	2.12	1.22
FeO	13.50	14.06	8.18
MnO	.16	.29	.14
MgO	2.63	.75	2.15
CaO	6.5	4.92	5.57
Na <sub>2</sub> O	2.67	3.13	2.43
K <sub>2</sub> O	2.41	3.93	3.57
P <sub>2</sub> O <sub>5</sub>	.57	.52	.46
H <sub>2</sub> O	.07	.08	.11
Total	99.57	98.64	98.68

It is difficult, without staining, to distinguish the granitic and anorthositic facies of Keene Gneiss; however the anorthositic facies tend to be more mafic, and the presence of quartz is diagnostic of quartz. As Keene Gneiss is followed across strike, and towards anorthosite, the granitic fraction becomes increasingly rich in andesine xenocrysts and xenoliths of anorthosite. Ultimately the granitic fraction constitutes no more than an interstitial filling between andesine grains and the gradation into anorthosite is essentially complete. The origin of Keene Gneiss seems quite clearly to be the result of commingling and hybridization between anorthositic, mangeritic, and charnockitic magmas. The high iron, titanium, and magnesium concentrations of the granitic fractions may be the result of mixing with late mafic liquids from the anorthosite. Such mixing may be responsible for the zone of mafic mangerite, gradational

into jotunite, within the Tupper-Saranac complex that Buddington (1939) and Davis (1970) referred to as transition rock. In such instances magma mixing would be more complete and the resultant rock would be more homogenous than Keene Gneiss.

Continue west on Hulls Falls Rd.

58.5 Junction Rt. 73. Turn (west) left towards Cascade Lakes.

64.5 Park in parking area on northwest side of Rt. 73.

#### STOP 9. CASCADE SLIDE XENOLITH\* FROM PARKING AREA (10 MINUTES)

In the stream bed above the falls there are several xenoliths and schlieren of marble  $\pm$  calcisilicate, together with light colored granitic layers and amphibolite. Note that the good parallel layering cannot be primary and demonstrates how sheet-like intrusion and tectonism can result in excellent layering. The fact that anorthositic rocks locally crosscut amphibolite indicates that they are intrusive surrounded by anorthosite. The largest of

\* Authored by John Valley from International Geological Congress Guidebook, 1989.

these bodies measures approximately 30 x 200 m in exposure, is compositionally zoned, and contains several unusual minerals. Most notably, the xenolith contains sanidine facies index minerals wollastonite, monticellite ( $\text{Mo}_{92-89}$ ), and akermanite as well as cuspidine, harkerite, vesuvianite, and wilkeite (Valley and Essene, 1980). Other minerals present include tremolite, garnet ( $\text{Gr}_{80-16}$ ,  $\text{And}_{80-15}$ ), spinel ( $\text{Mg}_{73}$ ), calcite, forsterite ( $\text{Fo}_{92}$ ), magnetite, clinopyroxene scapolite ( $\text{Me}_{78-50}$ ), quartz, and sphene.

Field relations, deformation and geochronology make it clear that these marble bodies were entrained within the anorthositic magma before the peak of granulite facies metamorphism. The exact timing of intrusion vs. regional metamorphism is still a matter of debate. Pre- rather than syn-metamorphic intrusion is favored, but in either case it is certain that both anorthosite and marble experienced the pressures and temperatures of granulite facies metamorphism (Valley et al., 1990). Thus, the mineralogy of these bodies may be used to study the P-T fluid conditions of granulite facies metamorphism. The origin of these minerals, which we believe was at low P and high T, is irrelevant in this regard because of the pervasive nature of the granulite overprint.

Several factors combine to make the Cascade Slide xenolith an unusually advantageous locality for fluid studies: 1) On a scale of 0.1 km the field relationships are relatively clear; a complex calcisilicate body is surrounded by anorthosite, so that any fluids infiltrating the xenolith must have passed through the anorthosite. 2) Mineral assemblages in the calcisilicates include many that either buffer or restrict  $f\text{H}_2\text{O}$  and  $f\text{CO}_2$ . 3) There is a large contrast in  $\delta^{18}\text{O}$  values between anorthosite ( $\delta^{18}\text{O}=9.7$  permil; Taylor, 1969; Morrison and Valley, 1988) and the core of the xenolith ( $\delta^{18}\text{O}$  up to 26.1). Thus the unusual character of this body makes it a sensitive monitor of fluid history.

Solid-solid mineral reactions at Cascade Slide indicate that P and T attained at least 7.4 kbar and 750°C, respectively (Valley and Essene, 1980; Bohlen and others, 1985). Valley and Essene (1980b) describe assemblages of akermanite + monticellite + wollastonite with equilibrium metamorphic textures as well as symplectic intergrowths of wollastonite and monticellite. At these temperatures and pressures, the presence of wollastonite, monticellite or akermanite requires that  $\log f\text{CO}_2$  be  $\leq 4.35$ ,  $\leq 3.32$ , or  $\leq 2.5$  respectively.

Further evidence that granulite facies fluid infiltration has not been important at Cascade Slide comes from oxygen isotopes (Valley and O'Neil, 1984; Valley, 1985). Any fluids ( $\text{H}_2\text{O}$  or  $\text{CO}_2$ ) passing through the xenolith would first have passed through the surrounding anorthosite ( $\delta^{18}\text{O}=9.7$ ). Subsequent exchange with the calcisilicates ( $\delta^{18}\text{O}=17.6$  to 26.1) would tend to homogenize this large premetamorphic difference with the result that  $\delta^{18}\text{O}$  in the xenolith would be reduced. The highest values of  $\delta^{18}\text{O}$  (26.1) in monticellite marble are thus very restrictive to theories of fluid infiltration and require fluid/rock  $< 0.1$ .

Three lines of evidence argue against the presence of fluid during the granulite facies metamorphism at Cascade Slide: 1) Assemblages of monticellite + forsterite + diopside + calcite + spinel plot in the fluid-absent field, including that if a fluid had existed,  $\text{PH}_2\text{O} + \text{PCO}_2 \leq 0.4$  kbar. 2) The large gradients in buffered values of  $f\text{CO}_2$  across the body and the fragile nature of the buffering assemblages would have been erased by  $\text{CO}_2$  infiltration, even by quantities as low as  $\text{CO}_2/\text{rock} = 0.001$ . 3) The preservation of high  $\delta^{18}\text{O}$  in the core of the xenolith and the sharp gradients of up to 18 permil/15 m would all have been homogenized if either  $\text{H}_2\text{O}$  or  $\text{CO}_2$  had infiltrated the xenolith in quantities greater than fluid/rock - 0.1.

Continue west on Rt. 73.

70.7 Junction Rt. 73 and Heart Lake Road. Turn left (south) for Adirondack Loj.

## END OF DAY 2

### DAY 3

#### Mileage

- Junction Heart Lake Rd. and Rt. 73. Turn left (west) on Rt. 73.
- 1.2 Olympic ski jump. Bear left on Old Military Road.
- 4.9 Junction with Rt. 86. Turn west (left).
- 11.3 Stop light. Continue on Rt. 86 through the Village of Saranac Lake. Continue north on Rt. 86 past hospital and out of Saranac Lake Village.
- 17.1 At junction with Rt. 186 turn west (left) and continue west.
- 19.3 Junction of Rts. 30 and 186. Turn north (right) on Rt. 30.
- 22.1 Park along low roadcuts on east side of road.

**STOP 10. EAST SIDE OF LAKE CLEAR. ANORTHOSITIC AND RELATED MAFIC ROCKS. (1 HOUR)**

This stop consists of a series of low, ledge-like outcrops along the northeast side of Rt. 30. The exposure provides excellent examples of block structure and crosscutting relationships between a variety of anorthositic facies, including Marcy-type, foliated gabbroic anorthosite, gabbro, and mafic ferrogabbro. Outcrop relationships indicate that the anorthositic rocks have been intruded and disrupted by gabbro and ferrogabbro (Cpx > Opx) which exhibits ophitic to subophitic texture and contains both small plagioclase grains (~AN<sub>40</sub>) as well as large (up to 10 cm) grains of blue-grey andesine (~AN<sub>48</sub>) believed to have been plucked from the anorthosite. The mafic content of the gabbro varies from about 25% to 75% and at the mafic-rich end it grades into Fe-Ti oxide-bearing ferrogabbro. At the southernmost end of the outcrops there occurs block structure of coarse, Marcy anorthosite with blue-grey andesine disrupted by gabbroic and ferrogabbroic material which exhibits "lobate" contacts with the anorthosite, suggesting coeval magmatism. Narrow, crosscutting veins of sulfide-bearing ferrogabbro are present in the outcrop and some parallel the road along shear zones which are associated with local mylonitization.

TABLE 4

	Woolen Mill	Tahawus Ore	Tahawus Ore Gabbro	Tahawus Olivine Metagabbro	Anorthosite Gabbro Rt. 9 Elizabethtown	FHRD 3 Cumulate	PXT Dike At Stop 13
SiO <sub>2</sub>	45.6	4.59	39.04	47.62	51.20	35.98	50.92
TiO <sub>2</sub>	3.5	18.58	6.78	.82	2.61	12.55	1.47
Al <sub>2</sub> O <sub>3</sub>	14.23	5.48	13.09	18.69	17.67	5.82	3.58
Fe <sub>2</sub> O <sub>3</sub>	18.42	66.37	19.09	11.40	10.43	25.81	17.73
MnO	.27	.28	.24	.14	.15	.30	
MgO	3.07	3.39	5.31	8.85	2.53	8.74	12.65
CaO	9.25	.31	9.77	8.29	9.37	10.84	9.57
Na <sub>2</sub> O	2.63	.22	2.02	2.76	3.10	.80	0.56
K <sub>2</sub> O	.79	.09	.66	.41	1.72	.07	0.17
P <sub>2</sub> O <sub>5</sub>	1.26	.0	.23	.12	1.02	.04	0.15
H <sub>2</sub> O	<u>.70</u>	<u>.1</u>	<u>.03</u>	<u>.2</u>	<u>.27</u>	<u>-.95</u>	<u>0</u>
Total	100.23	100.06	96.26	99.7	100.06	100.00	99.87

	Gabbroic Anorthosite Fine Grained*	Anorthosite Fine Grained*	Anorthositic Gabbro M1 Candidate LC14	Gabbro Disrupts Anorthosite LC13	Ferrogabbro Dikes LC4=LC2	Mangerite Crosscuts HK19
SiO <sub>2</sub>	52.30	54.26	52.07	47.09	44.13	59.27
TiO <sub>2</sub>	.29	.18	1.65	4.99	6.73	.60
Al <sub>2</sub> O <sub>3</sub>	22.58	26.51	19.84	13.33	9.07	20.30
Fe <sub>2</sub> O <sub>3</sub>	4.22	2.18	7.85	16.06	17.03	3.97
MnO	.04	0	.06	.24	.32	.05
MgO	2.46	.56	1.86	4.50	6.89	.95
CaO	10.22	9.38	8.24	10.16	11.97	5.05
Na <sub>2</sub> O	4.42	4.90	3.8	2.26	1.62	4.91
K <sub>2</sub> O	.68	1.10	1.92	.92	.35	4.47
P <sub>2</sub> O <sub>5</sub>	.01	.02	.37	.66	.32	.40
LOI	<u>1.75</u>	<u>.71</u>	<u>1.19</u>	<u>.19</u>	<u>.58</u>	<u>-0-</u>
Total	98.76	99.98	98.85	100.4	99.01	99.97

\*This texture includes subophitic pyroxene and is primary.

To the north, there occurs more disruption of Marcy-type anorthosite by gabbroic material which, locally, is the dominant rock type in the outcrop. Commonly, plagioclase laths in the anorthositic rocks exhibit parallel orientation which appears due to magmatic processes and may vary from block to block. Medium grained, light



feldspars and whatever pyroxenes are present and the margins of plagioclase grains have been widely replaced by myrmekite. This intimate penetration of mangeritic magma into intergranular spaces in the anorthosite is interpreted as indicating that the two rock types are coeval. A brief walk over the top of the outcrop reveals a large number of mangeritic dikes disrupting coarse, Marcy-type anorthosite.

- 40.7 Junction Rts. 3 and 30 in Tupper Lake Village. Turn right (west) on to Rt. 3.  
 46.5 Park alongside long roadcuts of pink granitic gneiss.

#### STOP 12. ALASKITIC ROCKS WEST OF TUPPER LAKE ON RT. 3 (30 MINUTES)

This long series of outcrops consists of pink granitic gneiss (Lyon Mt. Gneiss) which contains irregular interlayers of black, grey, and green hornblende-pyroxene-plagioclase rock. Coarse pegmatites disrupt and crosscut both the dark layers and the granite. Commonly the granites themselves exhibit crosscutting relationships with the dark layers. At both the eastern and western extremities of the outcrop there occur facies typical of the least contaminated granite. These consist of pink, medium grained alaskite whose major mafic mineral is magnetite. Small quantities of diopside and sphene are also present, probably due to contamination. The feldspar in these rocks consists almost wholly of mesoperthite or microcline perthite indicating that these are hypersolvus granites. U-Pb zircon data acquired from this outcrop, as well as three others in the northern and eastern Highlands, yield well constrained ages of ca. 1060-1080 Ma (table 1). The zircons exhibit morphological, zoning, and systematics characteristics that are typical of zircons crystallized from melts (figure 29) rather than those grown during metamorphism. Therefore, the 1050-1080 Ma ages of these zircons date their time of crystallization, and that of the alaskite, from a magma. Clearly, this magma could not have been extruded as a volcanic, since an abundance of evidence dictates that the entire region was undergoing granulite facies (Ottawan) metamorphism at 20-25 km during this time interval. Therefore, the alaskites are best interpreted as synorogenic, hypersolvus granites intruded at considerable depth during the late stages of the Ottawan Orogeny. Placing the intrusive events at the late stages of orogeny helps to explain the highly variable development of strain-related fabrics in these rocks, some of which are devoid of any evidence of strain but do contain good igneous textures. An intrusive origin for the alaskites is also consistent with their hypersolvus feldspars which require temperatures of  $\sim 800^{\circ}\text{C}$  at pressures of 7.5-8 kb and their unusually high Zr-concentrations of up to 2000 ppm, (table 3) which, for rocks of this composition require temperatures of  $\sim 1000^{\circ}\text{C}$  (Watson and Harrison, 1983). Both these constraints require temperatures well above those of regional metamorphism (figure 13) and preclude a metamorphic origin for these rocks. Finally, note that the alaskitic and related granitic gneisses contain no vestige, or even hint, of any primary volcanic characteristics.

Two major reasons (Whitney and Olmsted, 1989) for suggesting a metavolcanic origin for these rocks were 1) the extreme sodic (up to 10-11%  $\text{Na}_2\text{O}$ ) or potassic (up to 10%  $\text{K}_2\text{O}$ ) concentrations of some members which could be accounted for by alteration at the surface, and 2) the presence of semiconformable interlayers, including low-Ti magnetite deposits. The extreme compositions which are not uncommon in acidic volcanics and plutonics of the S. W. United States and may be due to metasomatic processes operating during late magmatism. While extrusive examples are most commonly described, intrusive examples of extreme compositions also exist in intrusive suites, and may be due to metasomatism involving alkali exchange.

- 52.3 Junction of Rts. 3 and 30 in Tupper Lake. Turn right (south) onto Rt. 30  
 55.8 Park across from long roadcut of mangerite.

#### STOP 13. LONG ROADCUT OF MANGERITE (30 MINUTES)

A long roadcut through grey-green mangerite extends for several hundred feet along the south side of the highway. The rock is typical of Adirondack mangerite and consists dominantly of coarse (1-3 cm) grains of mesoperthite (ternary feldspar) together with 5-10% quartz. A chemical analysis is given in table 3 (AC-85-6). Interstitial to the mesoperthite grains are concentrations of iron-rich clinopyroxene and orthopyroxene which exhibits an elongate habit suggestive of acicular texture. The pyroxenes occur in strained, elongate concentrations that are commonly accompanied by fine-grained plagioclase ( $\text{AN}_{25-30}$ ). At contacts between the plagioclase and mesoperthite a great deal of myrmekite is developed at the expense of the mesoperthite. Present in most thin sections are xenocrysts of andesine ( $\text{AN}_{45}$ ) which exhibit characteristic magnetite-ilmenite clouding.

These grains are strongly zoned outward to AN<sub>25-30</sub>. These textures record a complicated history for these rocks. It is not known whether the large grains of mesoperthite represent cumulates in the conventional sense, but they were clearly intruded by other magmas prior to complete solidification. At least one batch of this was ferrogabbroic and the other granitic. In the process, some of the mafic magma may have experienced supercooling thus giving rise to acicular orthopyroxenes. Coeval magmatism is also indicated by the fine grained, leucocratic granitic dike at the southern end of the roadcut. This lithology not only crosscuts the mangerite but immediately to the north (left) occurs as an angular xenolith in the mangerite demonstrating contemporaneity.

These mangerites belong to the Tupper-Saranac sheet of granitoids that border the Marcy massif on the north and west. Eiler and Valley (pers. comm.) have examined the variation of  $\delta^{18}\text{O}$  in this sheet and conclude that it consists of two subdivisions each of which is internally related by fractional crystallization and neither of which can be comagmatic with anorthositic members of the AMCG suite. These variations are different than those in other regional lithologies and the retention of these signatures indicates vapor-absent metamorphism.

Near its southern termination, the mangerite is crosscut by an irregular dike of green pyroxenite (table 13). This dike, which is similar in some respects to the orthopyroxenite-rich dike at Roaring Brook, consists principally of iron-rich clinopyroxene with a Mg-number of 25, corresponding in composition to clinopyroxene in the mangerite. While the origin of the dike remains uncertain, it is suggested that it represents late pyroxene-rich interstitial liquid filter-pressed for the mangerite.

Continue south on Rt. 30.

59.5 Park alongside roadcut of pink granitic gneiss.

#### **STOP 14. PINK HORNBLende GRANITE (1098 $\pm$ 4 MA) (15 MINUTES)**

This streaky, pink hornblende granitic is typical of the younger granitic rocks of the Highlands. A chemical analysis is given in table 3 (AM 86-6) and the mineralogy of the rock is dominated by mesoperthite and quartz with only a few small grains of sodic plagioclase and <5% hornblende. Buddington (1939) mapped this body of granite as younger than, and crosscutting, the mangerites and quartz-mangerites of the AMCG suite. This field-based interpretation is consistent with geochronology, and the intrusive contact passes through the sand and water covered valley between this stop and the next outcrop due north along Rt. 30.

Figure 5b shows the distribution of the younger granitic rocks whose ages fall into the 1090-1100 Ma range. Within the area shown on figure 5b, several occurrences of these rocks have been recognized but they do not represent a significant percentage of the designated area. These rocks are enigmatic, since they do not fit into any major chronological or petrologic group. At present, the sample base for this group is too limited to draw major conclusions on their origin, and additional study is required.

Continue south on Rt. 30.

60.6 Park in parking area at south end of Tupper Lake.

#### **STOP 15. LARGE ROADCUTS OF CALCSILICATE AND GRANITE (45 MINUTES)**

High steep roadcuts across from the parking area expose green, foliated diopside-rich calcsilicates containing layers of white, coarse grained pegmatite. Towards the north end of the cut the volume of granite increases until it becomes the dominate phase. On top of the northern portion of the outcrop a variety of disrupted, disharmonic features may be observed in migmatite and attest to the intrusive nature of the granite.

The highly strained calcsilicates are isoclinally folded together with the pegmatitic quartz-feldspar interlayers. However, near road level, in the middle of the outcrop, a shallow-dipping vein of pegmatite, cuts across foliation and is not folded. This layer is continuous with a steeply dipping, isoclinally folded layer. The only self consistent explanation of this relationship is that the emplacement of granitic material took place syntectonically and outlasted folding.

At the far southern end of the roadcut contaminated, garnetiferous granite has disrupted and incorporated xenoliths of amphibolite. It is common for Adirondack granites to become garnetiferous towards their contacts with biotitic, amphibolitic country rocks, and good examples of this can be found throughout the region. In these instances, the garnets tend to poikilolitically enclose quartz and/or feldspar.

Turn around and head back through Tupper Lake Village on Rt. 30.

74.4 Junction Rts. 3 and 30 east of Tupper Lake village. Proceed east on Rt. 3.

75.1 Park in sandy area on north side of Rt. 3.

#### STOP 16. MARCY FACIES ANORTHOSITE (20 MINUTES)

Park in sandy area on L side of road and walk E to first outcrops. This coarse grained andesine rock is typical, in both composition and relict igneous texture, of the most voluminous member of the anorthosite series. This exposure contrasts with the gabbroic (noritic in part) metanorthosite of the "border facies" in having <10 percent mafic minerals and in being coarser-grained. Buddington (1939) interpreted the finer-grained border facies to be a relatively chilled sample of the parent magma.

The primary minerals are andesine (commonly An<sub>40</sub>-An<sub>50</sub> and locally antiperthitic), augite, hypersthene, ilmenite, magnetite and apatite; amounts and relative proportions of the mafic minerals vary considerably. Quartz is a normative mineral (up to 5%) but is only occasionally seen in thin section. The andesine occurs in two forms, as bluish-gray megacrysts dusted with extremely fine iron-titanium oxides, and as a clear, finer grained, recrystallized groundmass. Metamorphic minerals include garnet (locally present as coronas around hypersthene, oxides, and -rarely- apatite, and as discrete porphyroblasts where the anorthosite has been extensively deformed and recrystallized); secondary clinopyroxene, amphibole, and, less commonly, biotite, clinoziosite, and scapolite.

Zircons extracted from the anorthosite at this location are small and clear, and yield an age of 1040±43 Ma. Both Silver (1969) and McLelland and Chiarenzelli (1990) have interpreted these zircons as metamorphic. The presence of baddelyite (>1083 Ma) and zircon cores (>1113 Ma) are in agreement with this conclusion. Based upon mutually crosscutting relationships, McLelland and Chiarenzelli (1990) have shown the Marcy anorthosite massif to be coeval with the ca. 1130 Ma granitoids that surround it.

The oxygen isotopic composition of the Marcy anorthosite is ~2.5 permil heavier than other "normal" anorthosites; this anomaly was ascribed by Taylor (1969) to exchange with pervasive <sup>18</sup>O-enriched C-O-H fluids during regional metamorphism. However, Morrison and Valley (1988) have shown that the <sup>18</sup>O enrichment is a magmatic feature that was acquired before the anorthosite intruded the crust at shallow levels. Values of δ<sup>18</sup>O in the anorthosite of the NW lobe of the Marcy Massif are extremely homogenous (δ<sup>18</sup>O = 9.3 ± 0.2), which in conjunction with the preservation of magmatic features, indicates that the oxygen isotopic composition reflects magmatic values rather than exchange with metamorphic fluids.

#### END OF DAY 3

#### DAY 4

##### Mileage

- 0.0 Leave Adirondack Loj. Turn southeast (right) at the junction of Heart Lake Rd. and Rt. 73.
- 6.0 Cascade Lake.
- 12.1 Junction of Rts. 73 and 9N in Keene. Continue on Rt. 73 towards Keen Valley (south).
- 13.4 Junction of Rts. 73 and 9N to Elizabethtown; continue south on Rt. 73.
- 25.4 Junction of Rts. 73 and 9N. Continue east to Rt. 87 (Northway).
- 27.7 Junction with Rt. 87. Proceed south on Rt. 87.
- 39.7 Exit Rt. 87 at North Hudson and proceed west to Newcomb on Blue Ridge Highway.
- 56.4 Park along edge of road in large cut in metagabbro.

**STOP 17. OLIVINE METAGABBRO IN ROADCUTS ON BLUE RIDGE HIGHWAY (20 MINUTES)**

Steep roadcuts on either side of the highway expose good examples of Adirondack olivine metagabbro. The rock consists of round, ~.25 cm coronas of red biotite and brown hornblende on oxides set in a garnetiferous matrix of green, spinel-clouded plagioclase and subophitic pyroxenes. Garnet forms well developed coronas between clinopyroxene (originally olivine?) and plagioclase. Olivine is not abundant in this outcrop although it is widespread throughout the rest of this relatively large body. A whole rock analysis given in table 4 (olivine-metagabbro-Tahawus).

There are a large variety of olivine metagabbros in the Adirondacks ranging from Mg-rich to Fe-rich. These are exposed throughout the region but are especially abundant in proximity to bodies of anorthosite. As seen in figure 2, the southern and eastern margins of the Marcy massif are especially rich in olivine metagabbro. Here it is suggested that these bodies are representative of the magmas ponded at the crust-mantle interface that gave rise to the parental magmas of the anorthosite. The bodies now exposed at the surface are interpreted to be late plutons that ascended, without ponding, after the major mass of AMCG had risen and provided crustal pathways. This suggestion is consistent with geochronological data indicating that the gabbros are contemporaneous with the AMCG suite (table 1, samples 21, 22). Given the possibility of this scenario, further detailed petrological studies of the olivine metagabbro should be undertaken. Garnet coronas developed in olivine metagabbros have been of petrologic interest for over 100 years and have been extensively discussed by McLelland and Whitney (1980).

Continue west on Blue Ridge Highway.

57.9 Turn north (right) to Tahawus.

64.3 Turn onto road to Calamity Brook.

65.5 Gated entrance to Cheney Poind. Park alongside large boulders from mine.

**STOP 18. MAGNETITE-ILMENITE ORES AT SANFORD LAKE (30 MINUTES)**

The ore (see Table 4 for chemical analyses of ore and related ore gabbro) in the Sanford Lake district consists of titaniferous magnetite and hemo-ilmenite in subequal amounts. Lamellae of ilmenite in magnetite originated via subsolidus oxidation-exsolution (Kelly, 1979). Green pleonaste spinel commonly forms as an exsolution product in magnetite. Iron sulfides occur as accessory phases. Both titanomagnetite and hemo-ilmenite form abundant small, rod-like inclusions in associated plagioclase sometimes rendering them black and opaque. The average composition of titanomagnetite and hemo-ilmenite in the principal ore deposits is given by Kelly (1979) as  $Mt_{81}Usp_{18}$  and  $Ilm_{94}Hm_6$  respectively. The ore occurs in two major modes: 1) as lean or disseminated ore gabbro, and 2) as massive, rich ore generally in anorthosite but locally within gabbro. The lean ore within gabbro is gradational into the host rock (with which it is commonly conformably layered (Ashwal, 1978, p. 106) but in anorthosite both lean and massive ore exhibit sharp contacts relative to host rock. With the exception of apatite-rich, and possibly nelsonitic, rocks near Cheney Pond, the concentration of  $P_2O_5$  in the ore deposits is strikingly low. Massive ore tends to be concentrated in lenses measuring 600-700 m in length and 150-300 m in width. It is not known whether this conformable configuration is the result of crystal settling, intrusion, or the accumulation of immiscible oxide-rich liquids. This uncertainty extends to the petrologic details of the origin of the deposits, although the evolution of these rocks is understood in the broad perspective. As seen at Stop 10, Day 2, the late differentiates of the anorthosite move toward pronounced enrichment in Fe, Ti-oxides thus yielding liquids of increasingly ferrogabbroic composition together with associated ultramafic cumulates. This suggests that the ores at Sanford Lake are the result of progressive differentiation of magmas residual from gabbroic anorthosite and that, at some point, these magmas became so enriched in iron and titanium that they either precipitate magnetite-ilmenite cumulates (Ashwal, 1978) or immiscibility of Fe-Ti oxide and silicate melts occurs (Kelly, 1979).

The stop at Sanford Lake takes advantage of excellent relationships exhibited in boulders on either side of the Calamity Brook Road at the gated entrance to the Cheney Pond Road. Over three dozen large, fresh boulders from the mines provide outstanding exposure of anorthosite, gabbro, ore-bearing gabbro, and massive ore crosscutting anorthosite. Several boulders containing both anorthosite and ore exhibit what appear to be coeval and pillowing relationships between the two phases. In other instances massive ore and ore-bearing

gabbro crosscut anorthosite. A number of boulders show irregular oxide-rich veins, some of which clearly contain separate fractions of oxide and silicate suggestive of liquid immiscibility. Many of the ore boulders contain xenoliths and enclaves of anorthosite and xenocrysts of andesine some of which are black due to oxide inclusions. Several boulders of good Marcy-type anorthosite are present as are some sheared, hornblende-bearing gabbroic anorthosite. Relationships seen in these boulders demonstrate that the Fe, Ti-oxide ore derives from the gabbros and bears intrusive relationships to the anorthosite. Polished slabs show the oxide phase to intimately penetrate and disrupt the anorthosite on a scale smaller than the grain size of magnetite and ilmenite in adjoining ore. This suggests that the oxide intruded as a liquid and this observation, together with evidence of liquid immiscibility in similar rocks, lead us to suggest that most, if not all, of the Sanford Lake ores were emplaced as immiscible liquids.

Retrace route to Blue Ridge Highway.

- 73.1 Turn west (right) on Blue Ridge Highway.
- 81.0 Intersection with Rt. 28N. Turn southeast (left) towards North Creek.
- 103.7 North Creek. Continue to junction with Rt. 28.
- 103.9 Junction Rts. 28N and 28. Turn north (right) towards North River.
- 109.9 Turn left on to Burton Mines Rd.
- 115.5 Gate to Barton Mines. Park at mineral shop.

#### STOP 19. BARTON GARNET MINE (1 HOUR)

The elongate open pit, oriented rough ENE-SWS, is located in a small olivine metagabbro body, which is in contact with gabbroic anorthosite gneiss on the north and with a fault contact against quartz mangerite on the south. Along the north wall of the pit, typical olivine metagabbro with well-preserved igneous textures is exposed. The igneous mineralogy of this rock was plagioclase-olivine-clinopyroxene-ilmenite. During metamorphism, coronas of two pyroxenes and garnet have formed between olivine and plagioclase, and coronas of biotite, hornblende and garnet have formed between plagioclase and ilmenite (McLelland and Whitney, 1980). Going S across the pit, the gabbro undergoes a nearly isochemical transition into garnet amphibolite, with garnet porphyroblasts commonly up to 0.3 m and rarely up to 1 m in diameter. It is this garnet amphibolite that constitutes the ore; interestingly, the modal garnet in the ore is approximately the same (roughly 15-20%) as in the coronitic gabbros. The composition of the garnet is approximately 43% pyrope, 40% almandite, 14% grossular, 2% andradite and 1% spessartite; zoning, were present, is very weak and variable. Toward the W end of the pit, garnet hornblende with little or no plagioclase is locally present, probably representing ultramafic layers or pods in the original gabbro. In the more mafic zones of the roe body, garnet porphyroblasts are rimmed with up to several cm of hornblende; where the host is less mafic, the garnets have plagioclase ( $\pm$ hypersthene) rims. The details of the ore-forming process are not well understood, however, the ore body probably represents a zone within the gabbro where  $fH_2O$  was locally higher during high grade metamorphism, facilitating diffusion and the growth of the large garnets and favoring the extensive development of hornblende at the expense of pyroxenes and plagioclase. Subsequently, temperatures must have continued to rise since late orthopyroxene and calcic plagioclase form at the expense of garnet-hornblende assemblages.

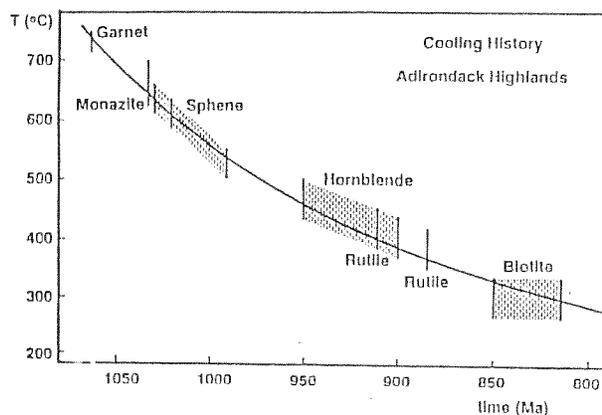


Figure 20. Cooling history of the Adirondacks.  
(After Mezger et al., 1991.)

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Chapter K

# Geology of the Coastal Lithotectonic Belt - SW Maine and SE New Hampshire

By Arthur M. Hussey II and Wallace A. Bothner

Field Trip Guidebook for the Northeastern United States:  
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## GEOLOGY OF THE COASTAL LITHOTECTONIC BELT SW MAINE AND SE NEW HAMPSHIRE

by

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### INTRODUCTION

The Coastal Lithotectonic Belt occupies a zone bounded on the northwest by components of the Norumbega Fault System, most notably the Flying Point Fault in Maine and the Campbell Hill fault in New Hampshire, and on the southeast by the Bloody Bluff-Fundy Fault system (including the Gulf of Maine fault zone of Hutchinson et al., 1988, 1989) (fig. 1, inset). The latter separates classical Avalon Terrane (Williams and Hatcher, 1983) from the coastal composite terrane named Nashoba-Casco Bay-Miramichi Composite and Atlantica Composite by Zen (1989) and Stewart et al. (1991) or Avalon Composite Terrane by Thompson et al. (1993). The area in southwestern Maine and southeastern New Hampshire to be examined on this field trip is underlain by several lithotectonic packages of unfossiliferous metasedimentary and metavolcanic rocks (fig. 1) ranging in age from Late Proterozoic (?) to Middle Paleozoic. These rocks have been intruded by a variety of felsic and mafic plutons of Ordovician to Cretaceous age. On this field trip, we will be examining three of the lithotectonic packages in an oblique cross-section along the coast from Portland, Maine, south to Rye, New Hampshire. These fault-bounded packages include the Casco Bay Group, the Merrimack Group, and the Rye Formation (fig. 1). The geology of individual packages is moderately well-known, but relations between them are as yet poorly understood. We will look at some of the spectacular shoreline exposures along the southern Maine - New Hampshire coast, and talk about what is known of stratigraphy, age, metamorphism, deformation, and plutonism of these belts, and what is still speculative.

### STRATIFIED ROCKS

#### The Casco Bay Group

The Casco Bay Group (perhaps better referred to as the Casco Bay sequence, in our more modern understanding of these rocks as a suspect terrane separated from other packages by faults) is a sequence of metamorphosed volcanic rocks and metamorphosed pelitic, psammitic, impure calcareous metasediments. The first detailed geologic mapping of the area was done by F.J. Katz (1917) under the U.S. Geological Survey. Subsequent detailed mapping has been carried on by Bodine (1965), Hussey (1971a, 1971b, 1985, 1988), Swanson and others (1986), and Swanson (1992).

At the base of the sequence is the Cushing Formation, originally mapped by Katz (1917) as a deformed granodiorite pluton. The Cushing Formation consists of light gray metamorphosed felsic to intermediate tuff, crystal tuff, and volcanic breccia. Interbedded with these, near the top of the formation is a thin, persistent, dark gray garnet-biotite-grunerite granofels. The Cushing Formation also includes rusty pelitic schist. The metavolcanic rocks grade laterally to the east into well-bedded volcanogenic metasedimentary rocks that include calc-silicate granofels, quartzo-feldspathic granofels, amphibolite, marble, and thin, sporadic interbeds of subpelitic quartzo-feldspathic granofels with sillimanite. These are interpreted to be an easterly sedimentary apron proximal to eruptive centers of the volcanic rocks. A U/Pb age or zircons from the Cushing Formation at Danford Cove (Stop 2) gives an age of  $471 \pm 3$  Ma, (Hussey et al., 1993), suggesting eruption of the volcanics during the Middle Ordovician.

The rocks above the Cushing Formation are dominantly metasedimentary. Immediately above are interbedded quartzo-feldspathic metasandstone and metapelite (the Cape Elizabeth Formation), succeeded upward by mafic, intermediate and felsic metavolcanic rocks (Spring Point Formation). Above these volcanic rocks is a thin, but distinctive sulfidic black graphite-quartz-muscovite phyllite (commonly with phyllonitic structure, the Diamond Island Formation). Overlying the black phyllite is gray aluminous phyllite, which is locally very sulfidic (Scarboro and Jewell Formations), and which locally has green chloritic phyllite representing intermediate to mafic volcanic tuffs. In the middle of this pelite is a thin persistent ribbony bedded metalimestone (Spurwink Metalimestone).

The contact between the Cushing metavolcanic rocks and the overlying metapelites and feldspathic metasandstones of the Cape Elizabeth Formation is interpreted to be a slightly tilted angular unconformity on the

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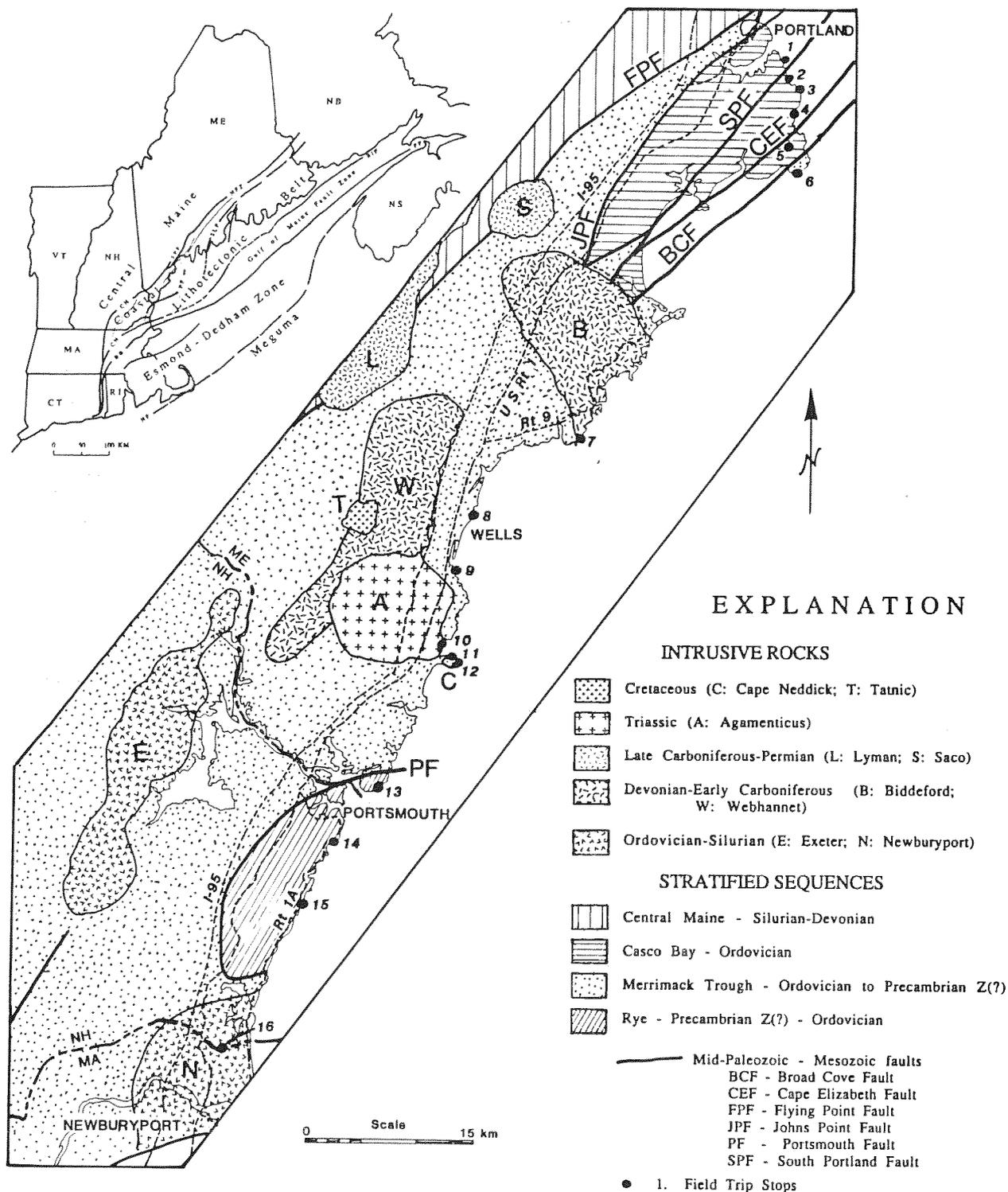


Figure 1. Generalized geologic map of SW Maine and SE New Hampshire. Inset shows boundaries and major internal faults of the Coastal Lithotectonic Belt. Offshore continuation based on aeromagnetic data (Hutchinson et al., 1988; Bothner et al., 1988; Macnab et al., 1990). Abbreviations on inset: NFZ-Norumbega Fault Zone; FPF-Flying Point Fault; SLF-Sennebec Lake Fault; CH-Campbell Hill Fault; CN-Clinton-Newbury Fault; BB-Bloody Bluff Fault; FF-Fundy Fault; BIF-Belle Isle Fault; NF-Nauset Fault.

basis of grits at the base of the Cape Elizabeth, and the fact that the Cape Elizabeth Formation regionally overlies different lithologies of the Cushing Formation (Hussey 1985, 1988, 1993). The age of the stratified rocks above the Cushing Formation is Middle to possibly Late Ordovician.

### The Merrimack Group

The Merrimack Group is a thick sequence of calcareous turbidites, which has been stratigraphically subdivided into three formations, the Kittery, Eliot, and Berwick Formations (Katz, 1917; Billings, 1956; Hussey, 1961; Bothner et al., 1984). The Kittery Formation preserves a variety of primary sedimentary structures, including graded bedding, climbing ripple cross-bedding, flame structures, small scale channel cut-and-fill structure, load casts, flute casts, and parallel lamination. Bedding thickness varies greatly, ranging from 1 cm to 2m. Most of the Kittery Formation has been interpreted to represent turbidites deposited in a variety of deep-sea fan subenvironments. Paleocurrent indicators such as dip of foreset beds of cross-bedding and direction of facing of flute casts suggest sediment transport in the direction S80 W (Rickerich, 1983, 1984). The source would appear to be to the east.

The Eliot Formation consists of light gray and pinkish gray calcareous metasilstone (the pinkish hue due to weathering of fine-grained ankerite and limonite) and dark gray metapelite. Graded bedding is moderately common, but the gradation from metasilstone at the base of a bed to metapelite at the top is quite abrupt, in contrast to grading in the Kittery Formation which is more gradual.

The Berwick Formation, as described by Bothner et al. (1984) is a well-layered, laminated, purplish-gray biotite granofels with thin, often discontinuous calc-silicate layers, stringers, pods, and occasional boudins. Bedding, rarely graded, in the granofels ranges from a few centimeters to 20 centimeters and contains thin, often rusty-weathering pelitic interlayers.

Recent reexamination of low-grade rocks previously correlated with the Cape Elizabeth Formation has resulted in a new correlation with parts of the Merrimack Group. Low-grade calcareous metasilstones and metapelite originally mapped as part of the Cape Elizabeth Formation in the southern part of the outcrop belt of the Casco Bay Group are now considered on strong lithic similarity to be an onstrike extension of the Eliot Formation. These rocks continue in a narrow fault-bounded belt to Freeport, just north of the map area of Figure 1. Similar, but thicker, and more variably bedded low grade metasilstone and metapelite east of the Broad Cove Fault in Cape Elizabeth, typified by the exposures at Two Lights State Park (Stop 6), are reinterpreted as onstrike equivalents of the Kittery Formation of the Merrimack Group. The rocks of both of these belts would appear to be too calcareous to produce the typical Cape Elizabeth lithology of higher metamorphic grade (garnet and higher).

### The Rye Formation

The Rye Formation consists of a series of variably mylonitized, polymetamorphosed calcareous and non-calcareous metasandstone and metasilstone, migmatized and non-migmatized metapelite, rusty-weathering sulfidic graphitic phyllite, marble, and amphibolite. Included within the Rye are abundant, now semiconcordant granitic and tourmaline-bearing pegmatitic layers (porphyroclastic augen gneisses) that share the same mylonitization history as their metasedimentary hosts and apparently had formed the basis for Billings (1956) subdivision of the Rye into metasedimentary and metavolcanic members. See Hussey (1980) for review. The age of the Rye is uncertain. It is assigned an age of Precambrian Z(?) to Ordovician based on cross-cutting metadiorite that yielded an age of 477 Ma (Rb/Sr, Gaudette et al., 1984) and a similar ~460 Ma  $^{40/39}\text{Ar}$  metamorphic age on hornblende from amphibolite not involved in later ductile faulting (West, 1993).

## INTRUSIVE ROCKS

Intrusive rocks in the Coastal Lithotectonic Belt range in age from Ordovician to Cretaceous. Within the field area the oldest rocks are the metadiorites and associated granites (Breakfast Hill) that intrude the Rye Formation. Gaudette et al. (1984) report an Rb/Sr age of 477 Ma for metadiorite on Appledore Island in the Isles of Shoals. The metadiorite also crops out in the Portsmouth Harbor area (md, fig. 5). The sampled Breakfast Hill granite remains to be dated.

Late Ordovician (?) to Silurian intermediate to felsic rocks intrude the Merrimack Group post-orogenically. These include the Newburyport Quartz Diorite (450±15 Ma zircon, optional Stop 15) and the Ayer Granite (435±5 Ma zircon) dated by Zartman and Naylor (1984) and the Exeter Diorite redated by U/Pb zircon and sphene as Silurian

(Bothner et al., unpublished data). All three intrusions cross-cut tight, upright folds common in the Merrimack Group and show variable effects of later metamorphism. The Newburyport and Exeter plutons have contact metamorphic aureoles superposed on regionally metamorphosed Merrimack lithologies.

Devonian to Early Carboniferous plutons assigned to the New Hampshire Plutonic Series, which also intrude the Merrimack Group, include the Biddeford and the Webhannet Plutons. We will briefly visit only the Biddeford Pluton (Stop 7). This is a 15 x 10 km post-tectonic pluton composed of non- to slightly foliated medium-grained light gray biotite granite. Rb/Sr age of this pluton is  $354 \pm 12$  Ma (Gaudette et al., 1982). The Webhannet Pluton is composed of granite and granodiorite from which a Rb/Sr age of  $391 \pm 42$  Ma and a  $403 \pm 2$  Ma U/Pb zircon age was obtained (Gaudette et al., 1982). Both the Biddeford and Webhannet Plutons post-date all phases of deformation and the low grade regional metamorphism of the Kittery Formation, and have imposed medium grade thermal metamorphism up to 0.5 km wide on the surrounding rocks.

Triassic to Cretaceous age minor intrusions and an explosion breccia of the White Mountain Plutonic-Volcanic Suite are common in southwestern Maine and southeastern New Hampshire. In the field trip area we will visit two intrusive complexes, one of Triassic age, the Agamenticus Complex (Stop 10), and the other of Cretaceous age, the Cape Neddick Complex (Stops 11 and 12), and an explosion breccia (Stop 13). At Stop 9, we will examine a suite of basalt, diabase, and lamprophyre dikes that form a dike swarm intermediate in age between that of the Agamenticus and Cape Neddick Complexes. The Agamenticus Complex is an anorogenic complex consisting of alkaline syenite, alkaline quartz syenite, alkaline granite and biotite granite, emplaced during Triassic time (K/Ar age  $228 \pm 5$  Ma; Foland and Faul, 1977), essentially synchronously with development of Triassic rift basins, and opening of the central Atlantic Ocean (Brooks et al., 1989; Brooks, 1990).

The Cape Neddick Complex (fig. 4) consists of two major intrusive phases. The older phase consists of gabbro grading inward to anorthositic gabbro, forming an outer steep-sided funnel intrusion 1 x 1/2 km in dimensions. The younger phase is a smaller funnel intrusion of cortlanditic gabbro. This also forms two cone sheets within the older funnel. Foland and Faul report a K/Ar age of  $116 \pm 2$  Ma for the gabbro of the older, outer, funnel. Prior to emplacement of the gabbro, explosive volcanism formed a breccia that includes blocks of Kittery Formation, basaltic dikes in the Kittery, and felsic volcanic fragments. The Cape Neddick Complex has produced a contact aureole in the Kittery Formation. More feldspathic beds were partly melted and intruded as felsic dikes into the basaltic/diabase dike swarm in the Kittery Formation (Stop 11).

Near the south end of Gerrish Island in Kittery (Stop 13) a small explosion breccia will be examined. This includes fragments of the Rye Formation and earlier mafic dikes, and is cut by later diabase dikes. Brecciation ends very abruptly and is separated from non-brecciated Rye Formation by a thin ring of rhyolite porphyry. One of the interesting aspects of the breccia is the lack of fine matrix between larger blocks.

## STRUCTURE

### Folds

In the Casco Bay Group, early folds (F1) are minor recumbent folds. We will not see these in the stops within the Casco Bay Group, but they can be seen in seacliff exposures on the mainland opposite and just north of Richmond Island in the town of Cape Elizabeth (Cape Elizabeth 7 1/2' quadrangle), among other places. Later F2 folds are upright to very steeply overturned, and variably plunging. These control the major outcrop pattern of the formations of the Casco Bay Group. Minor parasitic F2 folds are seen at stop 4, and are common throughout most of the area where the units of the Casco Bay Group have not been migmatized. At a few areas, they are lacking, and only recumbent parasitic F1 folds are seen. Mineral lineations formed during regional metamorphism are parallel to hinges of the folds.  $^{40}/^{39}$ Ar dating of hornblendes from amphibolites within the Casco Bay sequence by West et al. (1993) suggest metamorphism of that sequence and most likely F2 deformation are Acadian. An alternative view has been advanced by Swanson (1992, 1993) from studies of late Paleozoic ductile faulting in the Cape Elizabeth-Harpswell area. He interprets the folding as the result of flower structures developed at restraining bends of the different segments of the Norumbega Fault System that pass through the area and are Alleghanian.

The Merrimack Group rocks have been subject to similar folding. F1 recumbent folds are particularly well preserved in exposures of the Kittery Formation at Two Lights State Park (Stop 6) and at Marginal Way in Ogunquit (Stop 9). At both localities they show general eastward vergence. Later upright F2 folds are less well developed at either locality, but are seen at the exposures at Moody Point in Wells (Stop 8), and are particularly well developed

and displayed in shoreline exposures just south of the Marginal Way. These folds vary in plunge from gently northeast to gently southwest. The age of folding is constrained by Silurian ages of major post orogenic plutons (e.g. Exeter, and Newburyport). F1 and F2 folds, both, therefore predate the Acadian orogeny. F1 folds of the Merrimack and Casco Bay sequences may be correlative and related to the same pre-Silurian deformational event; F2 folds of the Merrimack sequence must also relate to a pre-Silurian event, and therefore do not correlate with F2 folds of the Casco Bay rocks.

Mesoscopic recumbent folds have not been recognized in the Rye. Small, isoclinal intrafolial folds are common in quartzofelspathic schists and gneisses, amphibolite, and marble and are interpreted to represent the earliest fold structures. Later, tight upright folding deforms foliation, intrafolial folds, and granitic dikes and pegmatite bodies at outcrop scale and control map pattern. Except for some of the granitic dikes whose original attitude controls fold orientation, these folds plunge shallowly to the northeast and southwest and contain weak to very pronounced stretching lineations that parallel fold axes. Stretching lineation is defined by preferred orientation of elongate minerals, porphyroclast tails, quartz rodding (most isolated quartz hinges), and slickenlines, all of which correlate with the varying degrees of mylonitization that so beautifully characterizes the Rye Formation. Shear boudins at scales of a few centimeters (Swanson, 1992) to tens of meters often terminate with strongly folded "tails" that are drawn into the mylonitic foliation (Stops 13 and 14).

### Faults

High angle uniformly north-northeast trending faults in the field trip area are parts of the Norumbega Fault System (fig. 1). Included are the Flying Point Fault, Johns Point Fault, South Portland Fault, Cape Elizabeth Fault, and Broad Cove Fault. Movements along these involve both strike-slip and normal motion. Swanson (1992) postulates right lateral motion for the Flying Point and Broad Cove Faults from detailed studies of kinematic indicators in the rocks and adjacent to the breaks. Studies by West (1988) on the thermal histories of rocks on either side of the Flying Point Fault demonstrate considerable vertical motion along the fault. From  $^{40/39}\text{Ar}$  studies of the age of muscovite "fish" in mylonites along parts of the Norumbega Fault System on strike to the north, West et al. (in press) indicates that strike-slip motion along these faults may be of Permian age. Dip-slip motion may be much later, in the Mesozoic.

The Portsmouth Fault is a regionally significant, compound structure that can be traced from Gerrish Island, Maine, south for some 20 km. It has been interpreted to wrap around the south end of the Rye Anticlinorium (e.g., Lyons et al., 1991), but an alternative hypothesis now being tested from geophysical data is that the fault merges with the Clinton-Newbury fault farther to the south (fig. 1, inset). The fault contains zones of ultramylonite that separate high-grade mylonitized rocks of the Rye Formation from low-grade rocks of the Merrimack Group. A second ultramylonite zone (the Great Common Fault Zone) is mapped within the Rye (Hussey, 1980; Carrigan, 1984a,b,c; Swanson, 1988) that similarly offsets metamorphic isograds (fig. 5). Both ultramylonite zones and the earlier mylonitic foliation within the Rye display strong dextral shear fabrics. Subsequent brittle deformation of the ultramylonite and adjacent rocks produced a variety of brittle structures and pseudotachylite veins (Stop 13) that Swanson (1988) describes and interprets in terms of paleoseismic events of possible Alleghanian age. Age of strike-slip faulting is constrained only by cross-cutting Jurassic(?) and Cretaceous(?) dikes and explosion breccias and a ~330 Ma  $^{40/39}\text{Ar}$  muscovite age (West, 1993) adjacent to the GCFZ on New Castle Island (fig. 5).

### SUMMARY

The accretionary history of the collage of lithic packages that comprise the Coastal Lithotectonic Belt from southern New England to Maritime Canada and its relationship to terranes to the east and west continues to be debated (e.g., Wintsch et al., 1992). Many of the arguments delimiting the character and boundaries of the terranes within the Nashoba-Casco Bay-Miramichi and Atlantica Composite Terranes have been summarized by Stewart et al. (1991 and references therein). Onshore-offshore-onshore connections of lithotectonic packages across the Gulf of Maine have been attempted with some success using available ocean-bottom bedrock samples, seismic reflection, and aeromagnetic mapping by Hutchinson et al. (1988, 1989), Bothner et al., 1988, and Brooks and Bothner (1989).

The three lithic packages in the southern part of the Coastal Lithotectonic Belt reviewed in this field trip are bounded by the Flying Point, Portsmouth, and Clinton-Newbury faults. These major faults are dominated by dextral strike slip motion in at least part of their displacement histories. In both the Portland and Portsmouth areas, higher-grade metamorphic rocks (Casco Bay sequence and Rye Formation, respectively) are interleaved or juxtaposed against lower-grade metamorphic rocks of the Merrimack Group. Thermochronologic data presented by West et al. (in press)

suggests an Acadian metamorphic history for the Casco Bay rocks on the east side and an Alleghanian metamorphic history on the west side of the Flying Point fault. Only pre-Acadian and Alleghanian metamorphism is so far recognized in the Rye Formation (West, 1993). The character of the minor structures studied by Swanson (1988, 1992) and Carrigan (1984) and the pattern of map scale folds and related faults led Swanson (1993) to propose the existence of a positive flower structure formed during dextral transpression at a restraining bend in the Flying Point Fault. Alleghanian dextral transpression was appealed to by Bothner et al. (1988) and by Brooks and Bothner (1989) to account for the pattern of aeromagnetic anomalies interpreted as faults bounding lithotectonic "slivers" and/or axial culminations on and offshore. Although many specific questions remain, particularly those related to timing of early ductile deformation, polymetamorphism, and to lesser extent, magmatism within rather than across individual packages, the later more brittle structural events at and between their bounding surfaces seem to point to strong progressive transpressional to extensional tectonism from late Paleozoic to mid-Mesozoic.

### ROAD LOG, DAY 1

Saturday, board the bus at Ogunquit Motel in Ogunquit at 0800. Take U.S. Route 1, north to junction of 109 in Wells, then 109 north to the entrance to the Maine Turnpike. Take the Maine Turnpike north to exit 7.

#### Mileage

- 0.0 Road log begins at the junction of the exit 7 connector and U.S. Route 1 in South Portland. Turn left onto Rte. 1.
- 1.1 Right turn on Broadway Avenue at stoplight.
- 1.8 At stoplight, bear left on Broadway.
- 4.6 At stopsign, turn right on Pickett Street.
- 4.7 At stopsign, turn left into campus of Southern Maine Technical College.
- 4.8 Turn right and park in parking lot. Walk to the steps down over the seacliff, then walk north to exposures along the shore.

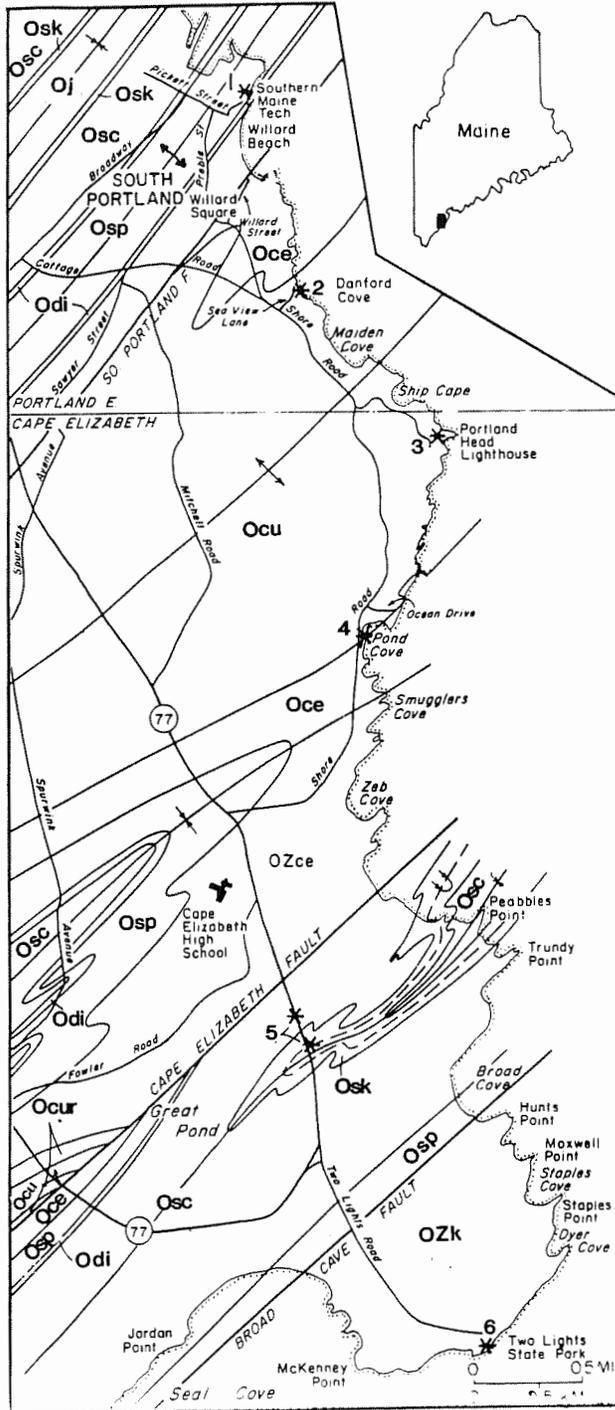
**STOP 1. SPRING POINT, DIAMOND ISLAND, AND SCARBORO FORMATIONS OF THE CASCO BAY GROUP (30 MINUTES).** At the north end of the exposures, near the remains of old Fort Preble, is the type locality for the Spring Point Formation, here consisting of metamorphosed mafic to intermediate volcanic rocks. Distinct to obscure elongate light colored patches suggest the original pyroclastic nature of these rocks. In sharp contact with the Spring Point Formation is dark gray to black graphite-muscovite-quartz phyllite of the Diamond Island Formation, a stratigraphic unit of the Casco Bay Group above the Spring Point Formation. Pollock (1993) interpreted the Diamond Island Formation to represent a shear zone related to right lateral strike slip movement on high-angle faults of the Norumbega Fault Zone. We interpret this rock to be a distinct stratigraphic unit, always appearing between the Spring Point and Scarboro Formations. The phyllonitic nature the Diamond Island fabric we interpret to be a function of the graphitic composition of the rock; shearing is most intense in this rock.

To the east and stratigraphically above the Diamond Island Formation is the Scarboro Formation consisting of non-rusty gray muscovite-quartz-garnet-biotite schist. The ledges visible on the east side of Willard Beach are part of the Cape Elizabeth Formation which lies stratigraphically beneath the Spring Point Formation. The South Portland Fault, a late Paleozoic or Mesozoic high angle fault, not exposed here, separates the exposures here from the Cape Elizabeth Formation to the east.

- 5.1 Return to entrance to Southern Maine Technical College. At stopsign, proceed straight on Fort Road.
- 5.3 Left on Preble Street.
- 5.6 Stopsign at Willard Square. Bear left on Preble Street.
- 6.2 Turn left on Shore Road.
- 6.3 Park on right side of Shore Road, opposite Seaview Avenue. Walk down Seaview Avenue sidewalk to steps that descend to the base of the seacliff.

**STOP 2. CUSHING FORMATION AT DANFORD COVE (20 MINUTES).** To the south of the steps are exposures of massive to weakly layered, strongly lineated fragmental metapyroclastic felsic volcanic rocks of the Cushing Formation. These were first mapped and interpreted by Katz (1917) to be a deformed granodiorite pluton, but exposures at this stop, as well as many other localities, preserve structures indicating fragmental volcanic origin. Volcanic pyroclasts exposed in the seacliff about 50 meters south of the steps are somewhat varied in composition and are strongly stretched in the direction of the major upright fold hinges in the area. From this

HUSSEY AND BOTHNER



EXPLANATION

- Oj: JEWELL FORMATION: Sulfidic and non-sulfidic garnet-mica schist; minor quartzite.
- Osk: SPURWINK METALIMESTONE: Medium gray thinly interbedded fine-grained marble and biotite phyllite. Typically has ribbony appearance on weathered surface.
- Osc: SCARBORO FORMATION: Same lithology as Jewell Formation.
- Odi: DIAMOND ISLAND FORMATION: Sulfidic coal-black quartz-graphite-muscovite phyllite
- Osp: SPRING POINT FORMATION: Medium greenish-gray chlorite-garnet phyllite; chlorite-actinolite-biotite phyllite; dark green blocky amphibolite; minor quartz-plagioclase-biotite-garnet gneiss and granofels.
- Oce: CAPE ELIZABETH FORMATION: Medium gray quartz-plagioclase-biotite-muscovite-garnet schist, and dark gray biotite-chlorite-muscovite-garnet schist.
- Ocu: CUSHING FORMATION: Quartz-plagioclase-biotite gneiss with or without microcline; locally preserves relict fragmental volcanic structures.
- Ocur: Rusty garnetiferous gneiss.
- Ozk: KITTERY FORMATION: Buff-weathering ankeritic and calcareous quartz-plagioclase-chlorite phyllite and dark gray phyllite.

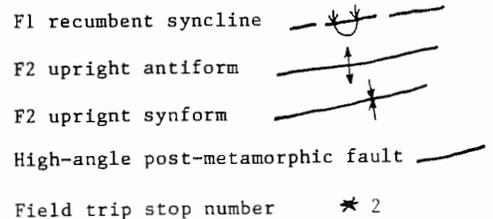


Figure 2. Geology of the South Portland-Cape Elizabeth area (field trip stops 1-6) after Hussey (1987).

locality, a sample taken from the least fragmental appearing of the metavolcanics has yielded a zircon U/Pb age of  $471 \pm 3$  MA (Hussey, et al., 1993). Approximately 200 meters north of the steps, the contact between the Cape Elizabeth Formation (above) and the Cushing Formation is exposed in the low-tide wave-cut bench.

Return to vehicles and continue on Shore Road.

- 7.3 Left turn into Fort Williams Park.
- 7.8 Parking lot for Portland Head Light.

**STOP 3. CUSHING FORMATION AT PORTLAND HEAD LIGHT. (15 MINUTES).** Rocks here are lineated but otherwise rather massive metarhyodacite. These can be viewed at low-tide from the wave-cut bench to which there are a few paths beyond the limits of the chain link fence.

Portland Head Lighthouse is one of the oldest in North America. Work began on the tower in 1787 by the Massachusetts Bay Colony, and was completed by the federal government in 1791. Portland Head Lighthouse sent its rays out over the waters for the first time on January 10th, 1791 (Mathan and Barry, 1991). The first keeper, serving for two years without pay, but with quarters and the right to fish and farm, was Capt. Joseph Greenleaf. At the end of two years, the federal government granted Capt. Greenleaf an annual salary of \$160. The present keepers' house was built in 1891.

Turn around and return to entrance to Fort William Park.

- 8.4 Left turn onto Shore Road.
- 9.4 Park and walk south to shoreline opposite the first house on the right. **All subsequent visitors to this site must ask permission from the owners of the house.**

**STOP 4. CAPE ELIZABETH/CUSHING CONTACT, POND COVE (20 MINUTES).** Exposure of the Cape Elizabeth/Cushing contact. The Cape Elizabeth Formation consists of quartz-plagioclase-muscovite-biotite-garnet schist, slightly micaceous feldspathic quartzite, calc-silicate lenses, and rare amphibolite beds. It is infolded with the Cushing Formation, here consisting of plagioclase-quartz-biotite-muscovite gneiss.

Continue on Shore Road.

- 10.8 Stop sign. Turn left on Rte. 77.
- 11.8 Park on shoulder at beginning of large roadcuts.

**STOP 5. SCARBORO AND SPURWINK FORMATIONS (20 MINUTES).** The northern set of roadcuts on either side of Route 77 here exposes the very sulfidic, rusty weathering phase of the Scarboro Formation. Rare 10 to 20 cm micaceous quartzite beds show that bedding at this locality is steep. A prominent fracture cleavage dipping moderately to the northwest is probably related to a thrust fault, part of the Cape Elizabeth fault zone, that is exposed on strike to the southwest at Ram Island Farm in Cape Elizabeth.

The southern roadcut exposes thinly banded ribbony metalimestone with biotite phyllite interbeds belonging to the Spurwink Metalimestone. Regional mapping shows that this outcrop lies at the core of a refolded recumbent fold. In the Spring of 1988, a block of the metalimestone 3m x 2m x 2m fell from the highest part of the roadcut and half-buried itself in the clay beneath the drainage ditch beside the highway, attesting to the fragility of seemingly solid ledge.

Continue on Rte. 77.

- 12.5 Bear left on Two Lights Road.
- 12.6 Take left fork staying on Two Lights Road.
- 13.6 Straight, into Two Lights State Park.

**STOP 6. KITTEERY FORMATION AT TWO LIGHTS STATE PARK; LUNCH STOP. (90 MINUTES).** The rocks along the shore at Two Lights State Park were long considered to be the type locality of the Cape Elizabeth Formation. Recent revisions to mapping and interpretation (Hussey, et al., 1993) now correlate these rocks with the Kittery Formation of the Merrimack Group. These rocks are distinctively of lower metamorphic grade, consisting of feldspathic and calcareous pinkish-gray weathering metasilstone with interbeds of dark gray chlorite-muscovite phyllite, and dull, brownish-gray weathering lenses of very limy metasilstone. The bedding style, thickness variations of the more resistant metasilstone beds is very similar to low-grade rocks of the Kittery Formation exposed at Moody Point in Wells, Maine. The rocks at Two Lights State Park are separated from

garnet-grade Casco Bay rocks to the northwest by the Broad Cove Fault, a high angle fault of probable Mesozoic age.

- Return to entrance to Two Lights State Park.
- 14.2 Straight on Two Lights Road.
- 15.2 Left on Wheeler Road.
- 15.3 Left on Rte. 77.
- 20.8 Right on Rte. 207.
- 23.9 Left on U.S. Rte. 1 at stoplight.
- 32.8 Rte. 1 bears right. Go straight, proceeding through downtown Saco.
- 33.5 Cross Saco River.
- 33.7 Left on Alfred Street
- 34.0 Left on Rte. 9.
- 34.1 Stoplight. Continue straight on Rte. 9.
- 39.5 Rte. 208 comes in from the left.
- 46.4 Right on Rte. 9 at Cape Porpoise Village.
- 46.7 Left on Wildes District Road.
- 48.1 Left at Wildwood Fire Company building.
- 48.2 Left on Turbats Creek Road. Becomes Shore Road in 0.7 mi.
- 49.7 Park on east side of road, opposite Walker's Point. Walk down short path to ledges at Blowing Cave.  
**CAUTION: Rocks may be extremely slippery if wet, and they slope down to the water.**

#### STOP 7. CONTACT ZONE OF BIDDEFORD GRANITE WITH THE KITTERY

**FORMATION (20 MINUTES).** The Kittery Formation consists of interbedded feldspathic quartzite, greenish calc-silicate granofels, and brownish gray biotite phyllite. Graded bedding is rare. Basaltic dikes up to 3 meters in width are abundant, and cut stringers and irregular pods of granite, here at the edge of the Biddeford Pluton, a late Devonian-Early Carboniferous biotite granite. One of the mafic dikes that cuts the Kittery predates metamorphism (here mostly contact) and intrusion of the Biddeford Pluton (stringers of the granite cut through this dike). Minor folds deform both the stringers of granite and the compositional layering of the Kittery Formation, and are probably related to late stages of intrusion of the granite magma of the Biddeford Pluton. The granite is, in places, weakly foliated parallel to contacts, but is generally massive, and fine to medium grained.

Blowing Cave is a sea cave scoured out of one of the larger diabase dikes at mid-tide level. It is most active when a moderate swell hits the shore at about two hours before and two hours after high tide.

- Continue on Shore Road; becomes Ocean Avenue.
- 51.7 Left on Rte. 9 at Kennebunkport Village.
- 52.0 Straight on Rte. 9 at stoplight.
- 56.3 Left on U.S. Rte. 1.
- 58.9 Straight at stoplight on Rte. 1 at Wells Corner (junction with Route 109).
- 60.2 Straight at stoplight (junction with Route 9B).
- 60.4 Left on Eldridge Road.
- 60.9 Right on Ocean Avenue.
- 61.2 Left on Webhannet Drive
- 61.3 Park on right, opposite Gray Gull Inn.

**STOP 8. KITTERY FORMATION AT MOODY POINT, WELLS. (20 MINUTES).** Walk to shoreline exposures of the Kittery Formation beginning in front of the first waterfront cottage, and from there south around Moody Point. Metamorphic grade here is chlorite zone, and the rocks preserve abundant ankerite and calcite. The rocks are thin to thick (2+ m) bedded metamorphosed feldspathic wacke with graded bedding and rare small-scale cross bedding, a typical flysch sequence. Flute casts can be seen at one locality, and suggest, after unfolding, current flow roughly to the southeast. Four anticlinal F2 hinges are exposed. These folds plunge gently southwest, and are slightly overturned with a northwest vergence.

- Continue along Webhannet Drive.
- 61.5 Left on Eldridge Road.
- 62.5 Right on U.S. Rte. 1.

- 65.8 Diagonal left on Shore Road in downtown Ogunquit.
- 66.3 Left on Israels Head Road.
- 66.7 Park in marked parallel parking slots along road near the miniature "lighthouse", at the entrance to the Marginal Way footpath. We will walk to the south nearly to Perkins Cove and then retrace our steps.

**STOP 9. KITTERY FORMATION AND MESOZOIC DIKE SWARM. (2 HOURS).** One of the finest public walks, the Marginal Way in Ogunquit, between Israels Head and Perkins Cove preserves excellent exposures of the Kittery Formation with its attendant basalt, diabase, alkaline felsite, trachyte, and lamprophyre dikes. Several generations of dikes are seen, displaying excellent examples of dilation, left en echelon terminations, chill margins, and xenoliths. The Kittery Formation here shows several facies of the general environment of deep submarine fan deposition. Where bedding is thick (up to 2 m) the lowest part of an individual bed commonly preserves granule-sized grains of quartz, plagioclase, and dark rock fragments. Upright to slightly overturned folds like the last stop are rare here, but recumbent F1 folds are well displayed, as are asymmetric, moderately overturned F3 folds. Several minor faults and slightly mineralized fracture zones are common, cutting both the Kittery Formation and older dike sets.

### ROAD LOG, DAY 2

We leave promptly at 0800 from the Ogunquit Motel

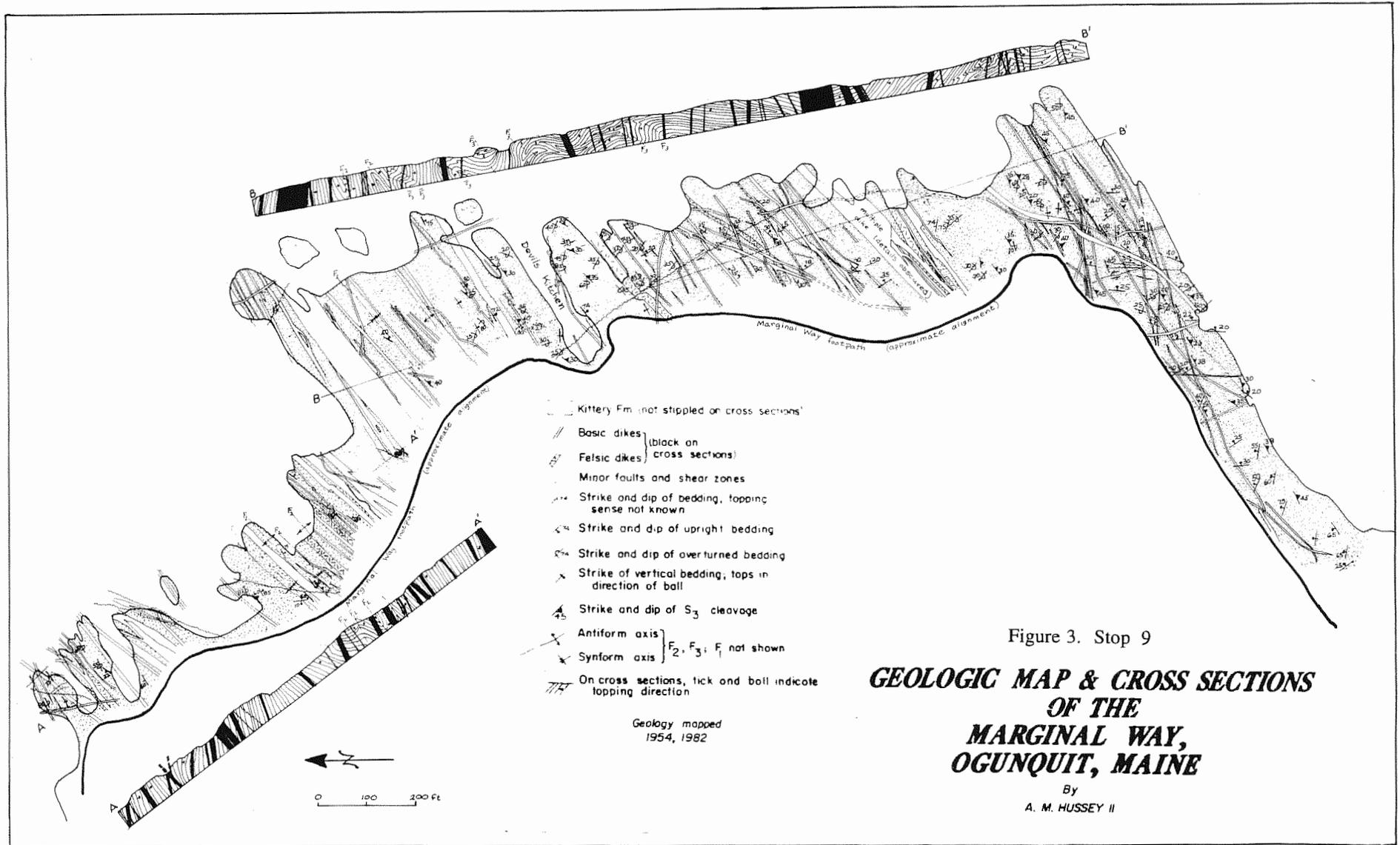
#### Mileage

- 0.0 Road log begins at the Ogunquit Motel, on U.S. Route 1, 2.3 miles north of the intersection of Rte. 1 and Shore Road in downtown Ogunquit.
- 2.3 Bear right, staying on Rte. 1.
- 5.5 Left on River Road.
- 6.5 Right on Shore Road.
- 6.8 Park at Cape Neddick Campground (ask permission to park, and to walk to exposures along the south bank of the Cape Neddick River).

**STOP 10. CONTACT OF AGAMENTICUS COMPLEX ALKALINE GRANITE AND THE KITTERY FORMATION. (30 MINUTES).** The Kittery Formation here consists of thin interbedded feldspathic and biotitic quartzite, purplish biotite pelite, and calc-silicate granofels, in the contact aureole of the alkaline granite phase of the Agamenticus Complex (Triassic age). Dikes of alkaline granite cut the Kittery Formation and are in turn cut by later basalt and diabase dikes of Triassic to Cretaceous age. The contact is exposed at the east end of the Cape Neddick Campground.

- Continue on Shore Road.
- 7.1 Stop sign. Straight on U.S. Route 1A.
- 7.7 Right at York Beach Village.
- 7.7 Left at firehouse.
- 8.1 Left on 1A.
- 8.2 Right on Nubble Road.
- 8.6 Park along shoulder of Nubble Road opposite Fort Hill Avenue.  
Walk to end of Fort Hill Avenue (about 200 meters), descend to shoreline exposures by a trail that starts at the side of the garage at the end of the avenue.

**STOP 11. CONTACT OF THE KITTERY FORMATION AND CAPE NEDDICK GABBRO. (30 MINUTES).** Here you will be able to observe the structures and mineralogy of the contact aureole in the Kittery Formation around the Cretaceous age Cape Neddick Gabbro. Numerous older Mesozoic dikes that cut the Kittery Formation are recrystallized by the Cape Neddick Gabbro. Heat from the gabbro has partially melted the quartzo-feldspathic beds of the Kittery, in places providing melt that has invaded dilated joints of the basic dikes. The more pelitic beds resisted melting and acted as semi-rigid struts that telescoped past each other into the quartzo-feldspathic material. During the cooling of the gabbro, a steep funnel intrusion, the crystallized outer part of the pluton, like a large crucible holding the still molten magma in the center, sat in a plastic, semi-molten mush of country rock. The fact that joints of many of the mafic dikes are dilated may be due to minor explosive events during emplacement of the gabbro magma. Joint blocks may have been held apart by small pieces of either the Kittery or the dikes wedged between the blocks.



9.2 Continue along Nubble Road.  
Left into parking area of Sohler Park.

**STOP 12. CAPE NEDDICK GABBRO AND ANORTHOSITIC GABBRO. (30 MINUTES).**  
We start out walking across weakly and steeply layered outer phase of the older intrusion where layering parallels outer wall of the complex. As we walk, generally westerly, notice that the percentage of dark minerals gradually decreases to nearly that of an anorthosite, and the weathered color is very light gray. Layering in the anorthositic gabbro in places is very rhythmic, marked mostly by zones 1 to 2 cm wide that are depleted in dark minerals. Continuing across the outlet of the small pond toward the set of cabanas, we approach the contact with a 20 meter wide zone of dark gabbro with abundant olivine. This is the outer cone sheet of cortlandtitic gabbro of the inner, younger, intrusive phase of the complex. Here, inside the cone sheet, the layering in the anorthositic gabbro is much gentler. Farther north along the shore we come to a zone where much of the gabbro transitional from the outer normal gabbro to the anorthositic gabbro is missing. Layering is very distinct and in places resembles channel cut and fill structure, but is anomalous in that "topping" would be interpreted as facing the outer edge of the pluton, rather than toward the center, and the layering is universally steep. Long, thin slabs of gabbro internally layered as above, occupy the transition zone, and are marked by harrisitic (crowfoot) augite on the sides of the slabs facing the center of the pluton. The section of the gabbro transitional between normal and anorthositic gabbro is probably missing due to slumping into the magma chamber after its precipitation on the near vertical wall. This may account for the fact that the layering becomes significantly gentler a few meters toward the center of the complex. The origin of the layering at this point is unclear. A crystal settling mechanism is clearly ruled out on the basis of steepness of the layering and the lack of igneous lamination. The origin of the truncation of layers is more enigmatic and perhaps will serve as the subject of lively discussion on the outcrop.

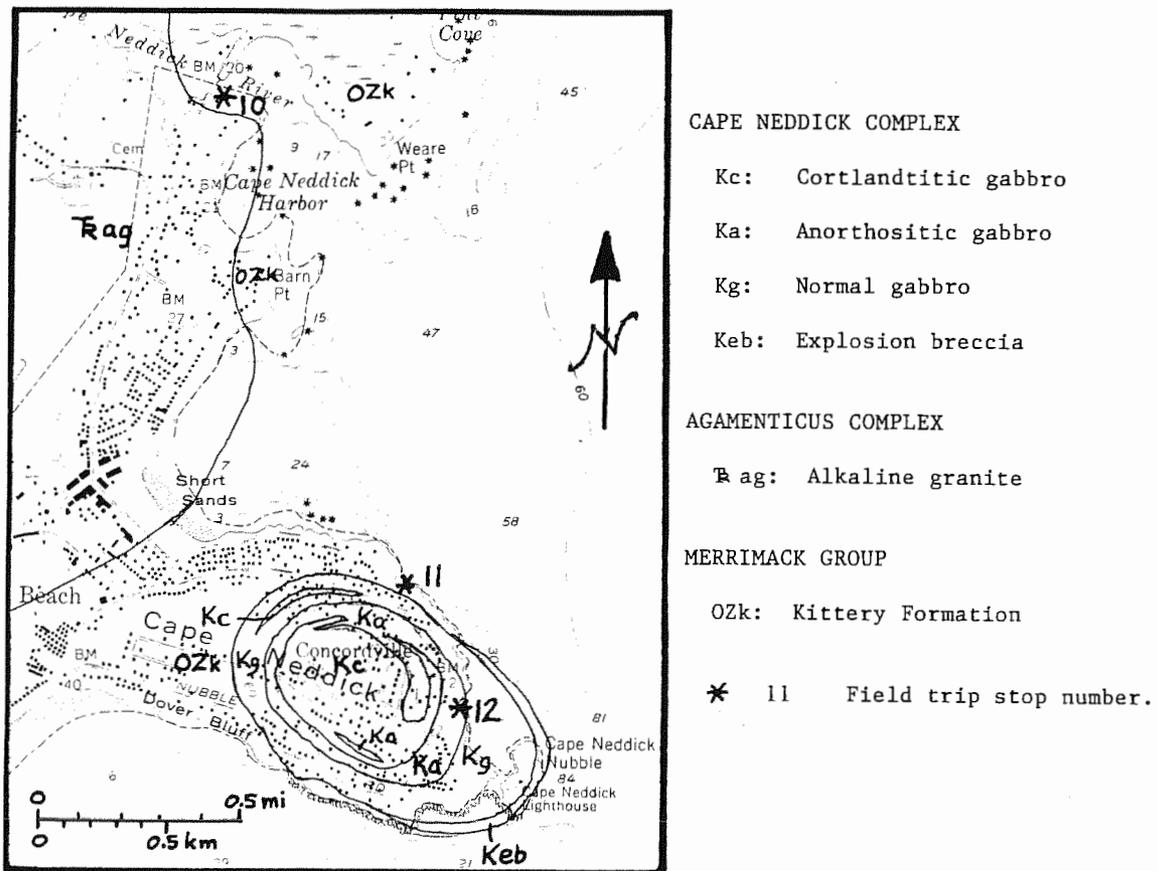


Figure 4. Geology of the York Beach area (field trip stops 10-12).

- Turn left at the exit of Sohier Park parking area onto Nubble Road.
- 10.2 Left on Rte. 1A.
- 13.2 Left on Rte. 103 South.
- 15.1 Braveboat Harbor Road on left. Stay on 103.
- 15.5 Payne Road on right. Stay on 103.
- 17.5 Turn left on Gerrish Island Lane.
- 17.8 Stop sign. Proceed straight and cross bridge to Gerrish Island.
- 17.9 Turn right at the end of the bridge. For the next 0.4 miles we obliquely cross the Portsmouth Fault Zone. The fault juxtaposes sillimanite-grade mylonitic rocks of the Rye Formation against biotite-grade Kittery Formation over a transitional zone up to 250-m thick. Ultramylonite and brecciated ultramylonite mark the steep northwest-dipping outside boundaries of the fault zone in the Portsmouth Harbor area. Both typical Kittery and mylonitic Rye lithologies are found within the fault zone (Carrigan, 1984).
- 19.0 Stop at the fork near the entrance to Fort Foster Park, disembark and walk in along dead end road, through the Delano Property (**ask permission**) to shore exposures (~15 minute walk). The bus will continue through the gate, take the first left and wait for us at "The Pavilion" where we will eat lunch before continuing our trek along the shore to the Fort where the bus will meet us to depart south.

**STOP 13. THE GEOLOGY OF GERRISH ISLAND (~3 HOURS, INCLUDING LUNCH).** Gerrish Island is underlain by the northernmost exposure of the enigmatic Late Precambrian(?) or Early Paleozoic(?) Rye Formation in fault contact with the Kittery Formation. Approximately 3 miles (5 km) of shore line outcrop exposes nearly the entire range of mylonitized metasedimentary lithologies of the Rye and the similarly deformed granitic and pegmatite injections, two major fault zones that record both ductile and brittle deformational history, abundant Mesozoic diabase and some rhyolite dikes, and matrix-free intrusive breccias related to White Mountain magmatic activity seen earlier in the day. A history of the stratigraphic nomenclature applicable to this area may be found in Hussey (1968, 1980) and Billings (1956) and will not be reviewed here, but evidence supporting the lack of a volcanic component in the Rye will be discussed on the outcrops.

We start our trek at the very "photoscenic" (A) Yellow Rocks (Swanson and Carrigan, 1984, stop 7D), the larger of two explosion breccia bodies exposed on the east side of Gerrish Island. This body consists of matrix-free randomly oriented blocks up to about one-meter in size of strongly deformed and heavily migmatized metapelite. The second vent breccia crops out ~1.5 km to the northeast and contains a 3-meter block of the Kittery Formation (brought up? or dropped down?). Unbrecciated, but strongly mylonitized, metapelite crops out just northeast of this vent structure and contains the assemblage sillimanite-garnet-mica-quartz-feldspar. Quartzo-feldspathic stringers and pegmatite lenses cross-cut foliation at very shallow angles and in turn are cut by ductile shears. The vent breccias also contain diabase dike fragments of probable Triassic age [the only dikes dated are those summarized by McHone (1984) from whole rock K/Ar age determinations acquired during the construction of the Seabrook Nuclear Power Plant which range from latest Permian to early Jurassic], are mantled in part by a "rind" of rhyolite, show variable sulfide mineralization at/near the margin, and are cut by still later diabases, one nearly 20-meters thick. We assume a Cretaceous age here based on the age of the nearby Cape Neddick Complex and cross-cutting relations.

The point at the east side of Swards Cove (B, Cedar Point of Swanson and Carrigan, 1984, stop 7C) consists of weakly unmylonitized ~25-meter thick amphibolite with calc-silicate lamellae. The layer is mappable across the southern end of Gerrish Island north of the second ultramylonite zone (Fort Foster Brittle Zone of Swanson, 1988). The amphibolite is recognized in a similar setting across Portsmouth Harbor at New Castle Commons, New Castle, NH. In both localities the amphibolite body has a map pattern of "pinch and swell", "mega-boudin", or, as Swanson and Carrigan (1984) noted, a "megaporphyroclast". Any of these terms apply as the interior part of the layer is less deformed than the margin and the overall shape is consistent with a pull-apart and rotated mass that moved dextrally. The same sense of shear is recognized in the surrounding quartzo-feldspathic augen gneisses. The protolith for the amphibolite here and at New Castle, where a young  $^{40/39}\text{Ar}$  age of 330 Ma has been obtained by West (1993), is interpreted to be a "dirty" calcareous sandstone. The "young" age is thought to reflect its proximity to the ultramylonite in Fort Foster Brittle Zone. An additional  $^{40/39}\text{Ar}$  age from amphibolite exposed in the central region of the Rye in New Hampshire has yielded an age of ~460 Ma for major metamorphism in the Rye.

From "Cedar Point" we walk along the beach of Swards Cove to "The Pavilion" (C) at Fort Foster Park to enjoy lunch and to examine the southern ductile fault zone first recognized by Hussey (1980) as the Southern Mylonite Zone (SMF, fig. 5). The fault zone has been studied in great detail and renamed the Fort Foster Brittle Zone (Swanson and Carrigan, 1984, stop 7B; Swanson, 1988). This 25- to 35-meter thick zone of very fine and

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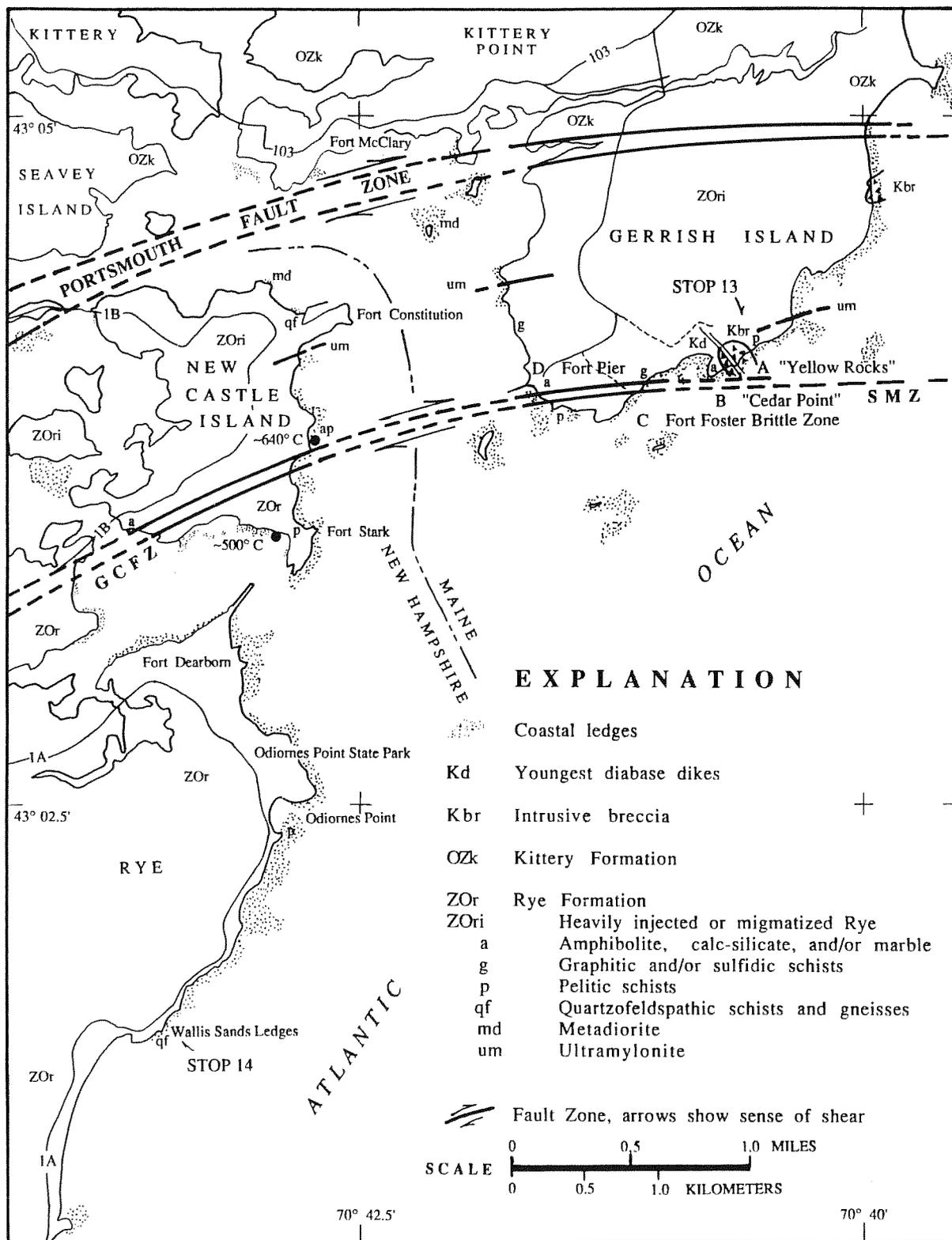


Figure 5. Geologic map of the Portsmouth Harbor area, southern coastal Maine and New Hampshire (base: USGS Kittery 7.5-minute quadrangle). Modified after Hussey (1980), Carrigan (1984c), and Swanson (1988). Filled circles north and south of GCFZ (Great Common Fault Zone) and SMZ (Southern Mylonite Zone) are sample sites for geothermobarometry.

evenly pin-striped chalky brown weathering chocolate-brown ultramylonite contains rootless pseudotachylyte veins emanating from slip surfaces that parallel and bound the shear zone. The veins typically occupy tensional fractures that Swanson (1988) argues formed during rapid layer-parallel shear when ultramylonitic rocks were at the transition from ductile to brittle conditions. The Fort Foster Brittle Zone contains a myriad of important brittle structures superposed on earlier (not much earlier) ductile structures that are elegantly described by Swanson (1988), and are interpreted in terms of paleoseismicity and regional strike slip kinematics. We have attempted to date the pseudotachylyte here by Rb/Sr methods (Boeckeler and others, unpublished data) but have so far been unsuccessful in obtaining a publishable age. Those results will be subject to discussion on the outcrops. Metasedimentary rocks caught along the northern border of the fault zone at the Pavilion include rusty weathering graphitic sulfidic schist and impure marble.

Continuing south of the Fort Foster Brittle Zone along the shore and/or the path we follow the ultramylonite for about 1 kilometer. Outboard (seaward) of the shear zone here, and on strike in New Hampshire, are lower grade mylonitized andalusite-garnet-biotite schist, quartzite, and lesser calc-silicate. As noted by Hussey (1980), Carrigan (1984b), and Brooks (1986) and now confirmed by garnet-biotite thermobarometry by Peter Welch (below) there is a significant difference in metamorphic temperatures on opposite sides of the Southern Mylonite Zone (Great Common Fault Zone (GCFZ, fig. 5) of Carrigan, 1984a). Fewer granite and pegmatite injections are noted, though when present are at least equally mylonitized. At the main pier (D), shear boudins of mylonitized quartzo-feldspathic gneiss is common. Much of this material can be demonstrated to be cross-cutting the metasedimentary assemblage.

- Leave main gate and return to Chauncey Creek Bridge.
- 21.2 Turn left onto Chauncey Creek Road.
- 22.8 Intersection with Route 103, continue left towards Kittery, ME.
- 23.9 Fort McClary State Park, dating back to 1715 as Fort William Pepperill and one of the opposing forts that protected Portsmouth Harbor through four later wars. Primary sedimentary structures (Bouma sequence) in metaturbidites of the Kittery Formation are beautifully preserved in ledges on the south side of the Fort (Rickerich, 1983, 1984; Hussey and others, 1984).
- 25.5 Junction of Rtes 103 and 236 just past entrance to Portsmouth Naval Ship Yard, follow 236 to avoid narrow streets of Kittery Center to Route 1.
- 26.5 Traffic circle. Drive nearly all the way 'round to follow Route 1S, pass through two lights.
- 28.2 Cross Memorial Bridge (Route 1A) to Portsmouth, NH - reverse direction, follow signs to and pass Strawberry Banke (reconstructed historical center dating to ~1630), continue to Route 1B.
- 28.5 Route 1B (New Castle Avenue) intersection, turn right towards New Castle.
- 29.3 Portsmouth Fault Zone (Kittery/Rye transition zone of Carrigan, 1984c) is exposed between the two bridges connecting Goat Island to Portsmouth and New Castle.
- 29.7 Enter New Castle, NH. Outcrops south of Riverside Cemetery are mylonitic granitic gneisses correlated lithically with the Breakfast Hill Granite.
- 30.6 Entrance to U.S. Coast Guard Station. Fort Constitution also dates from the Revolutionary War (then called Fort William and Mary built in the 17th century to protect Portsmouth from pirates); rebuilt in 1808, it served through WWII. Excellent mylonites, metadiorites, xenolith-rich diabase dikes of Triassic age crop out at and around the old fort.
- 31.0 Great Island Common (New Castle town park) where the continuation of the Fort Foster Brittle Zone, here named the Great Common Fault Zone (GCFZ, fig. 5) by Carrigan (1984a,b,c) and Swanson and Carrigan (1984, Stop 3), is well exposed along the eastern shore. North and west of the fault zone, pelitic schist and amphibolite heavily injected by granite and pegmatite are all strongly mylonitized. To the south and east, across the ultramylonites of the fault zone are lower grade pelitic schist, quartzite, and sheared pegmatite.

### Pressure-Temperature estimates across the GCFZ

by Peter W. Welch (UNH'93)

Electron microprobe analyses of amphibolite and pelitic schist sampled north and south of the GCFZ on New Castle Island (fig. 5) has confirmed that the petrographic evidence noted for Stop 13 that a significant difference in metamorphic grade occurs in the Rye across the fault. Relatively undeformed amphibolite exposed as a boudin mantled by highly sheared sillimanite-garnet-biotite schist sampled north of the fault yielded temperatures of 700 and 725°C at 5 Kbar using the plagioclase-amphibole thermometer of Blundy and Holland (1990). The garnet-biotite thermometer of Goldman and Albee (1977) yielded temperatures from the pelitic schist envelope of 625-675°C with some internal points reaching 725°C. Samples from a pelitic layer containing garnet, sillimanite, andalusite, and biotite at Fort Stark, south of the GCFZ, yielded garnet-biotite estimates between 480° and 540°C at 4 Kbar using

Ferry and Spear's (1978) method. Regardless of method, differences of about 150°C and about 1 Kbar occur across the GCFZ.

- Continue on Rte. 1B.
- 31.6 Wentworth-by-the-Sea resort hotel...hopefully to be refurbished. Excellent exposures of mylonitic Rye Formation crops out northwest of the hotel and to the southeast surrounding the saltwater swimming pool (Swanson and Carrigan, 1984, stop 1).
- 32.8 Intersection of Routes 1A and 1B, stay left on 1A towards Rye and Hampton Beaches.
- 33.3 Stay on 1A..
- 35.0 Entrance to Odiornes State Park.
- 35.3 Parking slots at south end of Odiornes Cove where at low tide one can observe part of a drowned stand of white pine (*Pinus strobus*). The 3250±200 year old (C-14, Goldthwait and Lyon, 1934) stumps record some of the transgressive history of the post-glacial Atlantic Ocean. In addition, there are abundant outcrops of mylonitic Rye, cross-cutting deformed pegmatites, and Mesozoic diabase dikes - a favorite locality for UNH student exercises.
- 36.4 Park in "slots" east side of 1A, north of beach.

**STOP 14. LEDGES NORTH OF WALLIS SANDS STATE BEACH.** Anastomosing mylonitic foliation is well displayed in both the metasedimentary Rye and in cross-cutting granitic "sills" where very strong C/S fabric is very well developed. Early tourmaline-bearing pegmatites are strongly deformed, large K-feldspar grains are megascopic sigma-porphyroclasts demonstrating the common dextral shear widespread in the Rye. Granitic layers, metasedimentary layers, and a sulfide-rich metadiabase(?) are pod-like in map view (shear boudins?). The Rye here is characterized by well foliated garnet-bearing quartz-feldspar-biotite granofels intercalated with thin (1-4 cm) pinkish calc-silicate layers. Both rock types contain isoclinal intrafolial folds and strong subhorizontal stretching lineation. However, feldspars are not severely deformed and are slightly coarser than quartz, which is flattened in the plane of foliation. The calc-silicate layers have a mineral assemblage of garnet (Ca-rich)-epidote-hornblende-biotite-feldspar-quartz-sphene that suggest temperatures of ~600°C (Rice, 1983) which indicate increasing metamorphic grade in the Rye southeast of the Great Common Fault Zone (Carrigan, 1984a). The granite has a blastomylonitic texture. Both plagioclase and microcline are fractured and rotated into C-planes enhanced by preferred orientation of fine-grained white mica. Quartz has undergone significant grain size reduction and has not totally recovered, a feature common to much of the Rye.

- 39.0 Cross Rye Harbor bridge.
- 40.4 Entrance to Jenness State Beach.
- 41.4 Turn right on Church Road, park, and cross Rt. 1A carefully.

**STOP 15. PELITIC SCHIST SCREENS AND DIKE SWARM.** Crops below rip-rap are dark gray garnet(coticule)-tourmaline-biotite-(fibrolite) and lighter gray quartz-feldspar-mica schists of the Rye Formation. The schist is preserved as "screens" several meters thick between half a dozen 0.5 to 4-meter thick multiple diabase dikes that occupy >50% of the outcrop area (nice dike swarm). The screens show little horizontal and vertical displacement. The pelitic schist is folded about shallow south plunging axes that approximately parallel the stretching lineations and are further cut by numerous ductile mostly dextral shears similar to those noted at other outcrops of the Rye to the north. In thin section, the pelitic schist contains abundant fine grains of idioblastic tourmaline, garnets (as massive coticles and as single idioblastic grains with inclusion rich cores and clear rims), kinked biotite, and highly strained quartz (good deformation lamellae, subgrain development, mortar texture). Fibrolite(?) is now massive fine grained white mica. Some chlorite formed at the expense of both biotite and garnet (possibly from dike emplacement).

- 42.5 Route 1A/Route 111(old 101D) intersection. Little Boars' Head.
- 43.0 North Hampton State Beach.
- 43.3 Route 1A/Route 27 (old 101C) intersection. Ledges at Plaice Cove expose mylonitic felsic gneiss, ultramylonite (only at low tide), and several cross-cutting premetamorphic mafic dikes with well-developed foliation. Drive "behind" Great Boars Head, Rye not exposed beneath drumlin hill except at low tide.
- 45.4 Route 1A/Route 51 (101) Low grade (greenschist facies) Kittery Formation is exposed on Hampton Beach and has been used to illustrate the "symmetry" of the Rye "anticlinorium".
- 46.4 Bridge over Hampton Harbor Inlet, Seabrook Nuclear Power Plant to west built on Ordovician Newburyport quartz diorite. We pass over the cooling tunnels about under the bridge. The tunnels started in quartz diorite, passed through the contact with Kittery Formation lithologies, here contact

- metamorphosed to pyroxene hornfels (sillimanite is present in thin section), and extend several kilometer out to sea.
- 48.4 Turn right (west) on Route 286.
- 48.9 Low outcrops of coarse grained, porphyritic, and foliated Newburyport Quartz Diorite on south side of road (Shride, 1976, Stop 3, 1976).
- 49.2 Turn left at POINT SEAFOOD diner.
- 49.9 Turn left on Seabrook Road (no sign.)
- 50.8 **STOP 16 (OPTIONAL, TIME DEPENDENT).** Quarry of Newburyport quartz diorite is a coarse-grained nonporphyritic biotite quartz diorite with abundant clots of more mafic diorite and cross cut by several diabase dikes (Hon and others, 1986; Shride, 1971). This quarry provided samples that were dated by U/Pb zircon methods by Zartman and Naylor (1984) at 466 Ma. A younger 437 Ma age is reported from the porphyritic inner zone of the pluton. The age of the Newburyport quartz diorite remains one of the important constraints on the age of the Merrimack Group.
- 51.2 Turn left, then turn right, continue on Seabrook Road.
- 51.6 Intersection Route 1/Seabrook Road, turn left (south) on Rte. 1.
- 51.9 Lights, dangerous intersection. Cross carefully remaining on Rte. 1.
- 53.7 Bridge over Merrimack River, entering Newburyport, MA, remain on Rte. 1 to Boston or follow signs to I-95.

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Chapter L

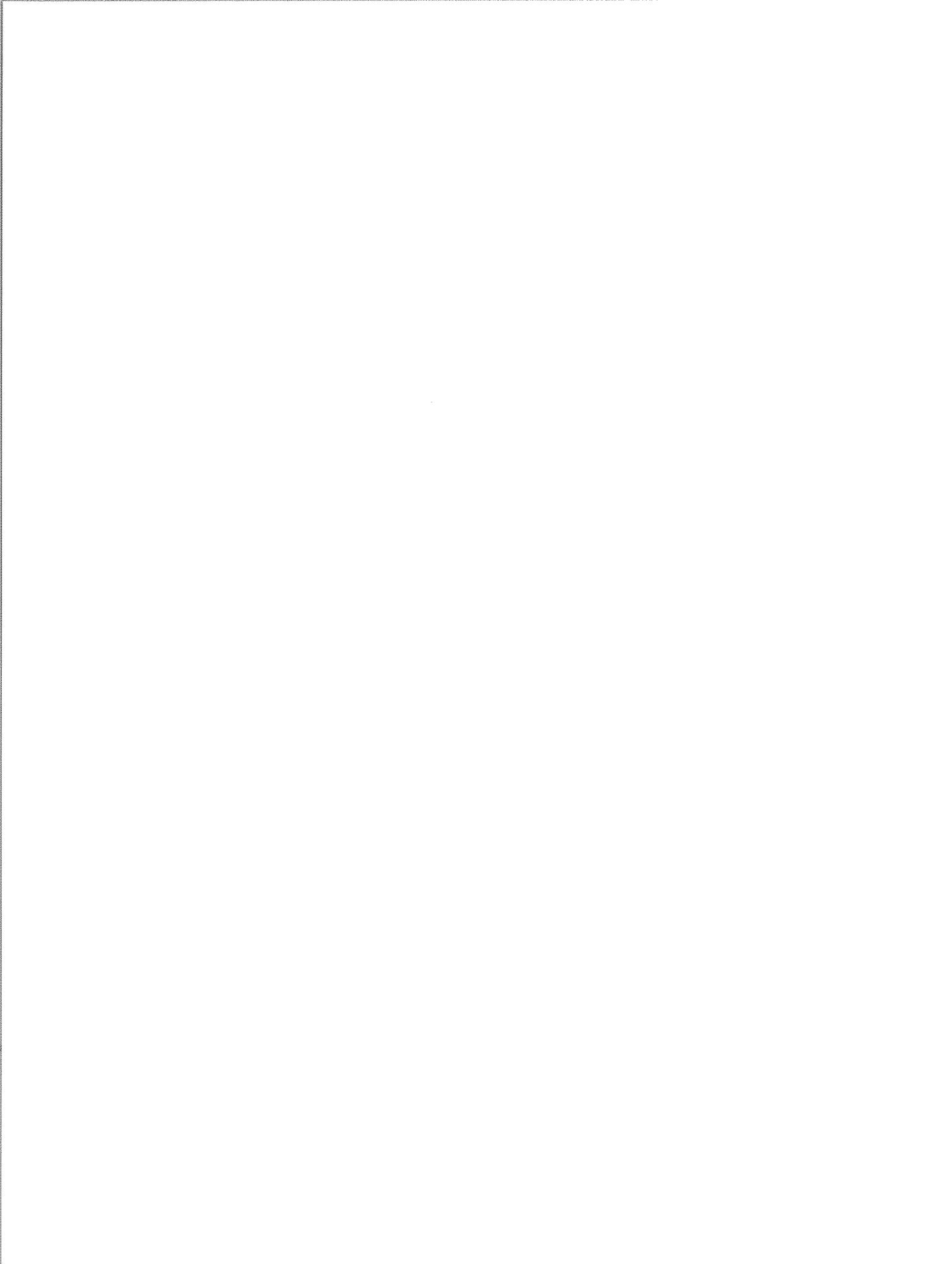
# Petrology and Field Relations of Successive Metamorphic Events in Pelites of West-Central Maine

By Charles V. Guidotti and Michael J. Holdaway

Field Trip Guidebook for the Northeastern United States:  
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## GUIDOTTI AND HOLDAWAY

PETROLOGY AND FIELD RELATIONS OF SUCCESSIVE METAMORPHIC  
EVENTS IN PELITES OF WEST-CENTRAL MAINE

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## INTRODUCTION

The primary purpose of this field trip is to examine metamorphism in west-central Maine of three distinct events, M2, M3, and Carboniferous, and to note similarities and differences in their field relations, mineral assemblages, and textures. We will see that (1) metamorphic pressure ( $P$ ) increased in general from north-northeast to south-southwest; (2) successively younger events occurred progressively farther to the south; and (3) at any given place, where a younger event truncated an older event, the younger event occurred at higher  $P$ . A close spatial relation between plutonic intrusion and metamorphism will be demonstrated.

Stratigraphic units in the region are primarily Siluro-Devonian with local areas of Ordovician. Virtually all the rocks are metasedimentary, primarily pelitic and psammitic with lesser amounts of impure carbonate.

## METAMORPHIC EVENTS AND MINERAL ASSEMBLAGES

Metamorphic events in west-central Maine are recognized as distinct episodes when (1) they are clearly distinct in time of origin as recognized by geochronology or relative age as seen from field relations, and (2) they are clearly different in  $P$  of origin as recognized by differences in pelitic mineral assemblages and geobarometric estimates. Because of the close association between plutonism and metamorphism, these events appear to be related mainly to when, where, and at what depth plutons and sill-like bodies of S-type granitic magmas were emplaced throughout the region. The work summarized below comes mainly from the studies of Holdaway et al. (1982, 1988), Guidotti et al. (1983), Guidotti (1989), and Dickerson and Holdaway (1989), plus numerous references cited therein. Guidotti and Cheney (1989) have summarized the development of concepts concerning metamorphism in west-central Maine. They have also provided a very complete list of references on the region.

In describing mineral assemblages we list only the AFM minerals; unless otherwise stated, all specimens also contain muscovite, quartz, ilmenite, and graphite, and commonly plagioclase, tourmaline,  $\pm$  pyrrhotite.

**M1, Regional Greenschist Facies**

Much of Maine has suffered a regional, syn-tectonic low-grade metamorphism characterized by chlorite and locally calcite and/or ankerite. Some of these rocks contain biotite, which may have been produced by this event. In areas that have *not* been consistently at higher than biotite grade, metamorphism is identified as M1.

**M2, Staurolite-Andalusite Regional Metamorphism**

This event is characterized by the widespread occurrence of post-tectonic andalusite + staurolite + biotite and, at higher grades, andalusite + biotite. In almost every occurrence of M2, the rocks show extensive retrogression to lower-grade assemblages or progressive metamorphism to higher-grade assemblages. Retrogressive metamorphism is prevalent even where later metamorphic events are not obvious in the area. Occurrence of nearby M3 has a hinge effect on M2, either down-grading or up-grading the mineral assemblage, but typically leaving clear textural evidence of M2, except where M3 is in the Upper Sillimanite Zone.

M2 is best developed near the Lexington batholith (Figs. 1, 4) where Dickerson and Holdaway (1989) described three sub-events from oldest to youngest: M2n - hornfels containing cordierite + andalusite + biotite grading to cordierite + sillimanite + K feldspar + biotite as contact metamorphism around the north lobe of the batholith; M2 - andalusite + staurolite + biotite  $\pm$  garnet and andalusite + biotite  $\pm$  garnet regionally developed around the south lobe of the batholith, and M2s - andalusite + staurolite + biotite  $\pm$  garnet, andalusite + biotite  $\pm$  garnet, and sillimanite + biotite  $\pm$  garnet as contact metamorphism around the south and central lobes of the batholith (Fig. 4). M2 and M2s are nearly synchronous in age, but M2n appears to be older than M2 and M2s. Most of the M2 rocks, less of

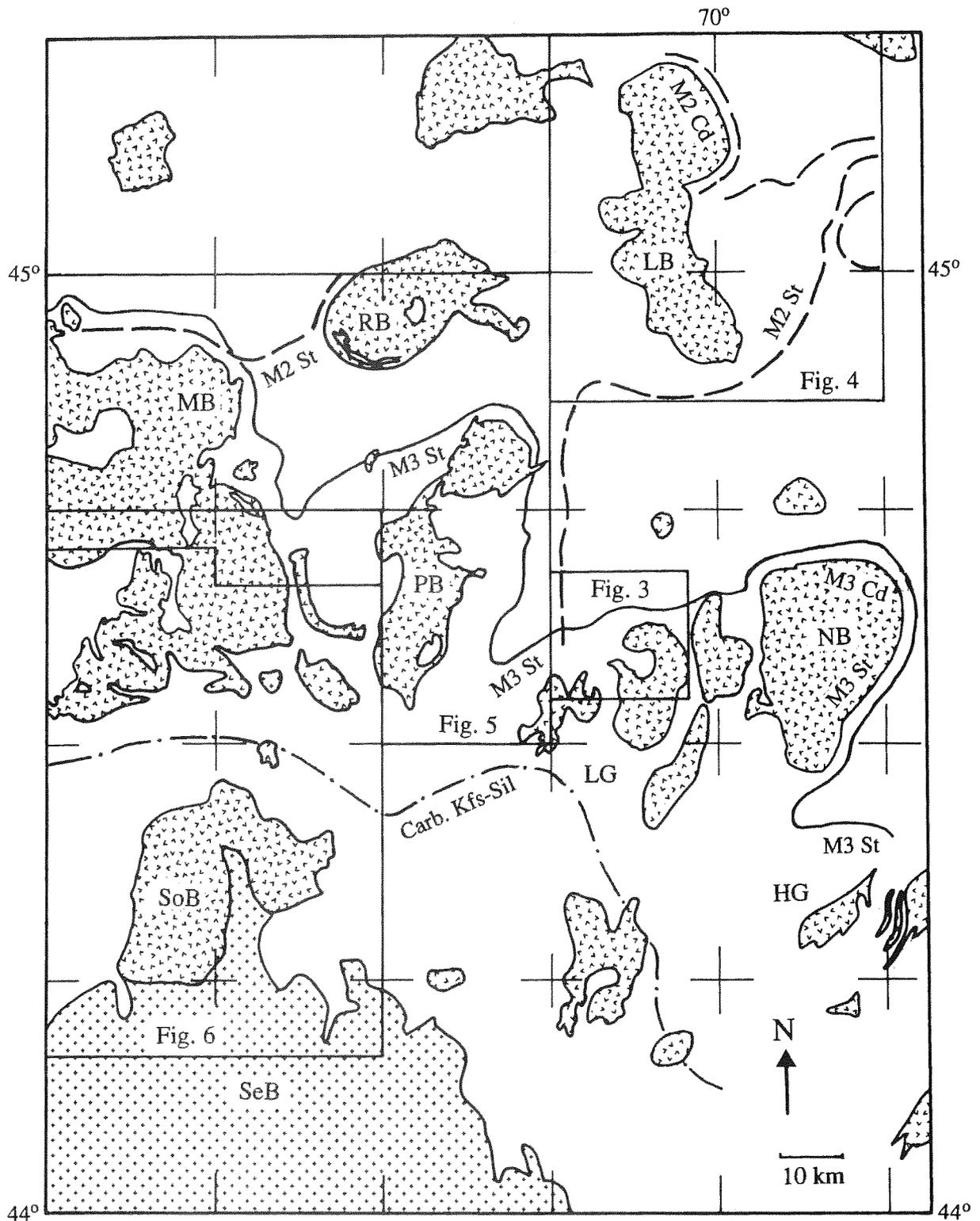


Figure 1. Metamorphic map of west-central Maine. Isograds shown are M2 staurolite (St) or cordierite (Cd) - dashed, M3 staurolite or cordierite - solid, and Carboniferous K feldspar-sillimanite (Kfs-Sil) - dash-dot. Solid NS and EW lines outline areas of Figures 3-6. Check pattern - Acadian granitic rocks; plus pattern - Carboniferous granitic rocks. Igneous abbreviations include RB - Reddington batholith, LB - Lexington batholith, NB - Norridgewock batholith, MB - Mooselookmeguntic batholith, PB - Phillips batholith, LG - Livermore Falls group of plutons including the North Jay pluton (SW corner of Figure 3), HG - Hallowell group of plutons including the Hallowell (SE Augusta quad.) and Togus (SE Vassalboro quad.) plutons, SoB - Songo batholith, SeB - Sebago batholith. Note that we are assuming contact metamorphism around the Reddington and Lexington, batholiths is approximately correlative with regional M2, and that around the Norridgewock batholith is approximately correlative with regional M3 (see text).

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the M2s rocks, and still less of the M2n rocks have been extensively retrograded to assemblages containing mainly muscovite and chlorite.

M2 is widely exposed in western Farmington and Kingfield quadrangles, and in much of the Dixfield, Phillips, Rangeley, Rumford, and Oquossoc quadrangles (Figs. 1, 5) which have not been affected by igneous activity, high-grade M3, or Carboniferous metamorphism. Locally, M2 isograds are clearly truncated by isograds of M3; this is most obvious in the southwest part of the Farmington quadrangle (Fig. 3). Most of these M2 rocks are extensively retrograded or prograded by M3, but appear to have originally contained assemblages such as staurolite + biotite + garnet, andalusite + staurolite + biotite ± garnet, and possibly andalusite + biotite ± garnet. Contact metamorphism around the Reddington batholith (Figs. 1, 5) produced hornfels assemblages much like those around the north lobe of the Lexington batholith (M2n), with cordierite, andalusite, and sillimanite, and very little staurolite or garnet. This metamorphism may also be M2, but could be younger as discussed below. M2 is also developed in Augusta quadrangle, as discussed in the section on M3.

The following idealized list of mineral assemblages represents the sequence of increasing grade in M2 (excluding M2n and Reddington batholith contact metamorphism). Assemblages on the same line represent bulk compositional variation.

- a. Biotite + chlorite
- b. Garnet + biotite + chlorite (not common)
- c. Chloritoid + biotite ± garnet (not common)
- d. Staurolite + biotite + chlorite + garnet (rare)
- e. Staurolite + biotite + garnet, Andalusite + staurolite + biotite ± garnet, Andalusite + biotite ± garnet
- f. Andalusite + biotite ± garnet
- g. Sillimanite + biotite ± garnet

Assemblage c occurs in the Carrabassett Fm. only, and is probably stabilized by low  $P$  and high Fe. Assemblage g is only seen in close proximity to intrusive igneous rocks. In assemblages e, Mg/(Mg + Fe) in all phases and Mn in garnet increase from left to right. Staurolite in assemblage e occurring *without* andalusite has Mg/(Mg + Fe) = 0.09 to 0.15 and that *with* andalusite has Mg/(Mg + Fe) = 0.17. Biotite in assemblage f overlaps all the biotite compositions of the subassemblages in e indicating that staurolite has broken down thermally by stage f. Assemblages e and f are most commonly seen and are clearly indicative of M2 (Fig. 4).

In the Lexington batholith area M2 rocks are pseudomorphed to various degrees to assemblages containing muscovite and chlorite. In other areas M3 readjustment matches the grade of the earlier M2. Possibilities of M3 minerals superimposed on M2 include chlorite + muscovite, biotite, garnet, or even staurolite. Chlorite + muscovite and/or biotite replace biotite, garnet, staurolite, or andalusite as pseudomorphs. If the M3 mineral is garnet, garnets may show two generations of growth. If the M3 mineral is staurolite, it grew as small crystals partially replacing former M2 staurolite within a pseudomorph largely consisting of layer silicates.

The  $P$ - $T$  conditions of M2 are difficult to evaluate due to the highly retrograded character of most of the rocks. Dickerson and Holdaway (1989) estimated  $P = 2.35$  kbar,  $T_{\max} = 605$  °C for M2n based on synchronous first appearance of sillimanite and K feldspar. They estimated  $P = 2.7$  kbar,  $T_{\max} = 570$  °C for M2 and  $P = 2.75$  kbar,  $T_{\max} = 590$  °C for M2s (Fig. 2) based on the breakdown of staurolite to andalusite-bearing assemblages and the fact that conditions of M2 and M2s must lie at  $P$  between that of M2n and that of M3 (3.1 kbar).

The age of M2 is known with certainty only from one area, that of the Lexington batholith, where Gaudette and Boone (1985) obtained whole-rock and mineral isochron Rb/Sr ages of 399 Ma. Smith and Barreiro (1990) found evidence for similar ages with monazite U-Pb dating of staurolite-grade rocks which presumably experienced both M2 and M3. The only available age for the Reddington batholith is a 365 Ma *minimum* age based on biotite  $^{40}\text{Ar}/^{39}\text{Ar}$  by D.R. Lux (unpublished). In other areas where M2 is developed, it can only be stated that M2 occurred prior to M3 (ages given below).

Heat sources for M2, both regional and contact, appear to have been the various S-type and biotite-bearing granitic plutons of the region. In general, it appears that sill-like bodies of magma extended laterally either below or above the staurolite-andalusite regional metamorphic rocks. Magmatic rocks may have been as close as 1 km from the M2 rocks. Areas where the grade of M2 drops rapidly below staurolite grade probably mark the limits of these sheets of magma. This relationship is clear for the Lexington batholith, where Stewart (1989) has suggested that the

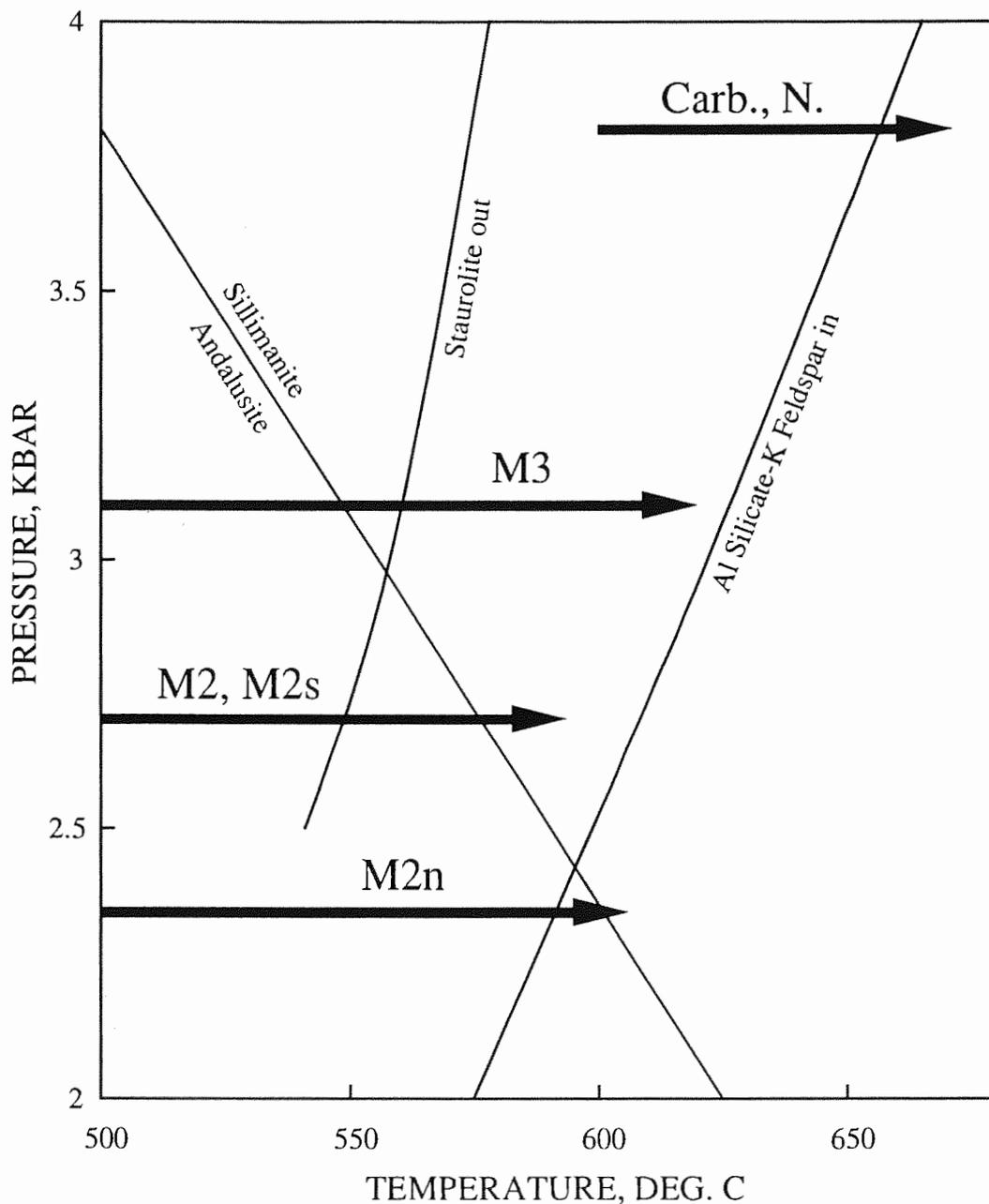


Figure 2. Approximate  $P$ - $T$  conditions for the various metamorphic events shown in Figure 1 and discussed in the text. Note that although the  $P$  variation with  $T$  for each event is shown as constant, the actual  $P$  variation depends on the compass direction of the metamorphic field gradient, and may increase, decrease, or stay the same. Conditions for the Carboniferous metamorphism are for the north side of the Sebago batholith (Fig. 1). Data from Holdaway et al. (1988) and Dickerson and Holdaway (1989).

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lack of geophysical evidence for extension of the southern lobe beneath M2 (Fig. 4) may be explained by lateral extension of magmatic rocks (now safely eroded away) *above* the M2 regional rocks. It is tempting to call on such downward metamorphism to explain the extensive hydrothermal system (Dickerson and Holdaway, 1989) necessary to retrograde M2 in the Lexington batholith area. A 1 km-thick granite sheet would also have provided the necessary *P* increase to explain the *P* difference between slightly earlier M2n and M2 and M2s.

The heat sources for other areas of M2 are less obvious, but extending the possible relations described for the Lexington batholith area, we suggest that lateral extensions of the Reddington batholith and early phases of the Phillips batholith may have provided heat sources at appropriate times and places for M2 to form. Finally, it should be clear that M2 need not have been everywhere synchronous. However, the general *P* increase that in most areas correlates with the inception of M3 probably marked the end of M2 throughout the region.

### M3, Staurolite-Sillimanite Regional Metamorphism

M3 is characterized by the widespread occurrence of post-tectonic garnet + biotite  $\pm$  chlorite, staurolite + biotite + garnet  $\pm$  chlorite and, at still higher grades, sillimanite + biotite + garnet. This thermal transition from staurolite to sillimanite clearly distinguishes M3 from M2, in which andalusite follows staurolite. Except for small amounts of late (?) chlorite, rocks of M3, rarely show any retrogressive effects. In general, areas of M3 exposure lie south of areas of M2 exposure, but there are many places where M3 truncates or is superimposed on M2. There are no localities where M2 alone can be found south of M3.

M3 is distributed across the region in four lobes of activity, from west to east, the Mooselookmeguntic lobe, the Phillips lobe, the Livermore Falls lobe, and the Hallowell lobe, (Fig. 1). Each of these lobes of M3 is also a center of plutonic activity. The Mooselookmeguntic lobe includes the Mooselookmeguntic batholith, the Phillips lobe includes the Phillips batholith, the Livermore Falls lobe includes several plutons in southern Farmington and northern Livermore quadrangles, including the North Jay pluton (Figs 1, 3), and the Norridgewock batholith (Fig. 1; designated Rome batholith by Moench and Pankiwskyj, 1988), and the Hallowell lobe includes the Hallowell and Togus plutons (Fig. 1). The M3 isograds are clearly related to these igneous bodies, although in some places lateral subsurface (?) extensions of the magmatic rocks are necessary to explain the distance between the staurolite isograd and the plutonic contacts. Between each of these lobes of magmatic activity is a sharp reentrant of lower-grade rocks (Fig. 1). The presence of these sharp reentrants provides for the possibility that adjacent lobes of igneous-metamorphic activity were not synchronous, but were emplaced during a time interval when the *P* was roughly constant throughout the region of M3.

There are more similarities than differences between the various lobes of M3. The Mooselookmeguntic and Phillips lobes are characterized by occasional sievy porphyroblasts of andalusite believed to be remnants of M2 which is exposed in areas of lower-grade M3. These andalusites are typically partly or wholly replaced by coarse-grained muscovite. The Livermore Falls lobe has very little andalusite because M2 was sub-staurolite grade throughout much of the Livermore Falls lobe. In the Farmington quadrangle, a few rare M3 localities which do contain andalusite are on the north and northeast edges of the metamorphic high and probably represent local areas of lower *P* during M3.

The Augusta quadrangle, primarily in the Hallowell lobe (Fig. 1), has sievy andalusite porphyroblasts in about half of the localities in the M3 staurolite and sillimanite zones. This area is unique in west-central Maine in that it has a few localities like West Sidney (Osberg, 1971), which contains andalusite + staurolite + cordierite + biotite + Mn garnet + chlorite + muscovite + quartz. No other locality in Maine, besides those in the Augusta quadrangle, contains staurolite + cordierite + muscovite + quartz. While there appear to be at least three heat sources in the Augusta quadrangle, there are no isograds that can clearly be distinguished as M2 as defined by Dickerson and Holdaway (1989). Novak and Holdaway (1981) used textural and phase rule analysis to suggest that the West Sidney locality, and others like it, are polymetamorphic. Our interpretation is that the Augusta area was undergoing magmatism and metamorphism during a *P* increase from about 2.4 kbar (similar to M2n) to 3.1 kbar (typical of M3), such that cordierite + andalusite + biotite were stable during the early stages and staurolite + andalusite + biotite were stable late in the process. Presumably, chlorite developed as an early retrogressive mineral to complete the sequence of events. In this area, the effects of M2 occurred early and those of M3 occurred later, and there was no *T* decrease between the two events. Because of this, textures are more difficult to interpret than in areas of *T* decrease between M2 and M3. Hence, M2 and M3 isograds coincide.

The Norridgewock batholith has an M3 age (see below) but the northeast contact zone, near the town of Norridgewock, contains cordierite to the exclusion of staurolite and garnet, and the southeast contact zone contains staurolite and andalusite. Apparently  $P$  in M3 decreased to the northeast.

An idealized grade sequence of M3 mineral assemblages (excluding the Norridgewock area rocks) is as follows:

- a. Biotite + chlorite
- b. Garnet + biotite + chlorite
- c. Staurolite + biotite + garnet + chlorite
- d. Staurolite + biotite + garnet, Andalusite (or chlorite) + staurolite + biotite + garnet, Cordierite + andalusite + biotite (Mg-rich, sulfidic)
- e. Sillimanite + staurolite + biotite + garnet
- f. Sillimanite + biotite + garnet

Garnet occurs in most rocks (excluding a) except the sulfidic cordierite schists. Muscovite and quartz, without K feldspar, were stable right up to pluton contacts. Chlorite is a common retrograde mineral in minor amounts.

CVG and MJH disagree concerning the stability of chlorite vs. andalusite in staurolite-grade rocks of M3. In MJH's opinion, relict M2 andalusite, preserved during M3, also became stable during M3, and produced local Mg-richer compositional domains. Staurolite and biotite locally adjusted their compositions to be in equilibrium with the remnant andalusite. MJH has measured staurolite compositions in andalusite-free rocks of M3 assemblage d with  $Mg/(Mg + Fe) = 0.11$  to  $0.16$ , and in M3 andalusite-bearing rocks of assemblage d with  $Mg/(Mg + Fe) = 0.17$  to  $0.19$ . By this scenario, most minor Staurolite Zone chlorite is retrograde, and the only prograde chlorite is that which occurs with staurolite near the beginning of the zone. Sillimanite formed from reaction of staurolite + muscovite + quartz and of andalusite at about the same time. Arguments in support of these ideas are given by Holdaway et al. (1988), and Holdaway and Dutrow (1989).

In CVG's opinion, most of the chlorite in Staurolite Zone rocks is prograde and remnant M2 andalusite remained out of equilibrium with the surrounding minerals during M3. Sillimanite first formed from reaction of staurolite + chlorite + muscovite + quartz and of metastable andalusite. Arguments in support of this conclusion are given by Guidotti (1974) and by Guidotti et al. (1991).

Holdaway et al. (1988) estimated average M3  $P$ - $T$  conditions of non-cordierite-bearing rocks by comparing the results of garnet-biotite geothermometry on assemblages d and e with the sillimanite-andalusite phase boundary. Using this method, the first sillimanite occurs at about  $550^{\circ}\text{C}$  and  $3.1$  kbar. For M3, the estimated  $P$  is  $3.1$  kbar, and  $T_{\text{max}}$  is  $620^{\circ}\text{C}$  (Fig. 2)

Age dates are available for several of the plutons (Fig. 1) correlated with M3, listed in decreasing age sequence:

Togus	394 Ma	Rb-Sr	Dallmeyer and VanBreeman (1981)
Hallowell	387 Ma	Rb-Sr	Dallmeyer and VanBreeman (1981)
Songo	382 Ma	U-Pb	Lux and Alcinikoff (1985)
Mooslookmeguntic	371 Ma	Rb-Sr	Moench and Zartman (1976, corrected by D.R. Lux)
Phillips	371	$^{40}\text{Ar}/^{39}\text{Ar}$	DeYoreo et al. (1989, minimum age)
Norridgewock	369 Ma	U-Pb Monazite	D.R. Lux (unpublished)
North Jay (Fig. 3)	356 Ma	U-Pb Monazite	D.R. Lux (unpublished)

The 394 and 387 Ma ages of the Togus and Hallowell plutons are consistent with the Hallowell lobe showing the transition between M2 and M3 in both time and  $P$  as discussed above. The remaining ages suggest that M3 varied in age from place to place.

For most of the plutons, the M3 isograds hug the contacts fairly closely. However, it appears that the Phillips pluton must have extended considerable distances (above or below the present erosion level) both east and west in order to provide sufficient localized heat to extend M3 staurolite and sillimanite zones several km from the contacts. Maps of igneous bodies one to two km above the present level and one to two km below the present level would have looked quite different from the map shown in Figure 1. The approximately  $0.4$  kbar pressure increase of M3 over M2 is explained partly by post-metamorphic tilting of the region down to the south and partly by plutonic

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magmas spreading laterally over the M3 terrain and/or producing volcanism on the surface. Presumably there was no  $P$  increase in the northeast portion of the Norridgewock batholith where cordierite persisted in M3.

### Carboniferous, K Feldspar-Sillimanite Regional Metamorphism

The only regionally developed K feldspar with sillimanite occurs in the southern part of the region, south of all the other metamorphic events (Evans and Guidotti, 1966). The second sillimanite isograd roughly parallels the contact of the large Sebago batholith, but lies 15 to 30 km to the northeast of the batholith (Fig. 1). South of the Sebago batholith, post-tectonic kyanite is present.

An idealized list of mineral assemblages of rocks which experienced late Carboniferous metamorphism is as follows:

- a. Kyanite + staurolite + biotite + garnet (south of Sebago batholith)
- b. Sillimanite + biotite + garnet (north of Sebago batholith)
- c. K feldspar + sillimanite + biotite + garnet (north of Sebago batholith)

Almost every pelitic specimen contains muscovite, but about half of those north of the batholith contain no K feldspar. Outcrops containing assemblage b are randomly interspersed with those containing assemblage c (Holdaway et al., 1988). Migmatitic rocks are also widespread at grades above the second sillimanite isograd. There appears to be a tendency for the metamorphism to plateau at or near the conditions of the K feldspar-sillimanite isograd. Perhaps also, metamorphic grade drops off rapidly to the northeast, leaving the rest of M3 relatively less affected by Carboniferous metamorphism. The major Carboniferous effect northeast of the second sillimanite isograd is the development of retrograde chlorite in many specimens collected in this area. South of the batholith, assemblage a indicates higher  $P$  and perhaps lower  $T$  than on the north side (Thomson and Guidotti, 1989).

Holdaway et al. (1988) found minor differences between sillimanite + biotite + garnet assemblages of M3 northeast of the second sillimanite isograd and identical assemblages of the Carboniferous metamorphism southwest of the second sillimanite isograd. Biotite in the Carboniferous assemblage contains more Ti than in M3, and both  $P$  and  $T$  of the Carboniferous assemblage was higher than in M3. Textural and compositional evidence suggests that the muscovite which formed during the Carboniferous metamorphism is prograde (Evans and Guidotti, 1966).

Holdaway et al. (1988) measured garnet-biotite  $T$ 's for assemblage b of  $640 \pm 17$  °C and for assemblage c of  $659 \pm 13$  °C, consistent with the absence of K feldspar in assemblage b. Thus small local variations in  $T$  could explain the observed variations in mineral assemblage. Small local variations in  $a(\text{H}_2\text{O})$  as suggested by Evans and Guidotti (1966) and Cheney and Guidotti (1979) are also a possible explanation for the difference between assemblages b and c. Holdaway et al. (1988) also calibrated the muscovite-almandine-biotite-sillimanite geobarometer to determine average  $P$  of  $3.8 \pm 0.2$  kbar for assemblage b and  $4.9 \pm 0.8$  kbar for assemblage c. This geobarometer has been recalibrated by McMullin et al. (1991), and use of this newest calibration will change these numbers. Thomson and Guidotti (1989) estimated  $P$  of about 6.5 kbar for assemblage a on the south side of the batholith.

This Carboniferous metamorphism has been dated by dating the correlative Sebago batholith. Hayward and Gaudette (1984) obtained an age of 325 Ma and Aleinikoff et al. (1985) obtained an age of 324 Ma. A more recent unpublished age determination by D.R. Lux of 296 Ma suggests that the metamorphism may be Alleghenian in age.

Lux and Guidotti (1985) demonstrated the cause and effect relationship between the Sebago batholith and the second sillimanite isograd to the north (Fig. 1), using geologic and petrologic evidence, and  $^{40}\text{Ar}/^{39}\text{Ar}$  hornblende cooling ages. The widespread occurrence of these highest-grade rocks in west-central Maine is a Carboniferous age feature caused by heat from cooling of the sheet-like Sebago batholith. Hodge et al. (1982) have shown with gravity studies that the batholith is thin and dips gently northeast. Guidotti et al. (1986) and DeYoreo et al. (1989) have modelled the thermal effects of a thin Sebago batholith dipping gently northeast and have shown that such an intrusion is adequate to produce the observed metamorphism in the overlying rocks. The levelling of metamorphic grade and the widespread occurrence of muscovite and of migmatites probably represent a buffering effect by the heat and water from the crystallizing batholith below. The nature of this Carboniferous metamorphism is thus quite similar to that of M2 and M3, which were produced by laterally spreading Devonian plutons farther north. Because the Sebago batholith dips northeast, the rocks of assemblage a on the south side of the batholith must have underlain the batholith before erosion (Thomson and Guidotti, 1989).

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## SUMMARY

The three major metamorphic events in west-central Maine are thus all produced by heat from cooling plutons over an age range from 400 to 296 Ma. In areas where the isograds do not conform to the plutonic contacts, the discrepancies probably relate to lateral spread of magma either below or above the presently exposed medium-grade metamorphic rocks. The petrologic differences between M2, M3, and Carboniferous metamorphism relate primarily to variation in  $P$  (or depth) of metamorphism. The facts that successively younger events were developed successively further to the south-southwest, younger events superimposed on older ones exhibit higher  $P$ , and  $P$  increased to the south-southwest within each event suggest that (1) the region has been tilted downward to the south-southwest during and since metamorphism and (2) apart from this tilting,  $P$  has increased at any given point during the age range of these events, presumably from emplacement of magmas at levels shallower than present exposures or from surface deposition of volcanic rocks. The total range of  $P$  over the time period of these events was from 2.35 to 6.5 kbar or about 14.5 km relative difference in depth.

## ACKNOWLEDGEMENTS

MJH acknowledges research support from the National Science Foundation, grant EAR-8904777, and CVG acknowledges support from National Science Foundation grant EAR-8817066. Much of the field-related study for the MJH portion of this project was done by Barb Dutrow (Farmington quadrangle) and Bob Dickerson (Lexington batholith). We appreciate the guidance provided by Dan Lux regarding the ages of plutons in the area.

## ROAD LOG FOR SATURDAY, OCTOBER 23, 1993

We will depart at 8:00 am from the Farmington Motel. Today we will examine outcrops in two areas, mainly M3 rocks of the Farmington 15' quadrangle (Fig. 3) and M2 rocks of the Lexington batholith area (Fig. 4). Detailed geologic maps are available for these areas: (Pankiowskyj, 1971, for the Farmington quadrangle; Pankiowskyj, 1979, and Boone, 1973 for the Lexington batholith area). General geologic relationships can be seen from the maps of Osberg et al. (1985) and Moench and Pankiowskyj (1988)

Proceed WEST 1.8 miles on Route 2 and SOUTH 7.7 miles on Routes 2 and 4 to the separation of Routes 2 and 4 in Wilton, where the trip begins.

**Mileage**

0.0 Proceed WEST on Route 2.  
0.2 Road cuts on left and right. Park on right beyond cut.

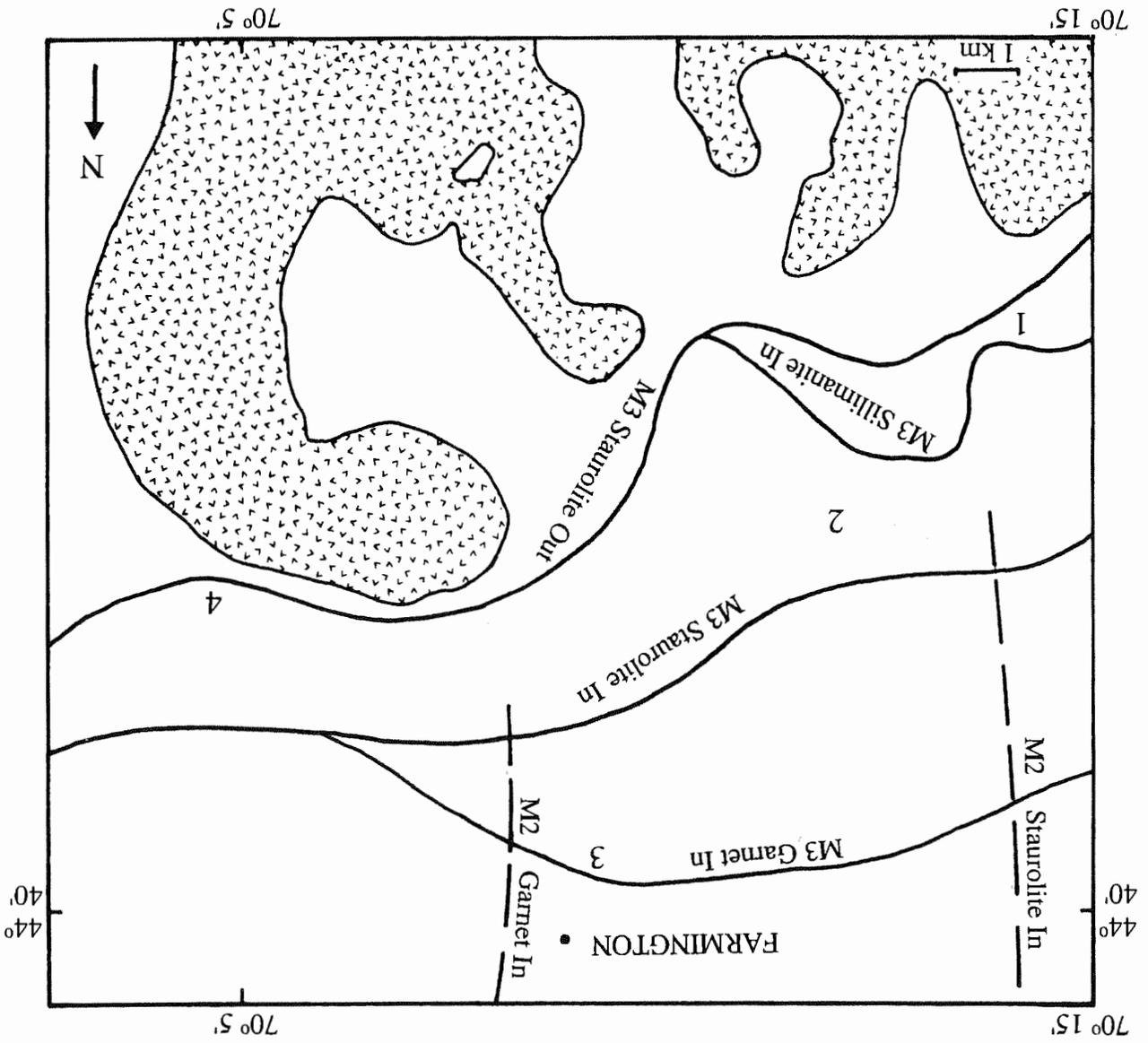
**STOP 1. M3 LOWER SILLIMANITE ZONE ROCKS OF THE SANGERVILLE FM. (25 MINUTES).** (Farmington quadrangle). Pelites containing sillimanite + staurolite + biotite + garnet are interbedded with biotite-rich psammitic rocks. Small biotite porphyroblasts and larger euhedral garnets are set in a matrix of partially oriented muscovite. Staurolite has partially reacted out, and sillimanite is visible in some hand specimens. The pelitic schists locally contain coarse muscovite pseudomorphs after staurolite (spangles) described by Guidotti (1968). In such specimens, most of the staurolite occurs as anhedral remnants within single-crystal muscovite pseudomorphs. Spangles are best seen near the middle of the exposure on the right side.

Stop 1 is 1.5 km north of the North Jay pluton, and the M3 isograds trend east-northeast in this area closely paralleling contacts of the plutons (Fig. 3). The culmination of this Livermore Falls lobe of metamorphism is associated with several small plutons in southern Farmington quadrangle, where grade reaches the Upper Sillimanite Zone (sillimanite + biotite + garnet). K feldspar is rarely, if ever, seen in M3.

Five km north-northwest of here in the M3 Garnet Zone, the rock is retrograded M2 staurolite schist with new garnet. M2 staurolite and garnet are replaced by mixtures of chlorite, muscovite, and biotite. The M3 garnet is subhedral and normally does not replace M2 garnet pseudomorphs. Both the M2 and M3 porphyroblasts are post-kinnematic. Retrograded and partially retrograded M2 staurolite is also seen near Varnum Pond and Temple in northwest Farmington quadrangle. The M2 staurolite isograd trends north 2-3 km east of the west boundary of the Farmington and Kingfield quadrangles (Figs. 1, 3) and is truncated on the south by M3. Turn vehicles around and return to the Route 2 and 4 separation point.

0.4 Proceed NORTHEAST on Routes 2 and 4.

Figure 3. M2 (dashed) and M3 (solid) isograds and field trip stops (1-4) in southern Farmington quadrangle (Fig. 1). Plutons (check pattern) are Acadian in age; North Jay pluton is in the SW corner of the map. Data from Dutrow (1985).



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- 3.0 Extensive road cuts on right. Park before cuts.

**STOP 2. M3 STAUROLITE ZONE PELITE OF THE SANGERVILLE FM. (25 MINUTES).** (Farmington quadrangle). Most of the rocks are staurolite + biotite + garnet schist with retrograde chlorite. The chlorite is locally abundant in greener portions of the road cut, and might be an overprint from the Carboniferous metamorphism 16 km south of here. Staurolite and garnet typically form anhedral porphyroblasts in a matrix of fine oriented muscovite and coarser, less oriented biotite. Staurolite crystals often grew in polycrystalline muscovite pseudomorphs which are similar to retrograde pseudomorphs of M2 staurolite found elsewhere. We have no direct evidence that M2 staurolite formed this far east; the pseudomorphs might also be of chloritoid, which could have formed in the M2 Garnet Zone.

Most M3 rocks in the Farmington quadrangle contain no andalusite, because M2 did not reach andalusite grade in the area. Continue northeast on Routes 2 and 4.

- 3.7 Road cuts on left and right are Sangerville Fm. mica schists, biotite quartzites, and calc-silicate rocks.  
6.0 Road cuts and bulldozer cuts of staurolite + biotite + garnet of Sangerville Fm. Chloritic alteration is more abundant in thin section than at Stop 2.  
7.5 Park on left across from road cuts in West Farmington..

**STOP 3. M3 GARNET ZONE ROCKS OF THE SANGERVILLE FM. (25 MINUTES).** (Farmington quadrangle). Most of the rocks in the road cut are biotite ± chlorite phyllite, but local horizons contain garnet. Muscovite and chlorite are fine grained and oriented, while biotite is slightly coarser and randomly oriented. Some garnets from this locality and others in the Garnet Zone of M2 and M3 (south and west of here) show two isolated stages of garnet growth with inclusion-filled, anhedral Mn-Ca-rich cores, locally an irregular graphitic boundary, and clear, euhedral Fe-rich rims (Dutrow, 1985). Chlorite is variable in amount and prograde. East of here the M3 Garnet Zone pinches out and grade in M2 drops to Biotite Zone (Fig. 3). Continue on Routes 2 and 4.

- 8.1 Take Route 2 SOUTHEAST, as Route 4 splits to the north at the Sandy River.  
11.8 Small road cuts on left are vesuvianite-bearing calc-silicate rock in calcareous horizons of the Sangerville Fm., and a granitic dike.  
12.5 Turn RIGHT on Routes 41 and 156 to Farmington Falls.  
12.8 Park near fire station on left and walk to bridge, northwest side.

**STOP 4. M3 STAUROLITE-ANDALUSITE ROCKS OF THE SANGERVILLE FM. (25 MINUTES).** (Farmington quadrangle). These schists contain staurolite + biotite + garnet with local coarse andalusite porphyroblasts. Muscovite and biotite are fine grained and oriented parallel to schistosity. In some areas, late chlorite appears to have replaced biotite. Subhedral garnet and euhedral staurolite porphyroblasts have grown after deformation. Where present, the andalusite is subhedral, very coarse, and relatively free of inclusions. Traces of sillimanite were observed in one thin section. Staurolite, biotite, and garnet with the andalusite are compositionally near the Mg-rich limit for M3.

The rare occurrences of andalusite in the Farmington quadrangle are all on the north or northeast side of the Livermore Falls lobe, suggesting that *P* decreased slightly in these directions during M3. Thirty km east-northeast of here at the northeast corner of the Norridgewock batholith, andalusite and cordierite are common and staurolite is absent. Metamorphic zones become narrower as well (Figs. 1, 3). The area is a lower-*P* portion of M3 as confirmed by Dan Lux's monazite age of 369 Ma on the Norridgewock batholith. Return to Route 2, continuing east. This is a good time for a snooze!

- 16.9 Turn LEFT (north) on Route 134 in New Sharon before the bridge over Sandy River. Most of the route between here and Stop 5 is over M1 Chlorite- and Biotite-Zone rocks.  
24.8 Outcrops in stream are quartzite and chlorite phyllite.  
25.2 Turn RIGHT (east) on Route 43.  
26.2 Road cuts on left and right to 26.6 mi. are quartzite and biotite-chlorite phyllite.  
29.5 Road cuts on right are Old Point pluton, a small northern satellite of the Norridgewock batholith.  
33.3 Without crossing the Kennebec River, proceed STRAIGHT (north) on Route 201A in Anson.  
37.9 In North Anson, extensive outcrops west of bridge over Carrabassett River are quartzite, biotite quartzite, and slightly calcareous rocks of the Fall Brook Fm. (Madrid equivalent).

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- 44.6 Cross the Kennebec River. The M2 staurolite isograd occurs about 1 km north of here, trending north-northeast.
- 45.7 In Solon, turn LEFT (north) on Route 201. Outcrops in stream are Fall Brook Fm. quartzites and biotite quartzites.
- 46.4 Turn LEFT (west) on Falls Road. Proceed straight on poorer quality road to Arnold's Landing, and park.

**STOP 5. M2 STAUROLITE ZONE, CARRABASSETT-FALL BROOK FMS. (25 MINUTES).** (Anson quadrangle). On the left are Fall Brook Fm. (?) quartzites and pelitic quartzites containing biotite and 2-cm andalusite. On the right are Carrabassett Fm. pelites (with silty layers) containing andalusite + staurolite + biotite + garnet and staurolite + biotite + garnet. Staurolite and andalusite stand out on the weathered surfaces. Altered porphyroblasts of andalusite, staurolite and garnet occur in a matrix of fine crenulated muscovite and slightly coarser, unoriented biotite. Andalusite is largely to entirely replaced by very fine muscovite (or sericite). Staurolite is partly replaced by very fine muscovite and some chlorite. Garnets are partly to fully replaced by chlorite. In some sections, late coarse chlorite cuts across the foliation and includes graphite inclusions parallel to the foliation. Coarse polycrystalline muscovite forms prism-shaped pseudomorphs which may have once been chloritoid.

As seen here and elsewhere, andalusite is *much* more common in M2 than in M3. In M2 of the Lexington batholith region at staurolite grade, Carrabassett Fm. pelites exhibit layer by layer variation from (1) staurolite + biotite + garnet to (2) andalusite + staurolite  $\pm$  garnet, and Fall Brook (and Madrid) Fm. semi-pelites and pelites exhibit layer by layer variation from (2) andalusite + staurolite + biotite  $\pm$  garnet to (3) andalusite + biotite  $\pm$  garnet (Fig. 4). Where the degree of alteration permits analysis, all minerals show increase in Mg/Fe and garnet shows increase in Mn from (1) to (2) to (3). The variation is clearly related to bulk composition; possibilities include increasing ankeritic dolomite or kaolinite/illite ratio in the original sediments with increasing Mg/Fe in the metamorphosed equivalent. Redox and sulfides are not factors as all opaques are hematite-free ilmenite, and graphite is ubiquitous but minor in amount.

Chlorite-grade phyllites are exposed 2 km east-southeast of here on the east side of Solon. Seven km west of here there is a rapid increase in grade to sillimanite + biotite + garnet (M2s) near the south and central lobes of the Lexington batholith (Fig. 4). The heat source for regional M2 was probably a lateral extension of the south lobe, perhaps extending *over* the present locality; this would also explain the retrogression. Return to Route 201 and proceed left (north).

- 48.7 LUNCH STOP, Arnold's Way Rest Area.
- 48.9 45th parallel. Elevation is 420 feet, not 352 feet.
- 49.6 Road cuts on right of Madrid Fm. psammites.
- 50.6 Road cuts on right of Carrabassett Fm. staurolite schist.
- 54.1 Turn LEFT (west) on Route 16 in Bingham. After crossing bridge, turn right (north) at 54.3 mi. Road cuts on left are Smalls Falls Fm. sulfidic-pelitic schists. Andalusite is rare.
- 55.8 Road cuts on left for next 0.8 mi. are Perry Mountain Fm. muscovite-rich pelites and Smalls Falls Fm.
- 58.5 Take LEFT fork toward Rowe Pond.
- 60.4 Drop off passengers opposite first road cuts on left, drive vehicles 0.25 mi. north and park near first pair of stone pillars on right.

**STOP 6. M2 LOWER STAUROLITE ZONE CARRABASSETT FM. PELITES (30 MINUTES).** (Bingham quadrangle). These rocks, from the beginning of the Staurolite Zone, are some of the least-altered rocks of regional M2. The primary mineral assemblage is staurolite + biotite + chlorite + garnet with local slender andalusite crystals up to 10 cm in length. **PLEASE** don't destroy the andalusite in place. Euhedral porphyroblasts of staurolite, garnet, and andalusite occur in a matrix of fine, oriented muscovite and chlorite with coarser, randomly oriented biotite. Andalusite is partly altered to fine muscovite.

This is one of the very few localities in M2 where prograde chlorite can be demonstrated in staurolite-bearing rocks. Elsewhere in M2, staurolite-bearing rocks contain andalusite + biotite instead of staurolite + chlorite as prograde Mg-rich limiting phases. Even here, staurolite + chlorite + muscovite + quartz is locally beginning to react to andalusite + biotite + H<sub>2</sub>O. The staurolite isograd here trends east-northeast, is 5 km from the central lobe of the Lexington batholith, and shows no relationship to the north lobe (Fig. 4). Metamorphic grade increases to the south. After examining these rocks, walk north 0.25 mi. to the culvert on the left opposite the first pair of stone pillars.

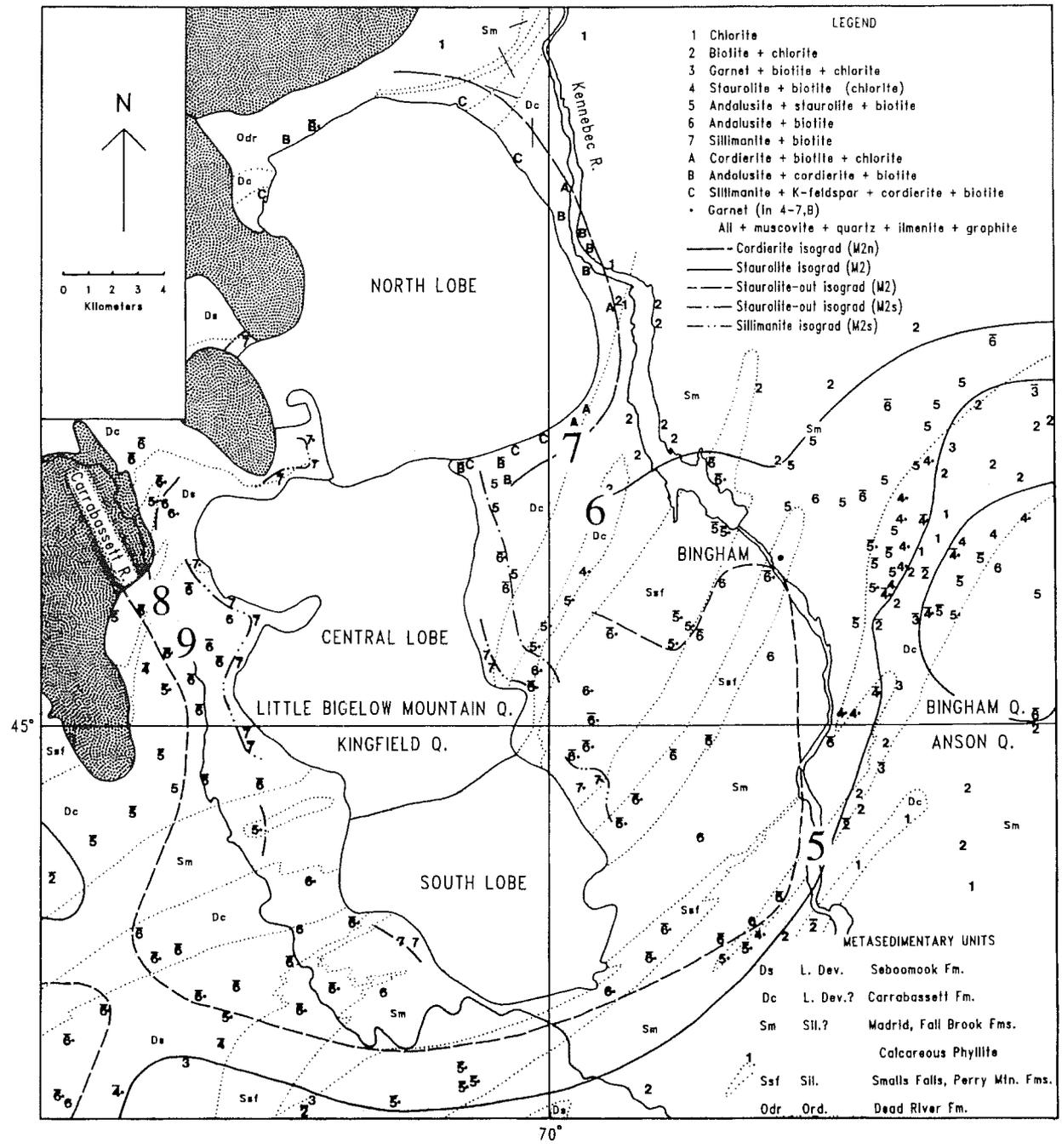


Figure 4. Metamorphic map of the Lexington batholith area (Fig. 1). Large numbers (5-9) give localities of field trip stops. Small numbers and letters indicate metamorphic assemblages as identified in Legend (upper right). Overbar - highest grade minerals more than 90% altered, isograd symbols are identified in Legend. White - Lexington batholith, stipple - older gabbroic rocks, country rock units identified at lower right. Modified from Dickerson and Holdaway (1989).

Carrabassett schists here contain chloritoid + staurolite + biotite + garnet + chlorite, transitional to the M2 Garnet Zone. Fine, oriented muscovite and chlorite form a matrix to the other minerals, which are unoriented. Euhedral chloritoid is partly replaced by chlorite. This is one of a few chloritoid localities in the area, it always occurs in the Carrabassett Fm. Chloritoid + biotite is probably stabilized by high  $Fe^{2+}$ , and low  $P$ . Proceed north-northwest.

62.7 Park just north of the red Scott marker in view of Clear Pond. THIS STOP WILL BE OMITTED IF WE USE A BUS FOR THE TRIP.

**STOP 7. M2n CORDIERITE ZONE CARRABASSETT FM. PELITES (25 MINUTES).** (Bingham quadrangle). The only cordierite found in normal pelites of the Lexington batholith area is that found adjacent to the north lobe of the Lexington batholith. South of the red marker is a glaciated outcrop of andalusite + cordierite + biotite hornfels of M2n which is difficult to sample. For fresh float samples follow the white blazes about 100 m west of the parking area, where cordierite makes ellipsoidal porphyroblasts and andalusite makes prismatic porphyroblasts in a matrix of oriented muscovite and unoriented coarser biotite.

The M2n event is a contact aureole that extends 1 to 1.5 km from the north lobe of the batholith. The rocks are only slightly retrograded, less than M2s and much less than regional M2. The sequence with increasing  $T$  toward the contact is (a) Chlorite, (b) Biotite + chlorite, (c) Cordierite + biotite + chlorite, (d) Andalusite + cordierite + biotite, (e) Sillimanite + K feldspar + cordierite + biotite (with muscovite). Staurolite is absent and garnet is very rare and Mn-rich. We think the north lobe is slightly older than the central and south lobes of the Lexington batholith, and staurolite-bearing M2s was superimposed on cordierite-bearing M2n (Fig. 4). Retrace the route south and east to the bridge at Bingham.

- 71.0 Do not cross the bridge over the Kennebec River. Continue STRAIGHT (south) on Route 16.
- 76.5 Road cuts of Madrid Fm. psammite on left over next 3 mi.
- 79.7 Route 201A joins Route 16. Continue SOUTH on Routes 16 and 201A without crossing bridge.
- 80.3 Turn RIGHT (west) on unmarked paved road.
- 81.0 Road cuts of Fall Brook Fm. quartzite. Pelitic layers contain biotite with no higher-grade minerals.
- 82.3 Road cuts on right in valley are M2 Fall Brook Fm. andalusite + staurolite + biotite + garnet schist, retrograded.
- 82.8 Cross Dunbar Hill Road. To the south 0.1 mi. is M2 Carrabassett Fm. andalusite + staurolite + biotite + garnet schist, and northeast, north, and west are M2 Carrabassett, Fall Brook, and Smalls Falls Fms. andalusite + biotite  $\pm$  garnet schists. All are significantly retrograded.
- 84.5 Turn LEFT (south)
- 84.6 Turn RIGHT (west)
- 85.8 Outcrops in this area are psammitic Fall Brook Fm. Northwest, 1.5 km is M2s Perry Mountain Fm. andalusite + biotite + garnet near the south contact of the Lexington batholith, south lobe. M2 and M2s merge in this area (Fig. 4).
- 87.1 Turn RIGHT (northwest) on Route 16.
- 88.3 Road cuts on right are psammitic Fall Brook and sulfidic Smalls Falls Fms. Three km west-northwest of here is diopsidic hornfels.
- 90.1 Road cuts on left and right are garnet + biotite + muscovite granite of the Lexington batholith, south lobe. The granite contains a few biotite-rich inclusions and streaks. West-northwest of here country rocks near the contact contain sillimanite + biotite with muscovite but no K feldspar (Fig. 4).
- 96.2 Road cuts of psammitic Madrid Fm. just west of granitic contact.
- 97.2 Glacial boulder on left of coarse, porphyritic biotite granite of the central lobe, Lexington batholith. Microcline phenocrysts are up to 10 cm long and commonly define a flow pattern.
- 98.8 In Kingfield, after crossing the Carrabassett River, turn RIGHT (north) staying on Route 16, which is now joined by Route 27. Outcrops of psammitic Madrid Fm. below bridge.
- 100.6 River outcrops of M2 Carrabassett Fm. andalusite + biotite  $\pm$  garnet schist occur discontinuously almost to Stop 8.
- 105.1 Road cuts on left of Seboomook Fm.
- 105.4 Park vehicles on right just south of Hammond Field Brook.

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**STOP 8. M2 ANDALUSITE ZONE SEBOOMOOK FM. PELITES (25 MINUTES).**

(Little Bigelow Mtn. quadrangle). Andalusite + biotite + garnet, largely retrograded are interbedded with quartzite. Andalusites are up to 15 cm in length, have a high aspect ratio, and have good crystal face development. In thin section, the andalusite porphyroblasts are seen to be nearly free of inclusions and completely replaced by very fine muscovite. Small garnet porphyroblasts are totally replaced by chlorite. The matrix is fine, oriented muscovite with slightly coarser, less oriented biotite, most of which has been replaced by chlorite.

This andalusite texture is commonly a function of proximity to igneous bodies, but here may also be a function of lithology. The Sugarloaf Gabbro is 1 km northwest, and the Lexington batholith, central lobe is 2.5 km east-northeast on the second, higher ridge back (Fig. 4).

M2 grade increases southwest to northeast from staurolite to Andalusite Zone. Andalusites closer to the contact are smaller than at Stop 8, but have the same high aspect ratio and well defined grain boundaries. Adjacent to the central lobe, is less retrograded M2s sillimanite + biotite  $\pm$  garnet with no K feldspar (Fig. 4). Proceed north.

105.6 Carrabassett Valley Rest Area. Turn around and head SOUTH.

107.0 Park vehicles on right.

**STOP 9. M2 ANDALUSITE ZONE CARRABASSETT FM. PELITES (25 MINUTES).**

(Little Bigelow Mtn. quadrangle). Andalusite + biotite  $\pm$  garnet schists here are almost entirely retrograded. Andalusites are rounded and less well defined, have a lower aspect ratio, and tend to be more sievy than at Stop 8, typical of most regional M2 andalusite. The andalusite is fully altered to patchy fine white mica. Garnet, where present is altered to chlorite. Schistosity is defined by muscovite and chlorite, which has replaced most of the biotite.

The Lexington batholith, central lobe is 2 km northeast of here. The large reentrant of country rocks east of here at Clay Brook Mtn. suggests that in this area the upper contact of the batholith is flat. However, topographic relations northwest and southwest of Clay Brook Mtn. suggest that at the outer edge, the batholith contact dips steeply west-southwest (Fig. 4). Continue south on Routes 27 and 16.

112.4 Kingfield; without crossing the bridge, proceed STRAIGHT ahead on Route 27.

115.7 Road cut on right of Seboomook Fm. M2 andalusite + biotite, partly sulfidic, largely altered.

119.7 Small road cut on right of Madrid (?) Fm. M2 garnet + biotite + chlorite.

122.4 Route 234 joins Route 27 from the left. Continue on Route 27 (all the way to Farmington). From here to 131.6 mi. the road passes through the New Vineyard Mountains. The rocks are mainly M1 Seboomook Fm. chlorite phyllites.

129.8 From here to Farmington, the road travels over M2 Garnet Zone phyllites, poorly exposed. West of here are M2 retrograded Staurolite Zone rocks (Fig. 1). A general southward increase in the width of the M2 Garnet Zone may be related to increasing  $P$  of M2 to the south, or to a decrease in the horizontal  $T$  gradient.

131.8 Turn LEFT (south) on Routes 27 and 4 through Farmington.

134.9 Turn LEFT (southeast) on Routes 27 and 2.

136.7 Farmington Motel.

**ROAD LOG FOR SUNDAY, OCTOBER 24, 1993**

We will depart at 8:00 am from the Farmington Motel. Today we will be seeing outcrops in the "old" Phillips, Rangeley, Oquossoc, Rumford, and Bryant Pond 15' quadrangles. Detailed geologic maps at this scale exist for these areas; Moench (1971), Moench and Hildreth (1976), and Guidotti (1963, 1965, 1977). Portions of the Rumford and Bryant Pond 15' quadrangles have been remapped at a 7.5' scale by Brady (1991). The manner in which these geologic maps fits into a larger geologic framework is shown in Osberg et al. (1985), and more recently in Moench and Pankiwskyj (1988). Stops 1-8 are shown on Figure 5.

**Mileage**

0.0 Proceed WEST on Route 2.

1.9 Bear RIGHT (north) on Routes 4 and 27 and continue through Farmington. We will stay on Route 4 until the town of Oquossoc.

9.3 Entering the Kingfield 15' quadrangle.

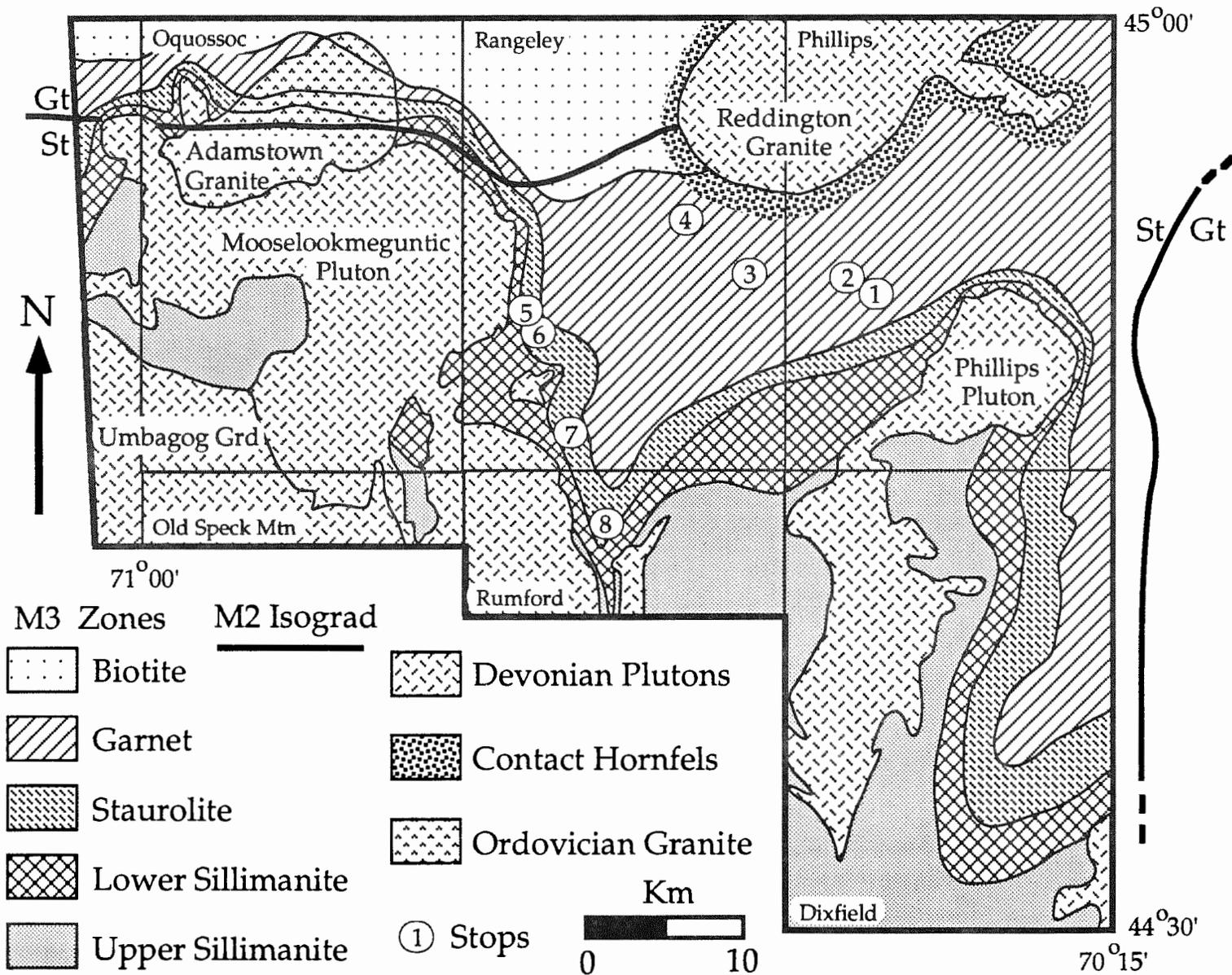


Figure 5. M2 staurolite (St) isograd, M3 zones, and field trip stops (1-8) in the Oquossoc, Rangeley, Phillips, and Dixfield quadrangles (Fig. 1).

## GUIDOTTI AND HOLDAWAY

- 10.8 Barren mountain to the north is Mt. Abraham; it is underlain by the Reddington Granite and its contact hornfels.
- 16.45 Spruce Mt. Rest Area; we are now in the Phillips 15' quadrangle and have entered the northern lobe of the Phillips pluton.
- 20.6 The road to the right goes into Phillips village where there are good outcrops of the Phillips pluton in the Sandy River. Stay on Route 4.
- 21.9 Phillips Middle School; New outcrops of the north contact of the Phillips pluton with the Carrabassett Fm. Continue north on Route 4.
- 24.2 Outcrops of Madrid Fm. on the left of Route 4.
- 24.6 Outcrops of the quartzose type of Smalls Falls Fm. continue to 24.9.
- 25.6 Carrabassett Fm.
- 25.9 Park carefully on the right taking care not to get stuck in soft sand.

**STOP 1. REMNANT M<sub>2</sub> ASSEMBLAGES IN CARRABASSETT FORMATION (30 MINUTES).** Watch out for cars. Here, the downhill (northerly) outcrops consist of fairly typical Madrid Fm. The uphill crops are in the Carrabassett Fm. Hence, the stratigraphic contact is present nearly in outcrop.

The Madrid Fm. consists mainly of thick bedded, fine-grained biotite granofels. Moderate amounts of calc-silicate interbeds are usually present. Here, they appear to have been pulled apart into boudins. Notice that much of the biotitic granofels has a slight purplish cast to it, presumably due to the biotite chemistry/color(?). To my knowledge, this commonly observed color for biotitic granofels associated with calc-silicates has not yet been explained.

Some granitic dikes (presumably from the Phillips pluton) cut the Madrid Fm. outcrops here. At the north end of the Madrid crops some of the surface weathering textures suggest remnants of fossils.

The Carrabassett Fm. consists of well bedded dark gray (graphitic), well foliated metapelite and biotite granofels. The bedding is typically on 4-8 cm scale and commonly it shows clear sedimentary grading (indicating tops to the south) as well as other sedimentary structures.

The outcrop of Carrabassett Fm. contains perfectly fresh staurolite that has persisted within the M<sub>3</sub> Garnet Zone. It contains staurolite + biotite + garnet + chlorite (possibly secondary here) ± andalusite, (+ ilmenite, pyrrhotite, and graphite) and so is a typical M<sub>2</sub> staurolite-grade assemblage for this part of western Maine. The staurolite crystals are euhedral and typically 1 cm in size and oriented in random directions with regard to the foliations in the rock. Moreover, they are larger in their dimensions than any crenulation cleavages in the rocks. Andalusite is only subhedral in shape and very difficult to see on the freshly broken surfaces. Some is visible at the south end of the crop about 2 m above the road bed. Can you find it? It occurs as anhedral to subhedral, pink megacrysts. On the upper portion of the outcrop andalusite is more clearly visible in some beds as 2-3 cm knobs that stand out in positive relief.

Hence, this outcrop appears to be an "island" of M<sub>2</sub> staurolite grade that has withstood the M<sub>3</sub> downgrading. Most likely this reflects inability of H<sub>2</sub>O entering some areas so that the downgrading was not effected. However, one other possibility is that the areas in which M<sub>2</sub> staurolite is preserved represent M<sub>3</sub> hot spots. This suggestion is based upon the geophysical study of Carnese (1983) which indicates that the sheet-like Mooselookmeguntic pluton extends beneath this whole area, reaching as far to the northeast as the Reddington Pluton. His cross sections suggest that it is not far below the surface such that local apophyses to higher levels may have produced M<sub>3</sub> temperatures still within the staurolite stability field. Obviously, resolution of such a suggestion will require a lot of detailed petrologic study.

- 26.35 Madrid Fm.
- 29.9 Cross the Sandy River and park on the right.

**STOP 2. VARIABLE PSEUDOMORPHING ON M<sub>2</sub> STAUROLITE AND ANDALUSITE (30 MINUTES) NO SMOKING** at this outcrop at the request of the land owner. The key thing to see at this outcrop of Carrabassett Fm. is the variable "downgrade" development (incipient to complete) of M<sub>3</sub> pseudomorphs of M<sub>2</sub> staurolite and andalusite megacrysts. This outcrop is another one of those "islands" within the M<sub>3</sub> Garnet

## GUIDOTTI AND HOLDAWAY

Zone in which M<sub>2</sub> staurolite and andalusite persists. Walk downstream along the north side of the river for about 50 yards to the extensive outcrops.

The uppermost river crops tend to have fresh staurolite with only slight rimming by muscovite. Some andalusite is replaced by coarse-grained muscovite. Note the random orientation of megacrysts. As one proceeds downstream for about 100 m, it is evident that the degree of replacement of staurolite by medium-grained muscovite and chlorite increases to the point where staurolite is gone, or remains only as rounded blebs in the core of a cuhedral pseudomorph. In most places this "downgrade" pseudomorphing involves a concentration of coarser laths of chlorite on the rims and somewhat finer-grained muscovite in the cores. Note that the pseudomorphs have the same spatial orientations with respect to S surfaces as do the fresh staurolite. People who are familiar with the work of Bell and his co-workers will recognize that the relationships between megacrysts and crenulation cleavages present in many outcrops we will see today do not agree with their suggestions.

At the lower outcrops there are also pseudomorphs of andalusite megacrysts up to 3 cm by 1 cm in size. Look across the river and you will see that some beds have extensive development of randomly orientated andalusite.

- 27.9 Madrid village and type locality of the Madrid Fm. Very large outcrops in the river and some on the highway to the north.
- 28.35 Into the Smalls Falls Fm. Abundant outcrops occur along the highway.
- 30.8 Turn to the LEFT into the Smalls Falls picnic area and drive to the back of the area (at 30.95 miles).

**STOP 3. M<sub>3</sub> METAMORPHISM OF SMALLS FALLS AND PERRY MTN. FMS. (1 HOUR).** This is the type locality of the Smalls Falls Fm. and a short distance to the north. on this stop we will see the Perry Mountain Fm., again at its type locality.

From a petrologic view point the Small Falls Fm. is an interesting unit because it contains 5-10 modal % of pyrrhoute and also is graphite-rich. As a consequence it shows the effects of sulfide-silicate mineral reactions to an extreme degree. The most straight forward aspect of these reactions is that the silicate bulk composition is moved to a very Mg-rich portion of composition space. Hence, at the appropriate grades one finds assemblages like andalusite or sillimanite + Mg cordierite + phlogopite. Fe-rich minerals like garnet or staurolite never occur in this unit. Moreover, the Ti phase is rutile instead of ilmenite as in the other Siluro-Devonian metapelites.

Walk to the outcrop at the edge of the plunge pool. On the smoothed surfaces you will find large chiastolite crystals - except that, as shown in Guidotti and Cheney (1975) they are now pseudomorphs composed mainly of margarite. As your eyes become more focused on the details of the textures present, you will see that other 2 cm pseudomorphic knots are present. They consist of aggregates of chlorite and phlogopite after cordierite. The original chiastolite and cordierite formed during M<sub>2</sub> but during M<sub>3</sub> they have been pseudomorphed.

Please don't hammer on the outcrops in the park area. The best opportunity for collecting margarite is at the road outcrops just uphill on the highway above the picnic area.

Return to the bridge and cross it. Just at the base of the bridge on the other side there are good chiastolite pseudomorphs which may be collectable. Walk up the steep trail along the river about 200 meters--to the point where the safety fence ends (**PLEASE** stay within the fence!). At the end of the fence it is safe to go down to the river outcrops again. There, in the bottom of an enlarged pothole there are some very well displayed pseudomorphs of cordierite. They consist of coarse-grained muscovite, Mg-rich chlorite, and phlogopite. Please do not hammer on them; take a picture instead!

Continue upstream along the river. Some 30-50 meters above the pseudomorphs there is a nice display of preserved sedimentary features in the quartzose beds of the Smalls Falls Fm. Continue upstream to the wooden bridge for a woods road, cross the river and return to Route 4. There, the bus will meet us in order to take us about 1/2 mile further up the highway. We will stop again on the left hand side of the highway, about 200 meters below the latter part of Stop 3.

Upon leaving the bus, very carefully cross to the other side of the road and form a group. Obviously the outcrops here are still Smalls Falls Fm. Upon walking up hill to the sharp bend in the road one abruptly encounters a new, very different unit, the Perry Mountain Fm. We are essentially at the type locality of this unit. Notice the well preserved bedding features; cross beds, graded beds (both giving tops to the south) etc. The main foliation is

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essentially parallel to bedding but a crenulation cleavage (S<sub>3</sub>) is also present and shows up in outcrop as a pronounced crinkling. In the context of the well preserved sedimentary features, keep in mind that these rocks have experienced two deformations and possibly three metamorphisms.

The groundmass of the pelitic beds is a nice "white schist" due to recrystallized muscovite. Garnet is partially replaced by coarse chlorite but minor "live garnet" has been observed. Hence, these rocks are shown on Figure 5 as M<sub>3</sub> Garnet Zone. The staurolite pseudomorphs are well developed and obvious - but can you find the pseudomorphs after andalusite? Look on the top of the southernmost bedding surface. It is clear that the original staurolite and andalusite are spatially, largely unrelated to any of the S surfaces and much greater in size than the spacing of any crenulation cleavages present.

As we continue to the north in the bus, we encounter outcrops of the Rangeley Fm. at about the town line of Sandy River Plantation.

- 31.1 At the entrance of the picnic area again. Turn LEFT on Route 4.
- 31.3 Woods road comes in from the left. This is where we will re-board the bus for the second part of Stop 3.
- 31.85 On a very tight, hairpin turn. This is the upper part of Stop 3. For later users of this guide driving in an auto, it will be possible to very carefully cross to the left hand side of the road and park on the wide shoulder.
- 32.1 Sandy River Township. Abundant outcrops of the Rangeley Fm. occur along the highway for the next several miles.
- 36.7

**STOP 4. EUHEDRAL PSEUDOMORPHS OF STAUROLITE IN RANGELEY FM. (20 MINUTES).** Here, there is a great display of euhedral staurolite pseudomorphs in a groundmass that is generally coarser grained than at Stop 3. Otherwise, the metamorphic textures, style, and history are very similar to that at Stop 3. This is a good outcrop for both collecting and photography.

Figure 5 shows this outcrop in the M<sub>3</sub> Garnet Zone, but garnets are hard to find due in part to the moderately sulfidic nature of the outcrops-thereby resulting in a somewhat Mg-enriched silicate bulk composition. Can anyone find "live garnet" or pseudomorphs after andalusite? About 1 km north of here I have found rocks containing euhedral pseudomorphs of both staurolite and andalusite. Hence, it seems likely that rocks at Stop 4 were at an M<sub>2</sub> grade that produced the compatibility of andalusite plus biotite.

- 37.1 Large outcrops of rusty weathering Rangeley Fm.
- 40.9 Large outcrops of the Rangeley Cgl.
- 41.25 Contact between Rangeley Fm. and the underlying Greenvale Cove Fm.
- 44.4 West side of Rangeley village. For the next 5 miles we will see many outcrops of the Quimby Fm. All of it is at low grades (M<sub>1</sub>?) and shows no evidence of having been affected by any of the higher grades that typify M<sub>2</sub> and M<sub>3</sub>.
- 49.5 Kamankeag Fm. slate member. These rocks are still at very low grades and a few miles to the north they contain Ordovician graptolites, Harwood and Berry (1967).
- 49.6 Contact with the volcanics member of the Ordovician Kamankeag Fm.
- 50.2 Metavolcanics on both sides of the road. This was Stop 1 of Guidotti (1970a). These outcrops belong to the Ordovician Kamankeag Fm. The rocks are typical greenstones (nearby pelites indicate Biotite Zone) containing actinolite + chlorite + epidote + albite and some calcite, quartz, sulfides, etc. The protolith is presumed to be basaltic flows but some pyroclastics are also present in other outcrops. Epidote-rich pods and clots occur in these otherwise massive to foliated volcanics. Possibly these represent volcanic bombs. In some cases, remnant plagioclase and pyroxene grains have been found in these rocks.
- 50.6 Oquossoc village; turn LEFT, (south) on Route 17.
- 55.9 Rangeley scenic overlook. Large outcrops of now metamorphosed (by M<sub>3</sub> at least) Quimby Fm. continuing to 56.7 miles.
- 57.6 Enter Rangeley Fm.
- 61.65 Heights of Land scenic overlook.

**STOP 5. HEIGHTS OF LAND SCENIC OVERLOOK (30 MINUTES).**

These outcrops are in the Rangeley Fm., and metamorphically in the "Transition Zone" as defined by Guidotti (1974). In particular, this means they are interpreted as having a five phase assemblage (sillimanite + staurolite +

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garnet + biotite + chlorite) which buffers the activity of H<sub>2</sub>O, i.e. we are in a narrow zone of a smeared out isogradic reaction. This is the locality of specimen Ra-a96-66 of Guidotti(1974). Sillimanite occurs only in thin section as fibrolite, and Mg-rich chlorite is present as sharply bounded laths that are mutually in contact with sillimanite as well as all other minerals present.

At this locality the Rangeley Fm. is mainly interbedded metapelites and biotite schist and granofels. Many sedimentary features are still preserved, including graded beds. As you inspect crops to the south you will see that various types of metaconglomerates are also present. The rusty weathering surfaces of these outcrops reflect the presence of up to 3 modal % of pyrrhotite in many beds. Randomly oriented, euhedral staurolite up to 1 cm in size are common in this outcrop. Irregular, sponge-like, 1-3 cm anhedral of pink andalusite are visible on some weathered surfaces of this outcrop. In some cases one can observe coarse-grained muscovite plates around these andalusite crystals.

The view to the southwest includes the Bemis Range and Elephant Mt. They are underlain by the Rangeley Fm. in a large inclusion (see Fig 5) within the Mooselookmeguntic pluton. Of interest is the fact that structurally the rocks in the inclusion define a tight syncline which is structurally coherent with the same syncline in the Rangeley Fm. at the east margin of the pluton. This implies that intrusion of the pluton did not structurally disrupt significantly the inclosing metasediments. To the west and northwest the broad lowland occupied by lakes is underlain by the Mooselookmeguntic pluton. On the west slopes of Elephant Mt. the grade of metamorphism rises to Upper Sillimanite Zone--i.e. the stability of staurolite is exceeded (Guidotti, 1970a).

Some granitic dikes are evident in these outcrops. Presumably they are derived from the underlying Mooselookmeguntic pluton. As noted in Guidotti (1970b), about 1.5 km to the north, the Union Water Power Co. drilled and cored down to about 900 ft. They allowed me to take a sample every 20 feet or so. At the surface the grade is Transition Zone (i.e. sillimanite + chlorite). With increased depth the grade rises and at about north 850 feet the hole hits adamellite (which field mapping by Moench could have predicted). About 150 feet from the granite there is an abrupt rise in muscovite and disappearance of sillimanite, suggesting K-metasomatism such as described by Green (1963) near the contacts of similar adamellites in the Errol quad.

62.55

#### STOP 6. STAUROLITE ZONE "SWIFT RIVER PSEDOMORPHS" (20 MINUTES).

This is a newly exposed outcrop of Rangeley Fm., which due to surface weathering, shows textural features extraordinarily well. **PLEASE DO NOT HAMMER ON THE FLAT, WEATHERED SURFACES.** You can probably find float boulders here or across the road that show the textures quite well.

The main textural feature to observe is the partial pseudomorphing of staurolite by muscovite and minor chlorite. Note that the remnant staurolite forms aggregates of small "new" euhedra oriented in the same direction as the original 1-2 cm megacrysts. This type of pseudomorph is typical of the "hinge line" along which M<sub>2</sub> and M<sub>3</sub> both attained nearly the same *T* and *P* conditions. I have called these "Swift River" pseudomorphs because they were first observed at outcrops on that river (at Stop 7). At higher grades, (transition and lower sillimanite zones), staurolite is pseudomorphed by aggregates of coarse muscovite forming rims around anhedral remnant staurolite grains (Guidotti, 1968). Eventually in the Upper Sillimanite Zone these aggregates re-crystallize to form large, 1-2 cm spangles of muscovite (Stop 9).

At the north end of this out crop you can see good examples of large, 1-3 cm, skeletal (i.e. highly poikiloblastic) andalusite crystals.

- 64.0 Arkosic member of the Rangeley Fm. containing calc-silicate "footballs" that display abundant evidence of diffusion.
- 64.5 A member of the Greenvale Cove Fm. consisting of interbedded calc-silicate and biotite granofels.
- 64.9 Pelite underlying the Greenvale Cove Fm. in the core of the Brimstone Mountain Anticline. At this outcrop there are nearly complete pseudomorphs of staurolite by coarse-grained muscovite, but they are hard to see unless the sunlight is favorable.
- 65.4 Outlier of the Mooselookmeguntic Pluton.
- 66.1 Metapelite in the core of the Brimstone Mountain Anticline.
- 66.4 Calc silicate member of Greenvale Cove Fm. again; then back into Rangeley Fm.
- 67.8 Swift River; park on the left a short distance below the bridge.

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**STOP 7. EVIDENCE FOR BUFFERING OF  $a(\text{H}_2\text{O})$  (30 MINUTES).**

Just across the bridge over the Swift River. Plenty of parking on the left, but watch embankments on entering parking area. Follow trail down stream on south side of River to where a clearing with a camp is encountered. Go to the far side the clearing, and then down to the Swift River. Specimens from this segment of the river were used as the starting point for the tourmaline paper of Henry and Guidotti (1985). The metamorphic grade here is M<sub>3</sub> upper Staurolite Zone.

Here you will see the "type locality" of the "Swift River" pseudomorphs. Many such pseudomorphs are present in these outcrops. However, the main interest now is the easily visible pyrrhotite-rich bands and the sulfide-silicate reactions that have occurred in association with the variable amounts of sulfide in different specimens. These reactions were the focal point of the study by Bussa (1973) and then Henry (1980). As you can see from the weathering surfaces, the amount of pyrrhotite varies from scattered amounts in a given bedlet to thin bedlets of almost pure pyrrhotite. Many of the specimens contain the assemblage staurolite + garnet + biotite + chlorite which is a limiting assemblage in terms of its phase rule variance. In any given outcrop Henry found a very systematic relationship between the modal amount of pyrrhotite and the Mg/Fe ratios of the silicate minerals and also their Mn-contents. In the case of the Mn-contents the systematic pattern includes the Mn content of ilmenite (and even the apatite!). In essence, due to the variable extent that the sulfidation reactions had proceeded, the activity of water was lowered such that the dehydration equilibrium that existed among the Fe-Mg silicates was driven to the right, i.e. more Mg-rich compositions for the minerals.

Two general conclusions which emerge from Henry's work are: (1) The activity of H<sub>2</sub>O varied on a very local scale that was controlled internally (buffered) by the mineral assemblage. (2) The fact that the mineral compositions are so delicately balanced with each other implies that a fairly close approximation to equilibrium was achieved.

Some calc-silicate nodules are well exposed along these river outcrops. They show clear evidence for small scale metasomatic transfer. No work on the diffusion that has occurred in these nodules has ever been done. Some small granitic dikes also occur in these outcrops.

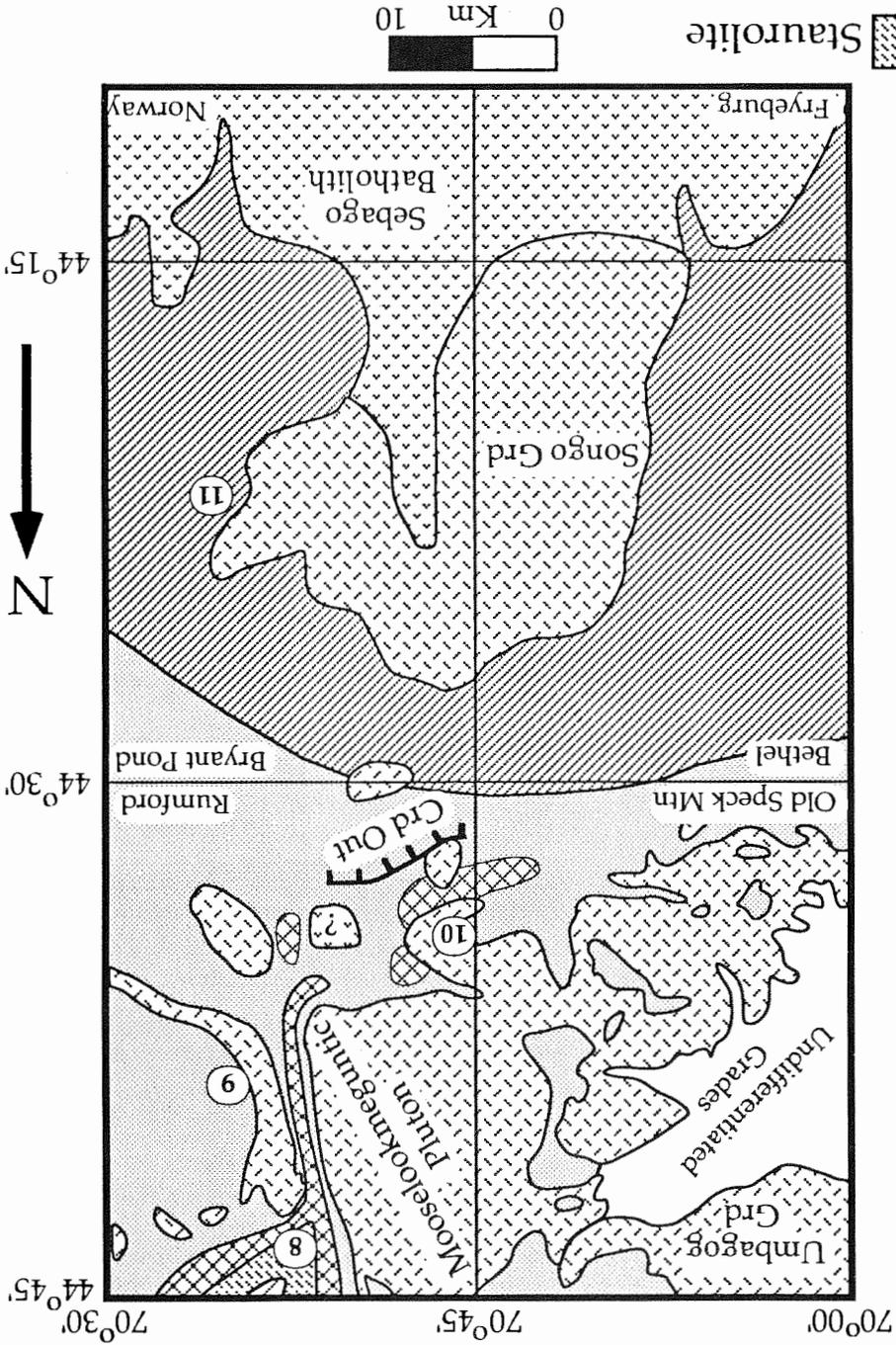
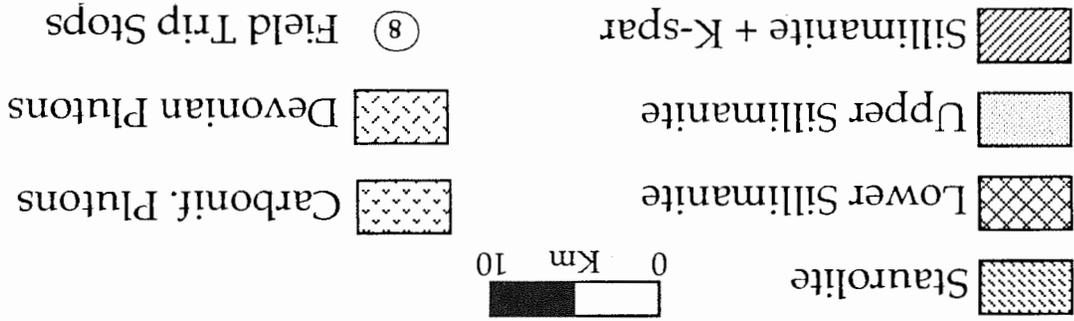
## 73.2 Coos Canyon picnic area

**STOP 8. BYRON MAINE, COOS CANYON AND BELOW ROUTE 17 BRIDGE (1 HOUR).** The rocks here, (see Figs. 5, 6) are part of the Perry Mountain Fm., (Moench, 1971). Only Devonian metamorphism can be discerned here and these rocks are now at M<sub>3</sub> staurolite grade. The diagnostic assemblage is staurolite + biotite + chlorite + garnet + ilmenite + pyrrhotite. In the northern part of the exposure (down on the river edge) staurolite euhedra up to 1 cm in length are found. Their spatial relationships with the various S surfaces present are as seen in earlier outcrops.

Textural evidence for an earlier, higher grade metamorphism (M<sub>2</sub>) is seen on bedding-slab surfaces. It consists of coarse muscovite, "Turkey Track" pseudomorphs after andalusite. These pseudomorphs are up to 10 X 2 cm in size and in a few cases still retain some fresh andalusite. (See also the outcrops at the north base of the bridge on the other side of the river.) The assemblage formed in this earlier event involved andalusite + staurolite + biotite + garnet. It is interesting that here (outcrops down Route 17), and elsewhere in the Rangeley area, totally fresh andalusite seems best preserved in rocks rich in sulfides and graphite. One can speculate that the fluid in such rocks had a lower % of H<sub>2</sub>O and thus reactions removing metastable andalusite (especially by forming hydrous minerals like muscovite) may have been inhibited. Some workers have appealed to dissolved Fe<sub>2</sub>O<sub>3</sub> in the andalusite acting as a stabilizer. The other common site for "fresh" relic andalusite is in quartz veins. This may result from relative inaccessibility to solutions bearing K<sub>2</sub>O.

Note how the pseudomorphs are unoriented within bedding planes. In addition, they are undeformed plus the muscovite in them is unoriented. Moreover, staurolite euhedra and tablets of coarse biotite are also largely unoriented. Although evident mainly from thin section study, these megascopic observations provide some indication that both M<sub>2</sub> and M<sub>3</sub> were largely static events. Some isoclinal folds are evident on the eastern wall of the gorge and a few scattered aplite dikes are also present in these outcrops. Although these rocks have been affected by two fairly high-grade metamorphism, it is remarkable how well graded beds and cross-beds are preserved. Possibly this reflects the static nature of the two metamorphism. Some chemical data is available for the minerals at this outcrop - see specimen 1-8/7/63 of Evans and Guidotti (1966), Conatore (1974), and Guidotti et al. (1975).

Figure 6. M3 staurolite, Lower Sillimanite, and Upper Sillimanite Zones, Carboniferous Sillimanite K Feldspar Zone, and field trip stops (8-11) in the Old Speck Mtn., Rumford, Bethel, and Bryant Pond quadrangles (Fig. 1)



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Now walk southward along Route 17 to inspect the rusty-weathering outcrops just to the south. These outcrops are unusual for the Perry Mountain formation in that usually this unit is gray-weathering. Because of the abundant pyrrhotite in these outcrops some sulfide silicate reactions come into play. For example, garnet is absent, staurolite becomes much less abundant or is absent, and in some of the latter the titanium phase is rutile rather than ilmenite.

Now continue walking to the south to inspect the outcrops about 200 m downstream from the Route 17 bridge. Inspection of bedding surfaces on the northernmost outcrops shows good development of andalusite pseudomorphs up to 4 cm long. On the large outcrop to the south (75 m) there are several nice folds, one especially so. These folds have a well developed, but annealed (polygonized) axial plane, crenulation cleavage (1-2 mm scale). In some cases (just north of the largest fold) this cleavage has concentrations of biotite flakes that formed by recrystallization of the mica rich portions of the original crenulation cleavage. As in all previous outcrops, the size of staurolite megacrysts is much greater than the scale of the crenulation cleavage, and typically they truncate this cleavage at variable angles.

Also present at this outcrop are the usual good preservation of sedimentary features, some granitic dikes, ptigmatic folding of quartz veins by the axial plane crenulation cleavage, and some irregular, cross-cutting, discontinuous veinlets largely composed of staurolite crystals which are very similar in appearance to staurolite in the groundmass. The origin of these veinlets is still under study. Any comments or suggestions would be welcome.

- 73.4 Road on the right to the lower outcrops of Stop 8.  
 77.5 Village store and town offices in Roxbury.  
 79.75

**STOP 9. SWIFT RIVER OUTCROPS, 2.25 MILES SOUTH OF ROXBURY VILLAGE (20 MINUTES).** The rocks here (Fig. 6) are coarse, swirled, migmatitic gneisses. Bedding is difficult to find. Coarse spangles of muscovite are prominent and in some cases have shapes suggestive of staurolite crosses. Many spangles contain euhedral garnets similar to those seen in lower grade rocks (e.g. Stop 10) that retain clear muscovite pseudomorphs after staurolite. Pegmatite veins and irregular masses are prominent throughout this outcrop.

These rocks are exposed over a wide area and were originally in the Upper Sillimanite Zone or possibly even Sillimanite + K Feldspar Zone. However, in many outcrops throughout this area the sillimanite is now largely resorbed (not sericitized!) and commonly is seen in thin section as inclusions in quartz and feldspar or as discontinuous aggregates that don't cross grain boundaries. Hence, the present assemblage is biotite + garnet. Minor chlorite seems to be present only as a very late alteration phase. A few parts of the outcrop on the river and associated loose blocks do retain coarse sillimanite visible to the unaided eye! Can you find it?

In the new outcrop now exposed on the other side of the highway one can find better examples suggestive of the muscovite spangles being formed from pseudomorphs after staurolite, (Southernmost, flat outcrop and by the benchmark). The "muscovite spangles" are more obviously aggregates of a large number of coarse flakes. Moreover, in these outcrops sillimanite is more readily found as fibrolite associated with biotite. In the northernmost part of this outcrop, in addition to abundant tourmaline, there are abundant clots of sillimanite and even some sprays of prisms up to nearly 1 cm in length. Even in this case where sillimanite is readily visible in thin hand specimen, the rounded aspect of the clots suggests some resorption.

Inasmuch as the sillimanite is now commonly resorbed in these rocks but garnet and biotite are fresh, it would appear that these rocks have been affected by a later, medium grade metamorphism. Moreover, because hornblende in plutonic rocks this far north shows disturbed spectra (Lux and Guidotti, 1985), we would suggest that this later metamorphism reflect the Carboniferous heating event associated with intrusion to the south of the Sebago Granite. Indeed, crops a little south of here show some signs of regeneration of sillimanite.

The rocks at this outcrop have been mapped by Moench and Hildreth (1976) as undivided Siluro-Devonian. Rocks here are superb examples of swirled up "soup" in which bedding seems to have been obliterated. This appearance gives the impression that they may have been partially molten at some point and even undergone some plastic flowage. Their appearance, areal distribution, and relatively abrupt transition into structurally coherent M<sub>2</sub> metamorphic rocks (e.g. at Stop 8) at least raises the question that the heat for M<sub>2</sub> (or M<sub>3</sub> ?) metamorphism this far south may in part have been associated with a core complex/extensional environment. Further north in the Rangeley area it seems clear that the heat for both M<sub>2</sub> and M<sub>3</sub> was supplied by plutons.

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- 81.25 Leave Route 17, turn RIGHT,(west) on road to Route 120.  
 81.55 Intersection with Route 120. turn to the RIGHT, (west) on Route 120.  
 84.3 Roxbury Notch. Immediately west of here we enter the broad lowland underlain by the Mooselookmeguntic Pluton. Here, we are crossing the south extending "finger" of Lower Sillimanite Zone rocks shown on Figure 6. This "finger" is restricted to the tops of ridges, thereby emphasizing the flat aspect of the isogradic surfaces.  
 89.0 Granodioritic phase of the Mooselookmeguntic Pluton.  
 90.7 Stop sign; turn to the RIGHT, staying on Route 120.  
 91.55 Stop sign; This is the end of Route 120 in the village of Andover. Turn LEFT, (south) on Route 5.  
 96.75 Two mica granite of the Mooselookmeguntic Pluton. From here, the view to the east shows Whitecap Mountain clearly. It is a mixture of (Carboniferous?, Guidotti et al. 1986) pegmatite and leucogranites.  
 98.8

**STOP 10. CARBONIFEROUS, LOWER SILLIMANITE ZONE (20 MINUTES).**

This outcrop involves well-bedded Perry Mountain formation. Even though these rocks have been exposed to as many as three high-grade events, graded beds are still present and show that the strata are overturned at this locality. These rocks were included in the studies of Evans and Guidotti(1966), Cheney(1975), and Cheney and Guidotti (1973,1979).

The grade here is Lower Sillimanite Zone and the typical assemblage involves sillimanite + staurolite + biotite + garnet + ilmenite + pyrrhotite. Staurolite is mostly replaced by coarse-grained, unoriented muscovite to form pseudomorphs that appear as the conspicuous white "eyes" seen in the pelitic portion of the graded beds. Usually, the staurolite is seen only in thin section but occasionally it is visible in hand specimen. In some places the muscovite "eyes" approximate the shape of staurolite and commonly contain euhedral garnet inclusions. Sillimanite is also present in the muscovite of the pseudomorphs and it typically crosses grain boundaries. The rough weathering surface of the pelitic beds is largely due to very abundant clots of fibrolitic sillimanite.

A crenulation cleavage at high angles to bedding is evident in parts of this outcrop. In thin section it is seen that this cleavage has been annealed to form polygonal arcs and has no effect on the muscovite pseudomorphs after staurolite. Moreover, the scale of the cleavage is much smaller than the dimensions of the staurolite which is now pseudomorphed.

These Lower Sillimanite Zone rocks trace directly up grade to the Upper Sillimanite Zone and then the Sillimanite + K Feldspar Zone that is associated with the Sebago Granite. Hence, it is believed that the last equilibration of the rocks at this Stop occurred during the Carboniferous heating event (DeYoreo et al., 1989).

- 99.45 Outcrops of Smalls Falls Fm. and pegmatite.  
 102.35 Intersect Route 2; turn LEFT (east)  
 102.95 Turn RIGHT (south) on Route 232. Take care; in places Route 232 is not well marked. After crossing the Androscoggin River, bear LEFT. Then at the next fork, bear RIGHT.  
 112.35 Junction with Route 26. Turn LEFT (south). The outcrops in the vicinity of this junction belong to the Songo Granodiorite, a Devonian (382 Ma) pluton that has had its hornblende Ar ages reset due to the intrusion of the Sebago batholith.  
 116.2

**STOP 11. MIGMATITES OF THE SILLIMANITE + K FELDSPAR ZONE (20 MINUTES).**

These commonly migmatitic rocks are in the extensive Sillimanite + K Feldspar Zone that was developed in response to intrusion of the Sebago Granite. This extensive area of high grade rocks appears to have formed above the low (northerly) dipping sheet of granite that constitutes the Sebago pluton, (DeYoreo et al 1989). Large numbers of pegmatitic bodies are present throughout this area and are "presumed" to be derived from the cooling granite below. Sillimanite is extremely abundant in metapelite compositions. K feldspar is also abundant, both in the groundmass and as coarse megacrysts. The latter commonly contain quartz inclusions that simulate bipyramids such as one would see in volcanic rocks. Muscovite is still stable so that the full assemblage is sillimanite + garnet + biotite + K feldspar (orthoclase) + muscovite + plagioclase (An<sub>30</sub>) + ilmenite + pyrrhotite.

Note that there most of the textural evidence for polydeformation and polymetamorphism so prominent in previous stops is now absent. Most likely this is because the degree of re-crystallization attained in these rocks has erased the evidence for the earlier events.

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From here continue southward on Route 26 to the town of Grey. There you will intersect the Maine Turnpike. From this point on the drive to Boston will be on freeways. The total driving time from Stop 11 to Boston should be under 3 hours.

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Chapter M

# Migmatites of Southern New England: Melting, Metamorphism and Tectonics

By Eileen L. McLellan, Robert J. Tracy, Thomas R.  
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# MIGMATITES OF SOUTHERN NEW ENGLAND: MELTING, METAMORPHISM AND TECTONICS

by

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## INTRODUCTION

The reputation of southern New England (Massachusetts and Connecticut) as a classic locale for high-grade metamorphism and migmatite formation has long been well established, but only in recent years have petrologists begun to appreciate the richness of the details of the Paleozoic tectonic framework within which these events took place. The traditional view of a roughly 100 million year periodicity of discrete deformational and metamorphic episodes (a Taconic orogeny at ca. 475 Ma, an Acadian orogeny at ca. 370 Ma and an Alleghanian orogeny at ca. 290 Ma) has given way to an appreciation based on modern geochronologic techniques that northern Appalachian Paleozoic orogenic events form more of a continuum on a regional scale. Similarly, the older view that Taconic orogenic features were restricted to western New England, Acadian ones to central and northern New England and Alleghanian ones to southeastern New England has undergone substantial revision. Increasingly detailed petrologic and geochronologic studies in both central and southern New England have emphasized that a better understanding of *styles* of metamorphism and deformation allows for both geographic and chronologic subdivision of the classic orogenies, and sheds much new light on the Paleozoic plate tectonic processes that formed southern and central New England. We emphasize to the reader that this new way of looking at New England geology is in its very early stages, and much mapping or remapping, petrology, structural geology and geochronology remain to be done.

Our purpose in organizing this trip is two-fold. First, we will visit some of the most classic migmatite localities in southern New England in order to see their features "on the hoof" and to discuss the petrology of high-grade metamorphism and migmatite formation. The characteristics of both the metamorphism and resultant anatexis must be understood in light of both P-T paths and durations of thermal events. Second, and equally important, we will use our discussions of the petrology of migmatite formation to stress the contrasts in style of high-grade metamorphism from place to place. The new models of southern New England metamorphism, and especially subdivision of the Acadian into eastern and western realms, are based on these contrasts in metamorphic styles and the plate tectonic reasons for them. The organization of this trip guide involves three parts: first we will introduce the reader to essential elements of the latest tectonometamorphic reconstructions of southern New England (with appropriate background references); second, we will review the important characteristics of high-grade metamorphism and migmatite formation in each of the areas visited by the trip; and finally, the bulk of the guide will be a detailed trip log with stop descriptions.

## A TECTONOMETAMORPHIC MODEL

In a recent attempt at a preliminary tectonometamorphic synthesis, Armstrong et al. (1992) reviewed both the styles and timing of Paleozoic metamorphism in western, central and southern New England. Broadly, metamorphism in this region is the result of three different orogenic events: the Taconian (Middle to Late Ordovician), the Acadian (Late Silurian to early Devonian), and the Alleghanian (Late Carboniferous to Early Permian). In addition, there is Ordovician high-grade metamorphism and migmatite formation in the Nashoba Block and Massabesic Terrane of eastern New England which forms the western edge of the so-called Avalonian realm, and which is not typically considered to be part of the Taconian orogeny *sensu stricto* (Aleinikoff et al., 1979). In this discussion, we will subdivide the Acadian into eastern and western realms, roughly divided by the Connecticut River; justification for subdivision will be provided below. Although it has long been recognized that these metamorphic belts trend generally north-south and that the younger belts (excepting the problematic Nashoba belt) step progressively to the east, they are not discrete by any means. There are substantial geographic overlaps between the Acadian and Taconian belts in western New England and between the Alleghanian and Acadian in east-central and southeastern New England. Figure 1A shows the major tectonic elements of southern and central New England

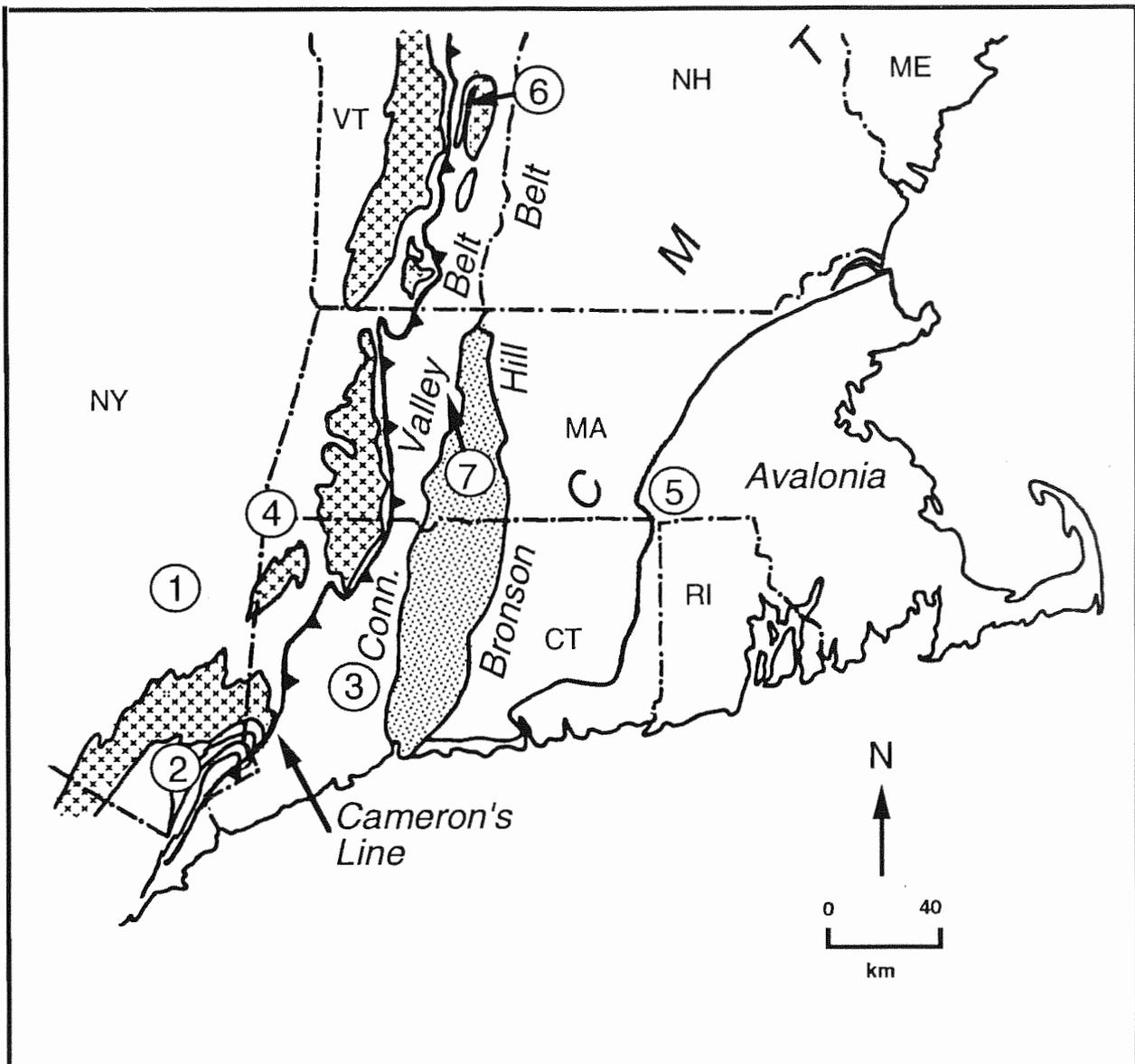


Figure 1. (A) Generalized tectonic map of southern and central New England, showing locations of several key tectonic and petrologic features discussed in the text. Grenville basement massifs are shown with cross symbol, and Mesozoic rift basin with dots. Relevant numbered references are: (1) Taconian foreland belt, NY; (3) Waterbury Dome, CT; (4) Taconic allochthons of foreland belt; (5) Honey Hill-Lake Char-Bloody Bluff fault system, western boundary of the Nashoba Block and Avalonia; (6) Chester Dome, Vermont; (7) Hatfield plutons and surroundings - EA-type rocks west of the Mesozoic basin; CMT - Central Maine Terrane. (Figure from Armstrong et al., 1992)

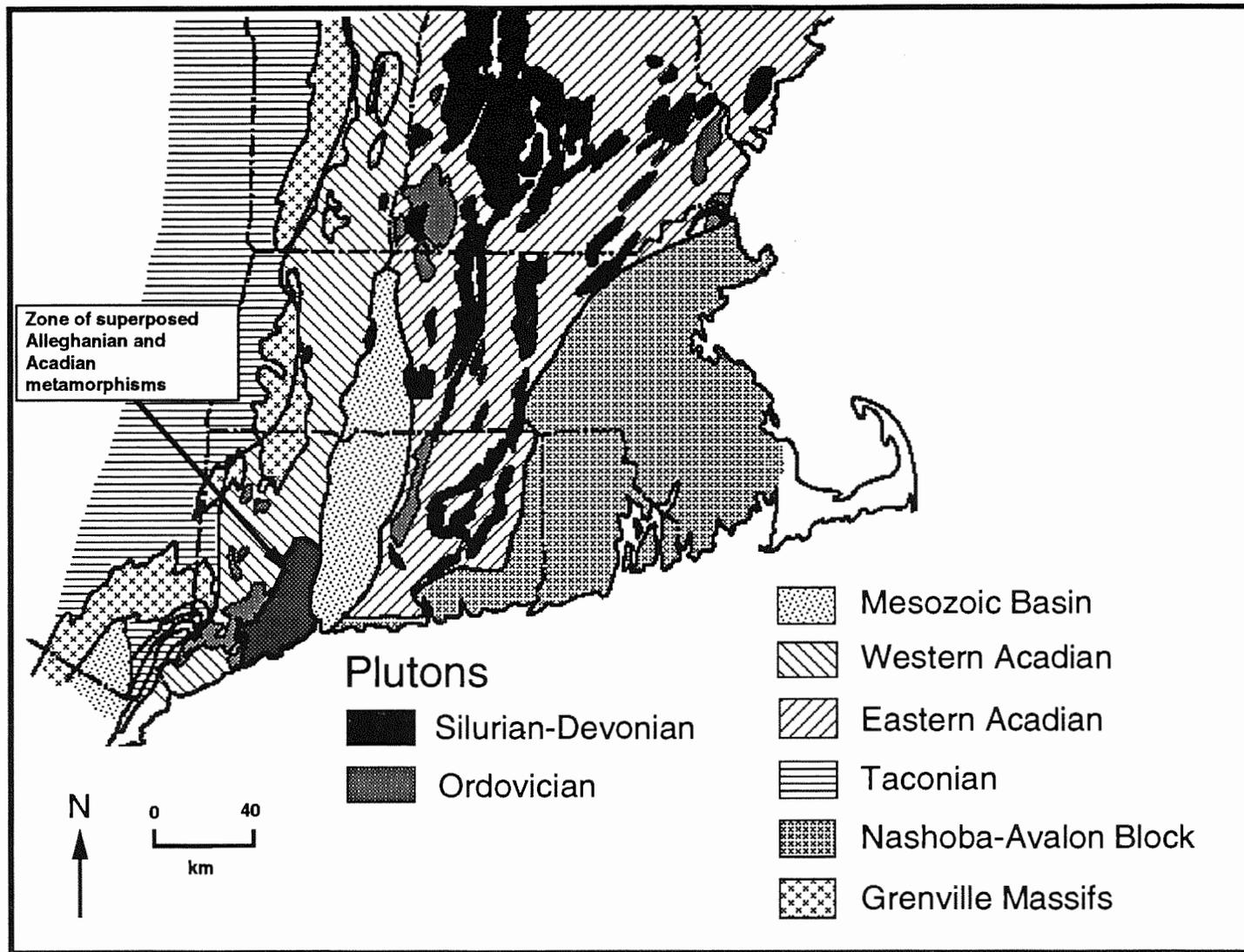


Figure 1. (B) Generalized geologic/plutonic/metamorphic-age map of southern and central New England showing the geographic distribution of Taconian, Eastern Acadian, Western Acadian and Alleghanian metamorphisms, as well as major plutons of Ordovician and Silurian-Devonian age and the Grenville massifs. Note that metamorphic areas are indicated where each metamorphism is the latest and highest-grade event, or in any case has produced the dominant metamorphic assemblage in the case of overlaps. (Figure from Armstrong et al., 1992)

referred to in this discussion, and Figure 1B denotes the parts of the region in which each of the Paleozoic metamorphisms was either the latest, the highest-grade or the most pervasive event in that area.

### **Taconian Metamorphism**

This trip will not visit any migmatitic localities of Taconian age, all of which are in extreme southwestern New England (Connecticut and adjacent New York), but a very brief discussion of Taconian metamorphism is necessary background for an understanding of the Western Acadian. The Taconian belt comprises greenschist to upper amphibolite facies rocks in western New England and New York (Drake et al., 1989). Taconian metamorphism, particularly that of high grade, may originally have extended farther east than the boundary shown in Figure 1B, but relatively high-grade Acadian metamorphism has overprinted and largely obliterated any evidence of Taconian metamorphism in Ordovician and older rocks east of a line roughly coinciding with the leading edge of the Taconian accretionary wedge (Cameron's Line and its northern equivalents). Metamorphic grade in the Taconian belt progressively increases from west to east (Robinson, 1983; Sutter, Ratcliffe and Mukasa, 1985), from greenschist or subgreenschist grade in New York and typically culminating eastward in kyanite- or sillimanite-grade rocks at the border with Acadian metamorphism in western Connecticut and Massachusetts. Only a small belt of presumed Taconian sillimanite+K-feldspar grade with associated migmatites is present in extreme southwestern Connecticut and adjacent New York (Robinson, 1983).

Stanley and Ratcliffe (1985) provided a thorough discussion of the North American margin before, during and after the Taconian orogeny. They postulated isograd geometry consistent with interpretation of metamorphism as a result of crustal thickening due to westward imbrication of the destructed ancient North American passive margin (including Taconian fore-arc and accretionary wedge regimes) onto the Middle Proterozoic Grenville and post-Grenville basement and associated parautochthonous and allochthonous rift clastics, post-rift slope-rise and carbonate platform sediments. Timing of the Taconian metamorphism remains somewhat unclear, and substantial diachroneity from north to south is likely. Estimates of time of attainment of peak metamorphic conditions range from >470 to about 445 Ma. Some of this range is undoubtedly real, and some is probably an artifact of geochronologic techniques, but there is no doubt that Taconian sedimentation was occurring as late as 460-455 Ma (see references in Armstrong et al., 1992).

Taconian metamorphism is typically of Barrovian style, with metamorphic field gradients or metamorphic arrays that lie roughly along the kyanite-sillimanite boundary in the P-T plane (Armstrong et al., 1992). The metamorphism appears to have been almost entirely the result of thermal relaxation of the imbricated wedge, with little or no contribution of magmatic heat. The plutonic map of the Appalachians of Sinha (1988) shows a few middle to late Ordovician granitoid plutons in western New England, but the ones that have been dated are post-metamorphic (i.e., post-440 Ma). The Taconian P-T path appears to be a "clockwise" one.

### **Eastern Acadian Metamorphism**

The metamorphic belt referred to as the Eastern Acadian (EA) comprises the Bronson Hill Belt and the Central Maine Terrane (CMT) (formerly known variously as Merrimack synclinorium or Kearsarge-Central Maine Terrane, among others) (Fig. 1). All of the first day's stops will be at outcrops in this terrane. The arbitrary western limit of this belt is the Mesozoic basin in southern New England and the VT-NH boundary (Connecticut River) in northern New England, although there are probably several thrust-bounded "excursions" of EA-style rocks west of the boundary, notably in east-central Vermont, in western Massachusetts (Hatfield-Whateley area) and southern Connecticut (near Bridgeport) (Lanzarotti, 1992, 1993). In the east, the EA belt is bounded by the Nashoba Block and the Avalon Terrane. The nature of this boundary and the time of accretion are uncertain at this time.

Tectonically, the EA belt can be broken down into: (1) the Bronson Hill Belt, the Taconian island arc - magmatic arc complex, possibly built up on Avalonian-age basement (Zartman and Leo, 1985; Tucker and Robinson, 1990, and (2) the CMT, an extensive back-arc sedimentary basin which was probably active from the Late Ordovician (*before* the end of the Taconian collisional phase) to the late Silurian (Moench and Aleinikoff, 1991).

In contrast to other terranes in southern and central New England (Figure 1B), the EA belt is characterized by extensive plutonism of both Ordovician and Silurian-Devonian age (Sinha, 1988) and the metamorphic style reflects the close connection between plutonism and metamorphism. Sillimanite + K-feldspar and higher-grade metapelite

assemblages and two-pyroxene metabasite assemblages are relatively widespread in the EA belt, and a narrow zone of true granulite-grade rocks, the highest grade of all metamorphic rocks of New England, occurs in the central part of this belt from northeastern Connecticut to south-central New Hampshire. A number of our first-day stops are in rocks that most petrologists would consider to be granulite-grade. For details of metamorphic grade, the reader is referred to metamorphic maps of Robinson (1983) and Zen et al. (1983).

Petrologic evidence accumulated over the years in the EA belt in central Massachusetts has increasingly indicated a so-called "anticlockwise" P-T path (see Tracy and Robinson, 1980 and Schumacher et al., 1989) in which early low-P, high-T Buchan-style metamorphism (essentially regional-scale contact metamorphism - see Pattison and Tracy, 1991) was succeeded by a later pressure increase or compression at near-peak T and during initial stages of cooling. This evidence includes: (1) large polycrystalline sillimanite aggregates that are pseudomorphic after andalusite (Rosenfeld, 1969); (2) abundant cordierite-bearing assemblages in which cordierite may show late breakdown to garnet-sillimanite-quartz symplectites (Tracy and Dietsch, 1982); (3) non-decrepitated and texturally early aqueous fluid inclusions (including in Kspar porphyroblasts) that yield low-P entrapment isochores, and late CO<sub>2</sub>-rich inclusions with higher-P isochores (Winslow et al., in press). Outcrop relationships suggest melting occurred early in the tectonic/metamorphic sequence, probably before the development of regional fold and thrust nappes as described by Thompson et al. (1968). All petrologic evidence points to regional-scale "contact" metamorphism thermally driven by heat input from voluminous plutons being emplaced in, or passing through, relatively shallow crustal levels. Vernon et al. (1989) have recently described a remarkably similar scenario in the Arunta Block, central Australia. Thompson and England (1984) and Kerrick (1991) emphasized the role of magmatic heating in production of Buchan-type metamorphism.

P-T data for the EA belt can be separated into two groups: for the higher-grade CMT in Massachusetts, Connecticut and New Hampshire, and for the somewhat lower-grade rocks in the Bronson Hill belt. Recorded peak conditions for the near-granulite or granulite CMT rocks are about 650-700°C at 3.5 - 4.5 kbar in New Hampshire (Day and Chamberlain, 1989) and 700-750°C at 5-6 kbar in Massachusetts and Connecticut (Robinson et al., 1989). The Bronson Hill belt shows lower T for similar pressures (550 - 650°C), except for possible relict high-T rocks on Fall Mountain in New Hampshire (Spear et al., 1990).

Recent geochronologic work has clarified the age relations of the plutonism and metamorphism. Most of the Silurian-Devonian plutons shown in Figure 1B that have been dated by U-Pb methods yield ages of 420-405 Ma, although there are several post-400 Ma ages as young as 360. Particularly important constraints are a 413 (5) Ma age on Kinsman quartz monzonite (Barreiro and Aleinikoff, 1985) and a 408 (6) Ma zircon age from a post-metamorphic quartz vein in southern New Hampshire (Zeitler et al., 1990). The first is important because of large volumes of Kinsman-type plutons in the EA belt, and the second because it puts a lower age limit on EA metamorphism.

Very recent work has revealed the possibility that at least some of the metamorphic features seen in the EA belt may actually be of Alleghanian age (Gromet and Robinson, 1990; Getty and Gromet, 1992). For example, lower Paleozoic cover rocks in the Bronson Hill gneiss domes show geochronologic evidence of both Acadian and Alleghanian thermal disturbances, but nearby basement gneisses only show evidence for Avalonian (ca. 650 Ma) and Alleghanian (ca. 275 Ma) disturbances. This raises the possibility of post-Acadian (possibly early Alleghanian) tectonic juxtaposition of cover and basement as far north as north-central Massachusetts and southern New Hampshire. See the Alleghanian section below for further details.

### Western Acadian Metamorphism

The Western Acadian (WA) metamorphic belt described here comprises much of western Connecticut and Massachusetts (west of the Mesozoic basins) and at least the eastern half of Vermont (Figure 1B). Its present eastern boundary may well represent a late to post-Acadian tectonic contact between EA and WA belts in which WA rocks may actually structurally overlie EA rocks (Hatch, 1988). Such a structural relationship would help to explain anomalous cooling ages reported from western New Hampshire and eastern Vermont (Harrison et al., 1989). The mapped western boundary is the limit of observable Acadian metamorphic effects that overprint the older Taconian belt. The overprinting relationship is quite subtle in many places in Massachusetts and Connecticut because of the

similarity of metamorphic grade in both Taconian and Acadian events along the line of overlap. Acadian thermal effects appear to drop off very rapidly to the west, and low to medium grade Taconian rocks in eastern New York show no effects of a later thermal disturbance, even though Acadian folding occurs well into New York.

The WA belt appears to be a zone of post-400 Ma tectonic and metamorphic reworking of older Taconian material (the fore-arc realm and telescoped accretionary wedge complex) as well as material tectonically emplaced from farther east as a consequence of pre-400 Ma orogenic activity in the EA belt, and perhaps synchronous with nappe-stage deformation in the EA belt (see below). One unusual feature of the WA belt is the relatively small variation in metamorphic grade from southern Connecticut to central Vermont. With the exception of an anomalous narrow low-grade strip along the eastern border of Vermont, the lowest grade observed in metapelites is garnet, with a substantial garnet zone in northern Massachusetts and southern Vermont (Robinson, 1983) but none observed farther south. Throughout most of the WA belt in western Connecticut (where some of our Saturday stops are located), western Massachusetts and southern Vermont, metapelite grade is kyanite-staurolite with sporadic occurrences of fibrolitic sillimanite in addition in western Connecticut and Massachusetts. Metapelites with sillimanite but not kyanite are very unusual. Incipient migmatization occurs in some of the kyanite-bearing metapelites, especially around Waterbury in western Connecticut where we will examine it, but migmatitic outcrops are in general far less extensively developed than in the EA belt of central Massachusetts or the Alleghanian belt in southeastern coastal Connecticut, for reasons we will discuss below.

Until recently, there was little quantitative information on the physical conditions of metamorphism in the western belt. Recent studies in metapelites in western Connecticut (Dietsch, 1988; Hames et al., 1989, 1991; Miller, 1990; Miller and Tracy, 1991, in press) and Vermont (Armstrong and Tracy, 1991; Armstrong, 1992) and in amphibolites in Vermont (Kohn and Spear, 1989; Laird et al., 1991) have yielded temperatures up to 730°C and pressures up to 10.5 kbar. A detailed review of these results can be found in Armstrong et al. (1992). In the main part of the WA belt, higher-P conditions are typical for most of the kyanite-staurolite and kyanite-staurolite-sillimanite assemblages. Virtually identical pelitic porphyroblast assemblages (grt-ky-sta-ms-bt-rt-ilm) occur at virtually identical P-T conditions (650-750°C at 9.5-10.5 kbar) in rocks as far apart as Gassetts, Vermont and Waterbury, Connecticut. In Connecticut, the P was high enough (up to 10.5 kbar) to stabilize muscovite in pelitic rocks at T as high as 730°C. Documented conditions close to one of our stops near Waterbury (just outside the Waterbury Dome) indicate that muscovite is at or near the limit of its stability within the kyanite field (Kerrick, 1991). This observation is important in light of reported kyanite-K-feldspar assemblages within the Waterbury Dome (O'Connor, 1978; Carmichael, 1978; Dietsch, 1988). We believe the structural relief of the Waterbury Dome has allowed exhumation of slightly higher-T rocks in which muscovite has broken down to kyanite-K-feldspar during associated migmatization (see later section on high-grade metamorphism and melting in the WA belt).

The style of metamorphism in the WA belt indicates virtually no contribution from magmatic heat, in contrast with the EA belt. There is a virtual absence of post-420 Ma plutons in western New England (Fig. 1B) except for syn- and post-Acadian pegmatites and the various bodies (many pegmatoid) of approximately 375-370 Ma Black Mountain Granite in southern Vermont. Metamorphism appears to have occurred purely as a result of tectonic/sedimentary overthickening (largely from the east, we believe) and resultant thermal relaxation and advective heating during exceptionally rapid unroofing, documented by Hames et al. (1989) and discussed by Armstrong et al. (1992), who report a calculated integrated uplift rate of 1.4 mm/yr and cooling rate of 13°C / m.y. We believe that these high uplift and cooling rates had a significant influence on the lack of significant migmatization in very high-T rocks in western Connecticut: there simply wasn't time for melts to form and coalesce (see further below).

Several recent Ar studies have provided constraints on the timing of WA metamorphism and post-metamorphic cooling history. Sutter et al. (1985) noted that most of the hornblende Ar cooling ages from the WA are in the range of 376-365 Ma and probably reflect cooling from a metamorphic peak achieved no later than 380 Ma. Hames et al. (1991) reported a 388 (±2) Ma cooling age on ky-grade rocks in SW Massachusetts, the same age as that obtained by Sutter and Hatch (1985) for staurolite-grade hornblende near the VT-MA border. Other studies summarized in Armstrong et al. (1992) suggest that a timing of 390 Ma for attainment of peak temperatures and associated pressures in the highest-grade rocks of the WA belt is entirely reasonable. Note that this timing puts the WA peak approximately 20 m.y. or more *after* the peak in the EA belt - magmatic arc.

### Alleghanian Metamorphism

The map in Figure 1B shows a fairly minimal and conservative interpretation of the geographic extent of late Paleozoic metamorphism in southern New England which might actually be much larger. The uncertainty in extent is primarily due to the fact that Alleghanian metamorphism almost everywhere is overprinted on at least one earlier metamorphism in older rocks, and the evidence for this later thermal disturbance may be very subtle. Two exceptions involve metamorphism of Carboniferous-age sedimentary rocks in eastern New England: low-grade rocks near Worcester, Massachusetts, and medium to high-grade rocks in the Narragansett Basin of Rhode Island and southeastern Massachusetts. These rocks were deposited after the Acadian, and thus have unequivocally been metamorphosed in the late Paleozoic Alleghanian orogeny. Elsewhere, the process of sorting out the presence of an Alleghanian thermal disturbance typically involves rather sophisticated geochronologic and petrologic techniques and interpretations (e.g., Wintsch and Aleinikoff, 1987; Harrison, Spear and Heizler, 1989; Gromet, 1989; Getty and Gromet, 1992) in both metamorphic and plutonic rocks.

In the portions of southern and coastal New England in which the occurrence of Alleghanian high-grade metamorphism and anatexis has been accepted for some time (e.g., Zartman and others, 1970; Wintsch and Sutter, 1986; Wintsch and Aleinikoff, 1987), the status of pre-Alleghanian metamorphism remains uncertain, particularly in southeastern Connecticut. In eastern Connecticut, north and west of the boundary with the Avalonian terranes, geologic correlations (e.g., Robinson and Hall, 1980) and presumed stratigraphic continuity with rocks to the north in the CMT and the Bronson Hill belt suggest that it is not unreasonable to presume relatively high-grade Acadian metamorphism overprinted by a later Alleghanian event. Getty and Gromet (1992) argued that the Acadian disturbance in the Willimantic area of eastern Connecticut (and perhaps farther to the north and west in the Bronson Hill belt) is restricted to cover rocks, and that Avalonian basement gneisses such as those exposed in the core of the Willimantic Dome show no isotopic evidence of a Devonian thermal disturbance. The implication of this interpretation is that basement and cover in much of southeastern New England were tectonically juxtaposed after the Acadian, but then deformed and metamorphosed together in the Alleghanian.

Permian age plutons serve to stitch together the underlying Avalonian basement and Devonian cover sequences; in our fieldtrip area we will see material derived from a pluton interpreted to show such a stitching relationship, the Stony Creek Granite (Stockman and McLellan, 1987) which is believed to be analogous to the unambiguously Permian-age Narragansett Pier Granite. However, the possibility remains that Avalonia docked with North America at several different times. For example, Wintsch and Aleinikoff (1987) argued strongly for an absence of Acadian effects in Avalonian rocks south of the Honey Hill - Lake Char fault system, but noted the presence of Devonian plutons in these rocks in Rhode Island (Hermes and Zartman, 1985). In summary, progress is being made in unravelling the complex tectonometamorphic relationships in southeastern New England, but an enormous amount of work remains to be done. The paper by Getty and Gromet (1992) provides an excellent recent summary of the status of work in the area.

Alleghanian metamorphic grade in the post-Acadian rocks of the Narragansett Basin has been documented to be as high as sillimanite, and there are extensive kyanite-staurolite-grade rocks in the southern part of the basin adjacent to the Late Paleozoic (ca. 275 Ma) Narragansett Pier Granite (Hermes and Zartman, 1985). Farther west, in the area we will visit on this trip, Alleghanian metamorphism in the Avalonian basement sequence reaches sillimanite+K-feldspar grade, with notable development of anatectic migmatites (Lundgren, 1966; Wintsch and Aleinikoff, 1987). Recent work on rocks west of New Haven and near Bridgeport, Connecticut has underscored the possibility of relatively high-grade Alleghanian metamorphism along the coast in western Connecticut (Lanzirotti, 1993; Lanzirrotti and Hanson, 1993). Detailed geochronologic work to the north in Waterbury, however, has indicated the absence of even a minimal late Paleozoic thermal disturbance in inland areas of western Connecticut (Dietsch and Sutter, 1987).

Critically important for the understanding of migmatite formation as a product of tectonic history is a determination of whether the necessary heat is supplied dominantly by conductive heating of overthickened crust or to convective transport from adjacent magmas (McLellan, 1989). The models of Alleghanian orogeny proposed by Getty and Gromet (1990) can be interpreted to suggest that heat for anatexis is supplied initially by underthrusting of Avalonian basement beneath Devonian cover rocks. This initiates melting in Avalonian basement to produce granitic plutons which subsequently rise through the crust and advect heat to high crustal levels allowing additional

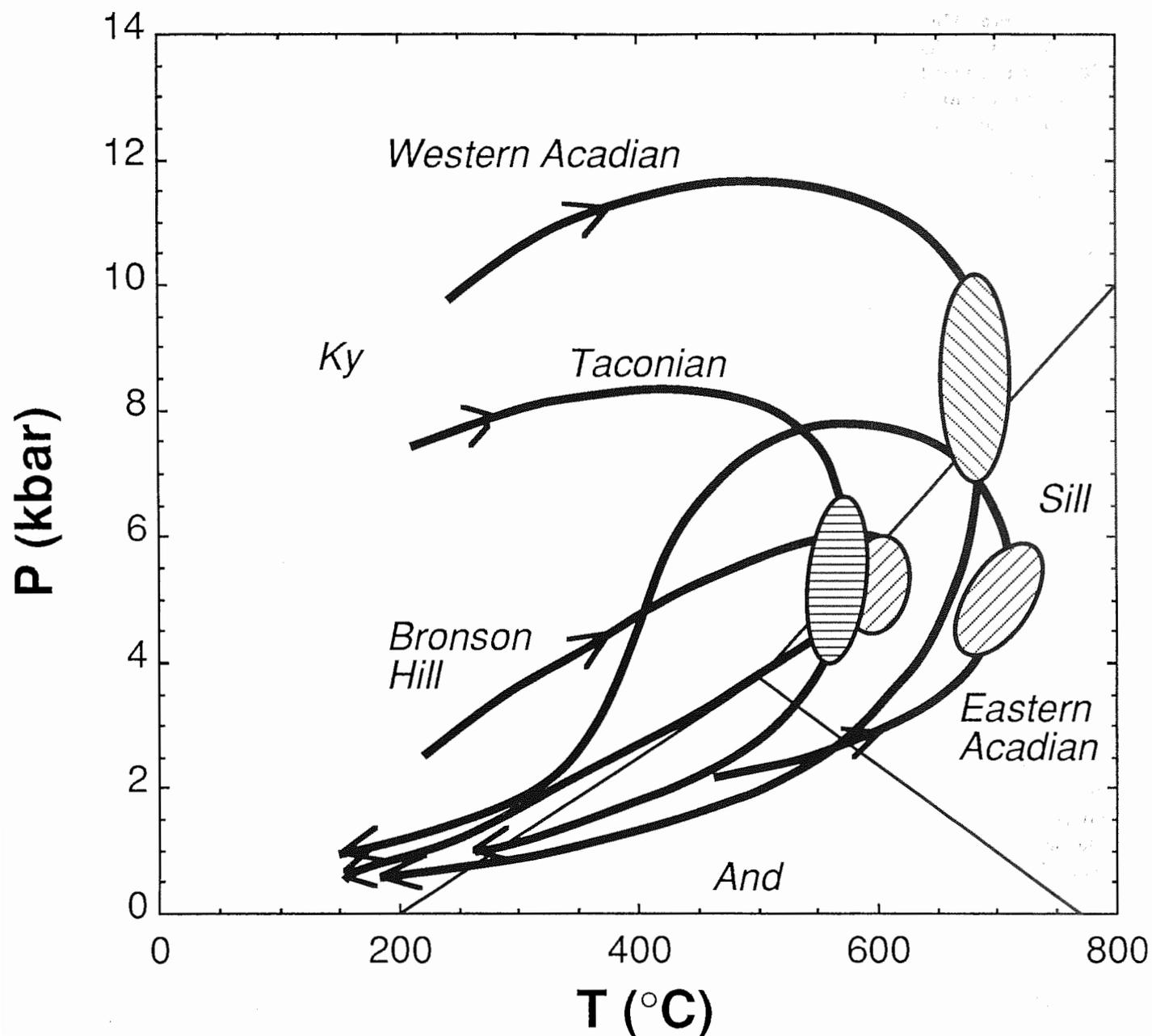


Figure 2. Summary P-T diagram showing the postulated typical shape of P-T trajectories for each of the metamorphic realms. Patterned ovals indicate the part of each path along which peak metamorphic P and T were recorded. Note that for broad belts such as the Western Acadian and Taconian in which a large range of peak metamorphic grade exists, we have shown only one typical path from a nested family of such paths. (From Armstrong et al., 1992)

anatexis in their vicinity. Further, their model calls for extensional unroofing of the overthickened crust, and this could promote melting as rock packets cross vapor-absent melting reactions during decompression. We will examine some evidence for this on our trip.

### Synthesis of Paleozoic Tectonometamorphism in Southern New England

The above brief discussions of tectonometamorphic evolution of the Taconian, Eastern Acadian, Western Acadian and Alleghanian realms in southern and central New England emphasize the tectonic complexity of the region and highlight the necessity of thinking in terms of exotic terranes (Zen, 1983) and of the possibility (even likelihood) that some of the metamorphic belts described were once far more widely separated than they now appear following the final accretionary events. Unfortunately, the present state of detailed structural, petrologic and geochronologic constraints (and even basic geologic mapping in many areas) for the whole of the region is not such as to permit more than a highly preliminary reconstruction or synthesis.

The earliest Paleozoic metamorphism in the southern New England terranes occurred in the Taconian belt of western New England, apparently related to closure of Iapetus Ocean. Convergence of the volcanic/magmatic arc above an eastward-dipping subduction zone with the Laurentian North American passive margin resulted in imbrication of marine sedimentary and volcanic rocks westward onto the parautochthonous platform carbonate and clastic rocks of the North American margin (Stanley and Ratcliffe, 1985) (all geographic references are to *present* orientations, and not meant to imply paleogeographic orientation). Thermal relaxation in the accretionary wedge, apparently little-aided or unaided by magmatic heat input, produced a Barrovian metamorphic terrane with "clockwise" P-T paths producing a Metamorphic Array or Metamorphic Field Gradient (MFG) roughly along the kyanite-sillimanite boundary in the P-T plane (Fig. 2). (Some parts of the Taconian belt in northern Vermont apparently contain relict early blueschist-type metamorphism overprinted by main-phase Taconian with a Barrovian MFG - Laird et al., 1984; Laird, 1987.) Taconian metamorphism appears to be substantially diachronous along the belt, from earliest events at about 500 Ma, or even slightly older, to late metamorphism in the Taconic allochthons of the foreland belt in NY at 445 Ma.

The final phases of the Taconian event were almost certainly overlapped in time by sedimentation in basins both east and west of the volcanic/magmatic arc (the future Bronson Hill belt). The eastern basin, the CMT, contains a thick package of metasedimentary rocks that represents a depositional sequence of Middle Ordovician through Late Silurian or Early Devonian sediments derived from a western source. Thermal maturation of this back-arc realm probably included major inputs from radiogenic heat as well as crustal extension and overthickening, and resulted in major crustal anatexis at 420-440 Ma. Relatively low-P (Buchan-type) EA metamorphism probably commenced by 420 Ma along the early part of what would become an "anti-clockwise" P-T path (Schumacher et al., 1989). Following this part of the metamorphic history of the EA belt, there was apparently a strongly compressional phase that created the so-called "nappe-stage" of deformation in which major east-to-west tectonic transport occurred in thrust nappes. This was accompanied by upward "mushrooming" of granitoid plutons as they were squeezed upward in the crust. This upward movement of more mobile hot plutonic rocks led to ultimate high-temperature compression of the metamorphic infrastructure and the increasing-P portion of the "anticlockwise" P-T path (Fig. 2) - see below.

There appears to have been a major transition about this time from westward-derived Silurian sediments (Rangeley to Lower Madrid Fms.) to an eastward-derived flysch sequence (Littleton Fm.) which ultimately covered the old Taconian magmatic arc terrane and the western "forearc" basin. The source of this flysch is problematic, but may be an eastern highland created by an outboard collision of the CMT with the eastern New England Avalon Block *sensu stricto*. Following this shift in sedimentation, there was apparently continued or renewed E-W compression and shortening in the EA belt that culminated in the 415 to 410 Ma nappe-stage deformation in which shallow crustal material was displaced to the west along fold and thrust nappes. As plutons migrated upward through the crust in the CMT, what might be described as "crustal overturn" occurred, and rocks which had been metamorphosed at low P *above and next to* plutons were rotated into deeper crustal levels *beneath* the mushroom-shaped or tabular plutons. Petrologic evidence cited above demonstrates that the first 100-200 C of cooling proceeded relatively slowly before significant decompression occurred.

The Western Acadian metamorphism was the end result of tectonic/sedimentary loading by the westward-directed nappes and associated flysch sedimentation. Recorded metamorphic pressures in the WA belt of up to 10.5 kbar indicate the existence of *at least* 30-35 km of material above the present erosional level. The substrate for the loading includes the Taconian assemblage of Grenville (Laurentian) crust, autochthonous or paraautochthonous pre-Silurian metasediments and the remains of the Taconian accretionary wedge and fore-arc. Onset of loading is heralded by deposition of the Silurian-Devonian Gile Mountain (VT), Goshen (MA) and Straits (CT) Formations. Peak metamorphic conditions of up to 700-730°C at 9.5-10.5 kbar appear to be completely the result of thermal relaxation with virtually no magmatic heat input (Armstrong et al., 1992). Like the earlier Taconian metamorphism, the WA is characterized by a typically Barrovian MFG that lies essentially along the kyanite-sillimanite P-T boundary, but extends to considerably higher P and T than any known Taconian conditions. Cooling and unroofing in the WA belt were very rapid, in contrast to the style of the EA belt. In fact, the drawn-out end stages of EA tectonism probably overlapped the onset of high-pressure metamorphism in the WA belt. The only plutonic rocks associated with the WA belt from central Vermont southward appear to post-date the high-grade metamorphism by 20-30 m.y. and are typically peraluminous granitoids. Most are actually pegmatites or coalesced pegmatite complexes on the scale of small stocks, except for the Black Mountain Granite pluton and related dikes in Vermont. They have been interpreted by Hames et al. (1989) and Armstrong et al. (1992) to represent the uppermost penetration of anatectic melts from the deeper crustal levels (>35 km) which resided long enough at super-solidus conditions for widespread melting to occur.

Although the peak of late Paleozoic (Alleghanian) thermal activity in most places seems to have been Permian (ca. 275 Ma), there is some evidence of magmatic activity in the CMT of northwestern Maine as old as 320 Ma (Lux and Guidotti, 1985). As noted by Wintsch and Aleinikoff (1987) and Gromet (1989), magmatism and associated high-grade metamorphism in southeastern New England in the Permian probably records the accretion of Avalonian terranes to eastern New England as part of the final assembly of Pangea. Thus where metamorphism occurs in unequivocal Avalonian crust along the Connecticut coast east of New Haven and south of the Honey Hill - Lake Char faults (where we will examine it), tectonic interpretation is relatively straightforward. However, the interpretation of mid-crustal late Paleozoic thermal disturbances and ductile deformation that extend quite far inland to the north and west is considerably more problematic. Documentation of apparent thermal decoupling of cover and basement in the Willimantic and Pelham Domes between 650 and 275 Ma (Getty and Gromet, 1992; Gromet and Robinson, 1990) requires post-Acadian and pre-Alleghanian juxtaposition of these tectonic elements. It also requires that the intense deformation of the Bronson Hill gneiss dome terrane (and the formation of the domes themselves) be at least in part, if not completely, Alleghanian. It is at present unclear how much of the high-grade metamorphism in the EA belt that has been previously assumed to be Acadian may in fact be Alleghanian. Clearly, much additional work on this problem is required.

### CHARACTERISTICS OF HIGH-GRADE METAMORPHISM AND MELTING IN THE EASTERN ACADIAN - CENTRAL MASSACHUSETTS AND NE CONNECTICUT

In the Eastern Acadian belt, the highest grades of metamorphism and the area of most intense migmatization are in central Massachusetts and northeastern Connecticut (Fig. 3). An excellent recent summary of metamorphism and melting in this terrane may be found in Schumacher et al. (1989), which we will briefly summarize here. Many of our central Massachusetts stops are in the highest grade zone (VI), which was defined on the basis of coexistence of *grt + crd + sil + kfs* in metapelites and in which the widely accepted granulite assemblage *opx + kfs + qtz* occurs in some mafic bulk compositions. As shown in Figure 3, metamorphic temperature in Zone VI commonly exceeds 650°C, and in south-central Massachusetts and northeastern Connecticut, exceeds 700°C over a fairly wide area. On the field trip, we will examine both pelitic and mafic migmatites in Zones IV through VI.

Metamorphic Zone IV is characterized by the first appearance in metapelitic rocks of K-feldspar resulting from the gradual breakdown of coexisting muscovite, and the subsequent transition to Zone V marks the final disappearance of muscovite. Both of these zonal boundaries (III-IV and IV-V) are approximate because they are bulk-composition controlled, particularly by Ca/Na in plagioclase, and in fact Zone IV represents the so-called second sillimanite isograd smeared out by multivariance (Tracy, 1978). The common metapelite assemblage in Zone IV is

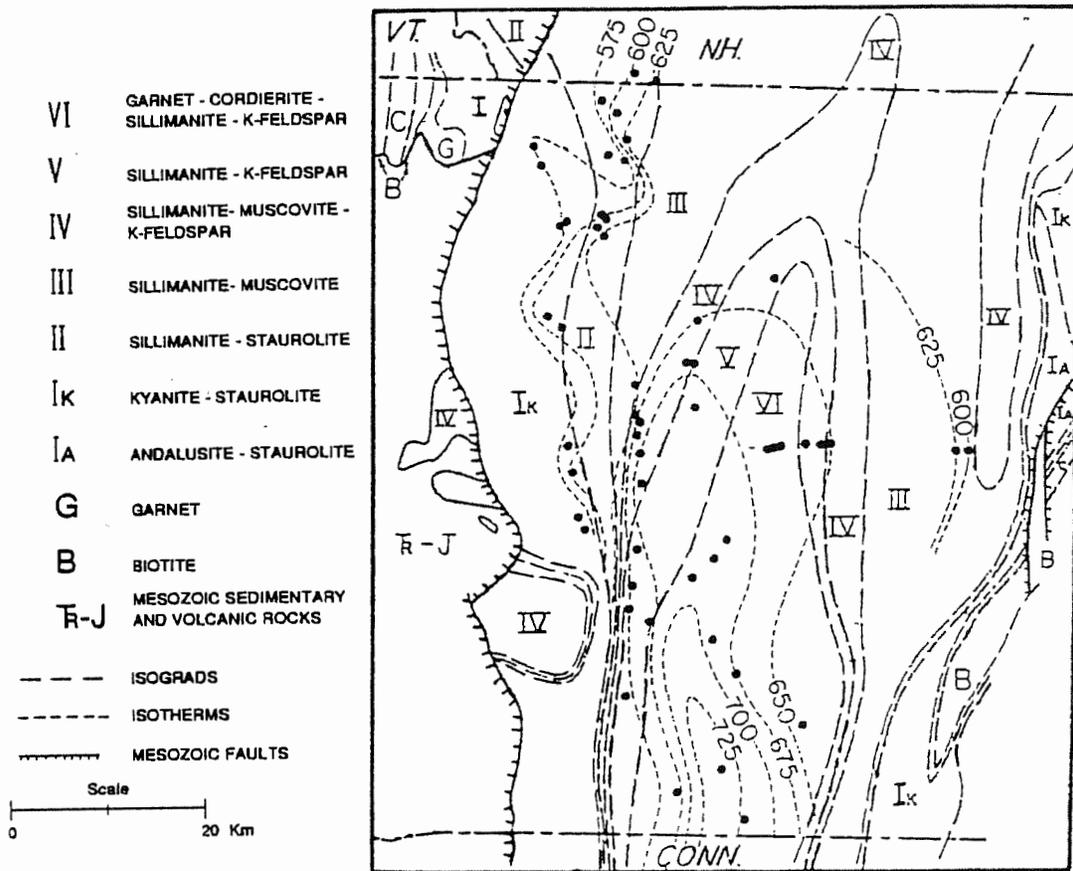


Figure 3. Contour map of the highest-grade part of the Eastern Acadian belt, showing the positions of estimated isotherms of peak metamorphism (short-dashed lines) and boundaries of metamorphic zones (long-dashed lines) for central Massachusetts and northern Connecticut. (From Schumacher et al., 1990).

thus *qtz-pl-ms-kfs-grt-bt-ilm* (*gr-po*) and that in Zone V is similar, but lacking *ms*. Zone VI is marked by the assemblage *qtz-pl-kfs-grt-bt-crd-ilm* (*+gr-po*) in typical pelites. In Zones IV and V, stromatic (layered) migmatites are especially common, and there appears to be a significant connection between muscovite dehydration and incipient H<sub>2</sub>O-saturated melting (Tracy, 1978; Tracy and Robinson, 1983). (Interestingly, Lundgren (1966) found essentially identical melting relations in Alleghanian migmatites of southeastern Connecticut.) There is a strong involvement of Ca in controlling the combined melting-dehydration reactions in metapelites, with melting beginning first in Na-rich bulk compositions and proceeding to Ca-richer compositions (Tracy, 1978; Thompson and Tracy, 1979). Fluid-saturated melting has been inferred as the cause of most of the stromatic migmatites in Zones IV through VI (Tracy and Robinson, 1983).

Fluid-absent metapelite melting involving the breakdown of biotite has been proposed by a number of investigators (Thompson, 1982; Clemens and Wall, 1981) and has been suggested as the origin of vein-type migmatites in Zone VI by Tracy and Robinson (1983) and Robinson et al. (1989). Especially large subhedral to euhedral garnet porphyroblasts appear to grow in the patchy leucosomes of this type of migmatite, similar to the occurrence described by Powell and Downes (1989). In one sample analyzed in detail (Robinson et al., 1989), the experimental equilibria for granite melting and experimental data for the six-phase assemblage *qtz-kfs-sil-grt-crd-bt* (Holdaway and Lee, 1977) were combined to estimate P, T and relative H<sub>2</sub>O activity. The calculated pressure of 5.2 kbar and temperature of 705°C agree well with independent thermobarometric estimates of 6.2 kbar and 685°C, and the aH<sub>2</sub>O estimate of 0.3 to 0.4 seems reasonable.

Similarly, mafic rocks in the EA belt have been shown to have melted in two ways, through fluxing by an H<sub>2</sub>O-rich fluid phase and through dehydration melting (Hollocher, 1985; Robinson et al., 1986; Schumacher et al., 1989). Hollocher has reported veins, anastomosing veins, small dikes and pods of igneous-appearing tonalite in amphibolite outcrops from Zone V, and has suggested two melting reactions involving the breakdown of cummingtonite. These were originally interpreted as fluid-absent, but now are thought to represent melting in the presence of a fluid phase because the estimated P-T conditions of the outcrops coincide with the H<sub>2</sub>O-saturated tonalite melting curve of Wyllie (1977). Simplified somewhat, these reactions are  $cum + pl + H_2O = hbl + grt + melt$ , and  $cum + pl + H_2O = hbl + opx + melt$ . Combined with subsolidus reactions, they produce a melting P-T grid for Zone V mafic rocks shown in Figure 4. In Zone VI, fluid-absent melting and hornblende breakdown appear to have caused a second episode of melting of mafic rocks, in the reaction  $hbl + qtz = opx + cpx + pl + melt$  (Robinson et al., 1986). Figure 5 (from Schumacher et al., 1989) summarizes information from central Massachusetts on estimated P and T for the six metamorphic zones and the probable P-T conditions of melting in both mafic and pelitic rocks.

#### CHARACTERISTICS OF HIGH-GRADE METAMORPHISM AND MELTING IN THE WESTERN ACADIAN - WESTERN CONNECTICUT

The highest metamorphic grades yet observed in the Western Acadian belt, along with the highest recorded P and T, have been found in the Waterbury area (see Fig. 6). Metamorphic grade in metapelites reaches *ky + kfs* in the Waterbury structural Dome (O'Connor, 1973; Dietsch, 1988), although outside the dome the maximum grade is typically *ky + st* or *ky + sil + st*, and all metapelites are *ms*-rich and free of *kfs*. There are rare patches of metapelitic rocks which contain *sil* alone, without *ky*. Recorded metamorphic temperatures are rather high for these grades, reflecting the unusually high pressures. Immediately north of Waterbury (Fig. 6), some pelites have yielded T up to 730°C at 10.5 kbar (Miller, 1990; Miller and Tracy, 1991, in press). Although one is initially surprised that abundant muscovite remains in such high-T rocks, examination of Figure 7 shows why. The estimated P-T conditions for most samples lie within the muscovite stability field at these high pressures (Holdaway, 1971; Kerrick, 1972; Thompson, 1982; Rubie and Brearly, 1990), and are consistent with the *ky-ms* coexistence in most samples.

Further examination of Figure 7 reveals that at temperatures only slightly higher than those commonly observed outside the dome, muscovite in *kfs*-absent pelites would break down in the simplified reaction:  $ms + pl + qtz = ky + kfs + bt + fluid$  (or melt). Of course, at these pressures the simple dehydration reaction is almost certainly metastable relative to the dehydration-melting equilibria, and the production of *ky-kfs* assemblages should also

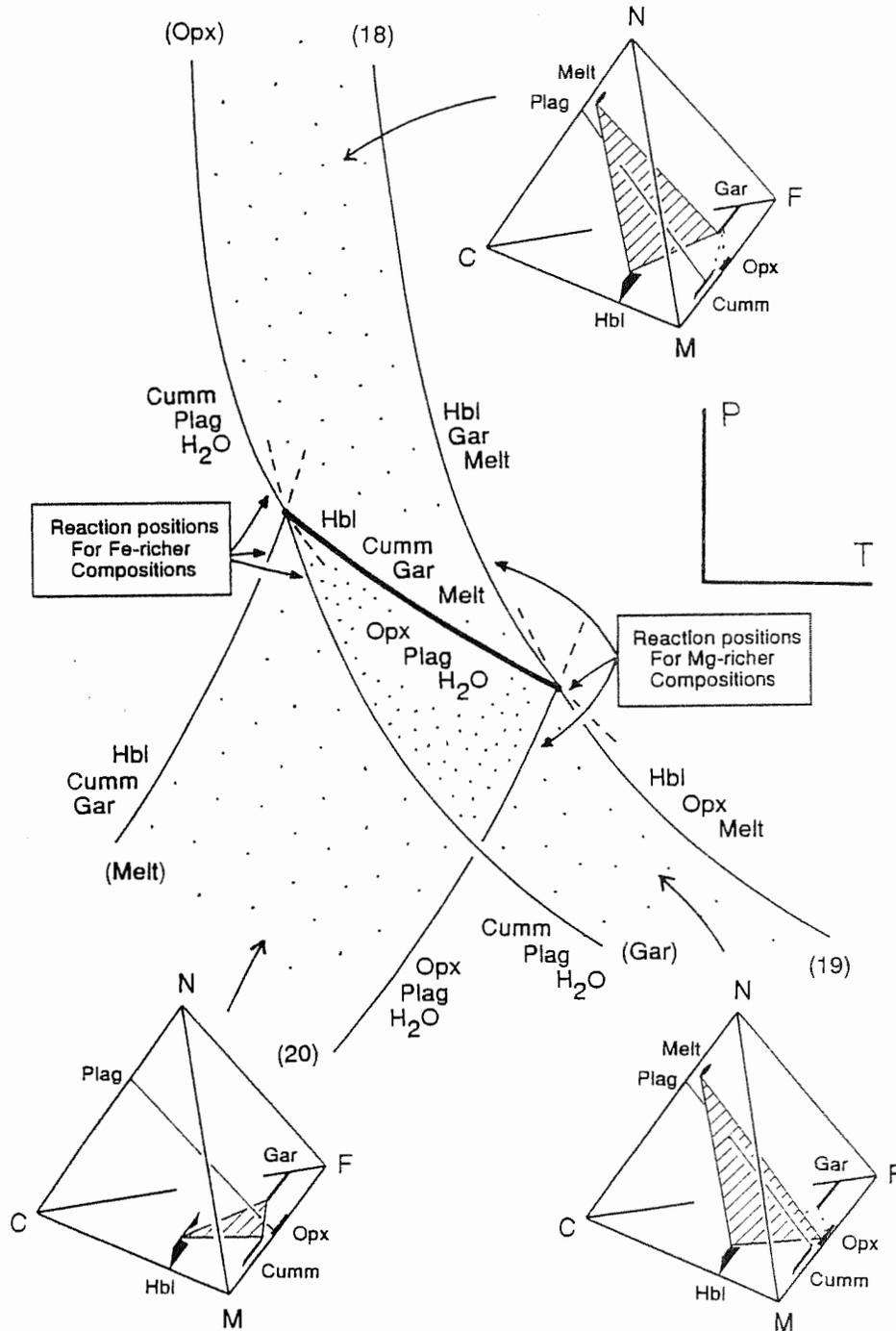


Figure 4. Hypothetical P-T diagram for divariant melting reactions in cummingtonite-bearing amphibolites, derived from combination of a subsolidus reaction described by Hollocher (1985) and two melting reactions from Robinson et al. (1986). Thin lines show relative positions of analogous sets of Fe and Mg reactions; numbers on reactions refer to reactions given in Schumacher et al. (1990). Several possible additional reactions have been left off for clarity. Stipled areas define P-T regions where divariant reactions occur, and the heavy line is a possible Fe-Mg univariant reaction. Abbreviations: cumm - cummingtonite, gar - garnet, hbl - hornblende, opx - orthopyroxene, plag - plagioclase. (From Schumacher et al., 1990)

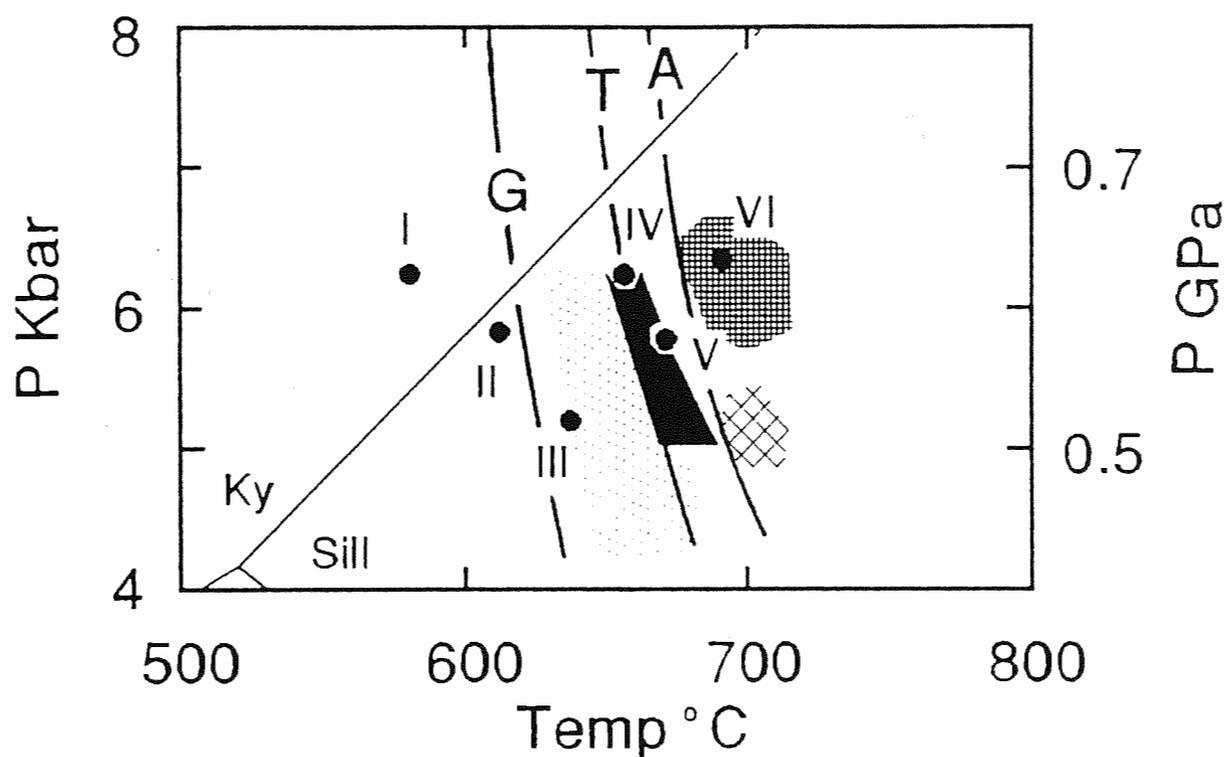


Figure 5. P-T summary diagram from Schumacher et al. (1990) showing average metamorphic conditions for the six metamorphic zones of central Massachusetts and adjacent areas (see Fig. 3), and showing the inferred approximate fields for melting in various bulk compositions. Stippled area - dehydration and melting of muscovite in metapelitic rocks (Tracy, 1978); black area - melting in cummingtonite-bearing amphibolites that accompany cummingtonite dehydration (as in Fig. 4) (Hollocher in Robinson et al., 1986); diagonal cross-hatched area - possible fluid-absent melting in pelitic rocks caused by biotite dehydration; square cross-hatched area - possible fluid-absent melting in mafic rocks caused by hornblende dehydration. Water-saturated melting curves for granite (G), tonalite (T) and amphibolite (A) (Wyllie, 1977) are given for comparison. (From Schumacher et al., 1990)

## Metamorphic temperatures (°C) within the Straits Schist, Western Acadian belt

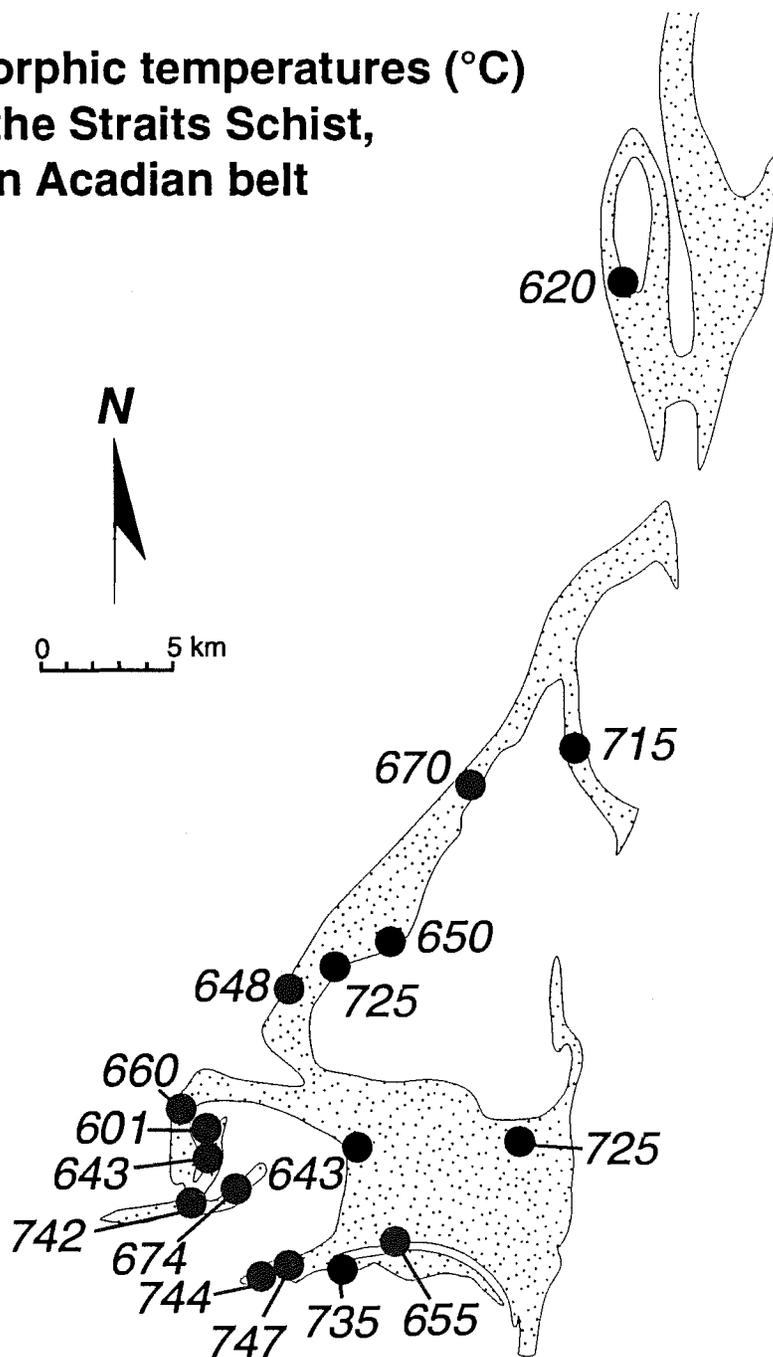


Figure 6. Map showing the distribution of pelitic Straits Schist in western Connecticut with measured metamorphic temperatures. The Waterbury Dome is at the southern end, and the Massachusetts state line is at the northern end of the map area. (Adapted from Miller, 1990; Miller and Tracy, in press)

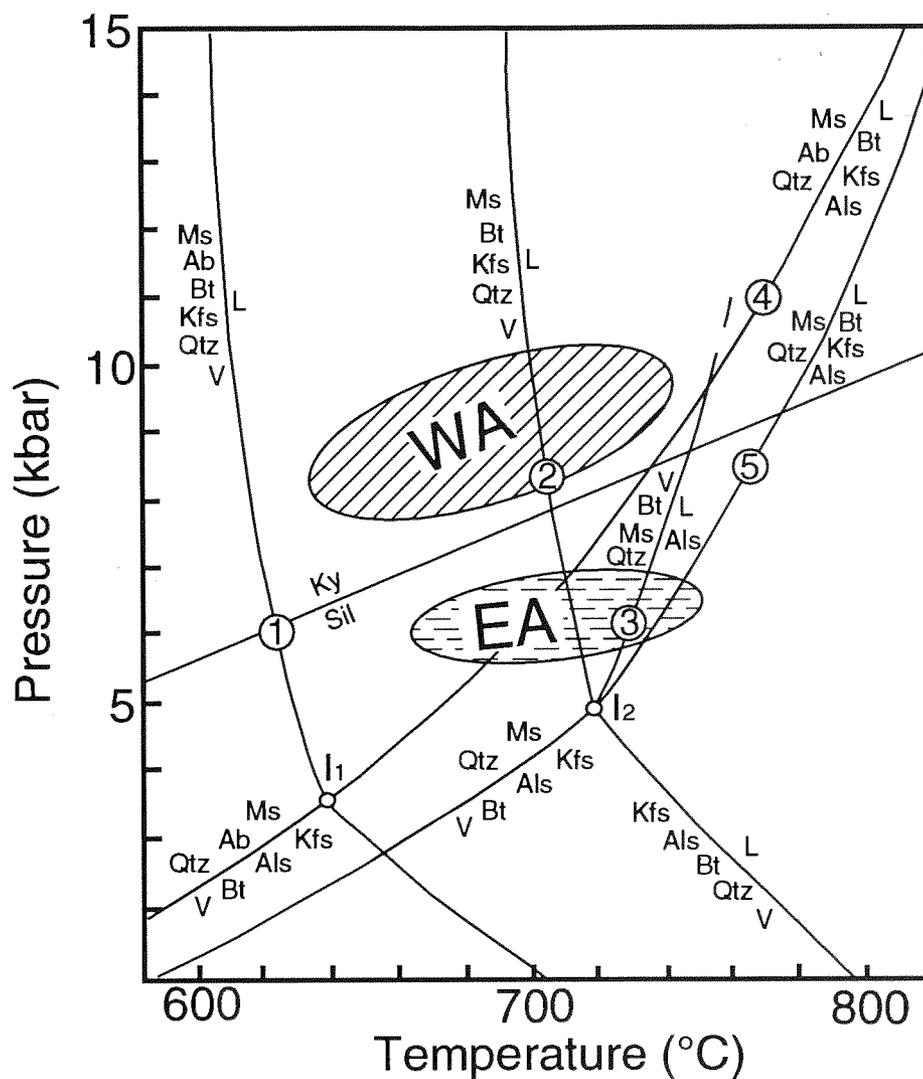


Figure 7. P-T diagram (adapted from Rubie and Brearley, 1990) showing dehydration and melting equilibria from Thompson (1982) along with fields of estimated peak metamorphic conditions for rocks in the Eastern Acadian belt (EA) in central Massachusetts and the Western Acadian belt (WA) near Waterbury, Connecticut. Kyanite-sillimanite boundary is from Kerrick (1991 - Fig. 3.46). Note the large overlap of EA metamorphic conditions with proposed dehydration-melting curves, as compared to lack of overlap of WA conditions with these equilibria.

involve production of granitoid melt (Thompson, 1982). The only *ky-kfs* assemblages known in southern New England occur in the Waterbury Dome in migmatitic outcrops; these rocks have recently been studied by Dietsch (1988). The Waterbury Dome is the southernmost of a series of structural domes that lie to the east of Grenville massifs from Vermont southward to Connecticut, and all appear to be structural culminations resulting from intersecting major anticlinal fold axes. Dietsch (1988) has estimated that the Waterbury Dome has a structural relief of several km, and this would probably be enough to have allowed exhumation of rocks up to several tens of degrees hotter within the dome than outside. The abundance of *ky*-bearing migmatites in pelite outcrops within the dome is probably best explained by local melting driven by local availability of H<sub>2</sub>O due to dehydration of muscovite, in a reaction such as  $ms + pl + qtz = ky + kfs + bt + melt$  (Fig. 7 - see Thompson, 1982)

The relative paucity of migmatites outside the dome is consistent with the fact that the estimated metamorphic conditions of *kfs*-absent samples lie on the low-T side of the dehydration-melting equilibria in Figure 7 portrayed by Thompson (1982). Research is currently underway to better characterize the reactions that went on in these high-grade rocks, but at the moment our speculation is that kinetics may play a major role in both subsolidus and melting equilibria. Petrologic studies noted above have demonstrated that the WA metamorphic event was probably a rapid one - both heating and subsequent unroofing and cooling appear to have proceeded much more rapidly than was the case for the EA belt. Although we know of little data or calculations bearing on this, we can speculate that the equilibrium formation and especially the coalescence of granitoid magmas to form migmatites is probably a kinetically slow process, particularly in the absence of pervasive aqueous fluids, although Rubie and Breatly (1990) presented theoretical calculations and data suggesting that *disequilibrium* melting (with reaction overstepping) may be quite rapid.

The character of the high-grade rocks in the WA belt suggests that only dehydration-melting reactions occurred, and that the extent of melting was controlled by the local availability of H<sub>2</sub>O through mineral dehydration (see Thompson, 1982 and Rubie and Breatly, 1990). This is to be expected where there is no intrusion and solidification of hydrous melts to contribute both to local heating and to local supply of aqueous fluids. On the other hand, the abundance of stromatic migmatites in the EA belt is entirely consistent with the role of plutons in metamorphism there (Fig. 7), and thus the local availability of fluid from solidifying magma. There are some outcrop features in the Waterbury area but outside the dome that indicate at least incipient melting. Localized small (< 1 m) coarsened areas and pegmatoid *kfs*-bearing "sweats" suggest melting was imminent, if not already starting. It is likely, however, as noted above, that most of the rocks simply didn't get quite hot enough or have sufficient residence time at above-solidus temperatures for melting to proceed very far. More voluminous melting in the terrane appears to have gone on only at greater depths (as noted in Hames et al., 1989) where the P-T trajectory spent longer in the P-T region of pelite melting. The abundant 360-385 Ma pegmatite dikes and sills are the likely evidence of deeper magma formation.

#### CHARACTERISTICS OF HIGH-GRADE METAMORPHISM AND MELTING IN THE ALLEGHANIAN - SOUTHERN CONNECTICUT

In contrast with the Acadian metamorphic belts, migmatites in the Alleghanian orogeny have received comparatively little attention. In part this reflects the difficulty of separating Alleghanian from Acadian effects in overprinted terrains, as discussed above, and in part it reflects the relative scarcity of traditional migmatites within the metamorphosed igneous rocks which comprise Avalonian basement. As an example of the former, only relatively recently have migmatites within the Massabesic sequence of New Hampshire (Aleinikoff et al., 1979) and the Pelham Dome of Massachusetts (Tucker and Robinson, 1990) been identified as Alleghanian in age. Direct evidence of anatexis is likewise limited; for example, Wintsch and Aleinikoff (1987) use leucosome chemistry and zircon isotope systematics to infer partial melting of a metasedimentary source at depth.

The relative scarcity of metasedimentary units limits the application of thermobarometric techniques, thus melting reactions are by and large unconstrained. The exception to this is the work of Lundgren (1966) in the Lyme area of southern Connecticut. This area, which we will visit at STOP 16, lies in the sillimanite-orthoclase zone; our thermobarometry suggests conditions of c. 730°C and 6.5 kbar for this region and for some of the gneisses in the Stony Creek area further west. The lack of muscovite in most lithologies means that anatexis is relatively limited; leucosomes are predominantly leucogranitic in composition and appear to develop only in association with high strain zones which may serve to focus water and so promote melting at the granite cotectic. More mafic

lithologies develop garnet-bearing leucosomes which are attributed to fluid-absent melting of biotite-bearing assemblages; the migmatites so produced are similar to those produced in Acadian Zone VI.

Although lack of suitable assemblages limits petrologic interpretation, these migmatites are amenable to detailed structural analysis, particularly in the large pavement exposures of the Connecticut shoreline. Stockman and McLellan (1990) have documented the progressive disaggregation of existing lithostratigraphy in response to increasing anatexis; on this trip we will view this progressive disaggregation at STOPS 12-15. It is particularly important to note that the structural style of migmatites in this region is very different from the typically stromatic geometry seen in the Acadian region; agmatitic and diktyonitic (McLellan, 1988) geometries are common. Thus the mechanism of melt segregation in these migmatites appears to be dominated by hydraulic failure in rocks undergoing simple shear (see McLellan 1988 for a discussion of shear-driven melt segregation). This is perhaps not surprising given the positive  $\Delta V$  of melting associated with dehydration melting reactions which will tend to increase the effective pressure of the melt and so promote rupture.

Thus it appears that migmatite development in the metaigneous rocks of the Avalonian terrain is intimately connected with deformation which both drives melt segregation and provides high-permeability, high-strain regions within which the kinetics of melting reactions are increased. There is some evidence in the Lyme area to suggest that melting and deformation occurred under conditions of rapid decompression (see discussion of STOP 16) which may be attributable to exhumation of overthickened crust developed during final accretion of Avalon to North America, as in the model of Getty and Gromet (1990). The source of heat responsible for this melting is unclear; the major exposed Alleghanian pluton in the region, the Narragansett Granite, lacks associated migmatites. We will examine the presumed analog of this granite, the Stony Creek Granite, at STOP 11, where we will have a chance to view the relation between granite and migmatites. However, the widespread occurrence along the Connecticut coast of pegmatites related to the Narragansett Pier Granite argues for the existence of a series of buried plutons which may have provided the necessary heat and perhaps contributed to uplifting.

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## ROAD LOG

### DAY ONE

Start trip at front entrance to Old Sturbridge Village.

- 0 Junction with Route 20; Turn right onto Route 20 (east)
- 0.3 Stop lights; continue straight on Route 20
- 0.5 Turn left onto New Boston Rd.
- 2.1 Bridge across Mass. turnpike. Park on right beyond bridge; cross over fence on right and head under bridge to large roadcut (north side of turnpike).

### STOP 1 CORDIERITE PEGMATITE

This outcrop is located within a narrow belt of gray-weathering aluminous schist with minor calc-silicate and has been assigned to the Silurian Rangeley Formation. The dominant rock-type is a coarse quartz-orthoclase-plagioclase-biotite-garnet-cordierite-sillimanite-graphite gneiss. Within the gneiss are two sills of garnet and garnet-sillimanite pegmatite with very strong deformational fabric. Along the contacts of the pegmatites are dark layers up to 0.3m thick which consist of biotite, garnet, sillimanite and cordierite. These "selvages" may be either restite layers from melting of the pegmatite or metamorphosed reaction rims between the pegmatite and the surrounding Rangeley

gneiss. In several places there are layers of almost pure cordierite (up to 4cm thick). Thermobarometry from this locality, using garnet-biotite thermometry on garnet cores and matrix biotite yield a temperature of 682°C; The assemblage sillimanite-garnet-cordierite (without quartz) yields a maximum pressure of 6.7 kbars.

Return to vehicle; turn around and proceed south on New Boston Rd. back to Route 20 in Sturbridge.

- 3.7 Intersection with Route 20. Turn left and head east on Route 20.
- 3.8 Turn right onto ramp for I-84 west.
- 7.8 Exit 1 Mashapaug Rd.; take ramp exit. Turn left at bottom of ramp onto old Route 15
- 8.1 Pass under I-84 east.
- 8.5 Park vehicle near entrance to abandoned Sturbridge Isle. Walk to outcrops and blasted rock behind old abandoned fuel island.

### **STOP 2 CORDIERITE PEGMATITE AT MASHAPAUG ROAD**

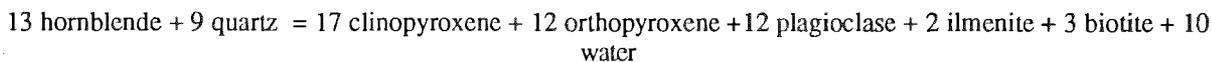
Thomson et al. (1992) interpreted formation of the cordierite-garnet pegmatite to be the result of fluid-absent biotite-dehydration melting reactions of the surrounding gneisses which include quartz-sillimanite-garnet-cordierite schists and interlayered biotite and calc-silicate granulites of the Silurian (?) Paxton Formation. The pegmatite (quartz-plagioclase-orthoclase-sillimanite-biotite-cordierite-garnet assemblage) contains large blue to dark lavender cordierite up to 8cm across. Some large dark patches within the outcrop appear to be pinnite alteration of original cordierite (Thomson et al., 1992). Some samples show evidence of cordierite breakdown during formation of new aggregates of garnet + sillimanite + quartz. Thomson et al. (1992) estimated P-T conditions during onset of cordierite breakdown to be 762°C at 5 kbars using the core composition of garnet aggregates and cordierite compositions taken far from the new aggregates. Using cordierite compositions near or in the aggregate ("final" cordierite compositions), Thomson et al. (1992) estimated P-T conditions of 570 to 590°C at 6.6 to 6.7 kbars, resulting in a "counter-clockwise" P-T path typical for the Eastern Acadian metamorphic regime as discussed by Armstrong et al. (1992).

Return to vehicle; turn around and return to I-84 west

- 8.9 Turn left onto I-84 west.
- 12.0 Cross into Connecticut.
- 13.0 Get off I-84 at Mashapaug Rd. exit (exit to right). Park on grass strip on left handside of road near bottom of ramp. Walk north across ramp to south end of large outcrop.

### **STOP 3 MASHAPAUG ROAD MAFICS AND TONALITIC MELT**

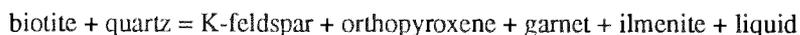
This outcrop is located within the Zone VI granulite facies of Tracy (1978). The rock-types on the SW facing part of the outcrop are pyroxene granulites containing no prograde amphibole. The dominant rock-type is a two-pyroxene granulite that has the proper bulk composition to have been a quartz-bearing amphibolite at lower metamorphic grade. Hornblende has completely broken down through completion of the reaction:



The two-pyroxene rock has undergone partial melting with recrystallized melts forming anastomosing tonalitic veins with coarse-grained pyroxene, quartz, and plagioclase. Melting probably occurred as a result of water influx from dehydration reactions involving hornblende breakdown, as in the reaction above. Thus the hornblende dehydration reaction can be reduced to the melting reaction:



An orthopyroxene- K-feldspar bearing granulite unit (with 10% albite component in the feldspar) occurs within the NW part (left-hand side) of the outcrop as a synformal infold within the 2-pyroxene granulite. This rock contains abundant segregations of coarse-grained orthopyroxene and K-feldspar bearing tonalite, which probably represents partial melts which developed during biotite breakdown (Robinson et al., 1986). The biotite dehydration reaction involved in melt formation can be reduced to:



Robinson et al. (1986) reported on the presence of red-brown biotite-quartz symplectites from both major rock units indicative of high temperature orthopyroxene + K-feldspar + water = biotite + quartz retrograde reaction. Such reactions may have occurred during evolution of rocks through the increasing P - decreasing T part of the regional "counter-clockwise" P-T path as discussed at stop 2. As noted by Armstrong et al. (1992), such a P-T path may have been the result of a regional "crustal overturn" during 420 to 410 Ma crustal anatexis within a back-arc environment or within discrete mylonitic shear zones during subsequent (410 to 385 Ma) Acadian compressional deformation.

- Return to vehicle; turn right at bottom of ramp and proceed north along Holland Rd.
- 14.6 Causeway over Hamilton Reservoir.
  - 15.3 Crossroads in town of Holland (note Windmill !); continue straight.
  - 17.4 Turn left onto Wales Rd.
  - 19.0 Intersection with Route 19; turn left onto Route 19 (south).
  - 20.2 Wales Post Office; turn sharp right onto Monson Rd.
  - 20.9 Monson Rd. bears left; continue straight on McBride Rd.
  - 21.2 Three-way junction; turn right (north) onto Mt. Hitchcock Rd.
  - 21.4 Tennessee gas pipeline; park on right side of road near outcrop.

#### **STOP 4 SILLIMANITE-CORDIERITE SCHIST OF THE RANGELEY FM.**

This stop provides an opportunity to collect fresh samples of sillimanite-cordierite bearing schist of the Rangeley Formation; the precursor to many of the pelitic migmatite bodies seen today. P-T estimates for rocks from this locality are 710°C at 6.7 kbars. Robinson et al. (1986) reported sillimanite pseudomorphs after andalusite from this locality, which may have formed during attainment of peak-T conditions along a "counter-clockwise" P-T path.

- Return to vehicle. turn around and retrace path back to Wales Post Office.
- 22.6 Wales Post Office; turn left (north) onto Route 19.
  - 23.8 Holland Rd. on right; continue straight ahead on Route 19 (north).
  - 26.1 Cross Route 20 and proceed straight ahead to stop light in center of Brimfield. Continue north on Route 19.
  - 29.6 Underpass beneath Massachusetts turnpike.
  - 32.6 T-intersection; turn right onto Route 67.
  - 35.6 Intersection with Route 9; turn left (west) onto Route 9.
  - 39.6 Turn right onto Route 32.
  - 42.7 Turn right onto New Braintree Rd. (just before railroad crossing).
  - 44.6 Park in barnyard on right and walk across hayfield on right (south) to large pavement outcrop.

#### **STOP 5 LITTLETON FORMATION "PASTUREITE"**

This outcrop is part of the Devonian Littleton Formation and is situated within an isoclinal syncline structurally above sulfidic schist and calc-silicate of the Frankestown Formation (Robinson et al., 1986; also known as Fitch Formation further west). The rock was dubbed "pastureite" by A.B. Thompson in honor of this beautifully polished glacially-smoothed outcrop. **Hammering is not allowed!** Along with several subordinate layers of calc-silicate, the rock is composed of medium-grained quartz-orthoclase-plagioclase-garnet-cordierite-biotite-ilmenite schist and gneiss and a network of slightly deformed, cross-cutting felsic veins consisting of quartz-orthoclase-plagioclase-cordierite with 2 to 4cm euhedral garnets. Garnets show little to no internal chemical zoning, indicative of high temperature, post-growth diffusional homogenization. Fe-Mg zoning, related to post growth cation exchange, is apparent along extreme edges of garnet where in contact with cordierite and biotite. Garnet-biotite and garnet-cordierite thermometry both yield temperatures of approximately 685°C for a pressure value of 6.2 kbars derived from coexisting quartz-sillimanite-garnet-cordierite (Tracy, 1976). In addition, sillimanite appears to be pseudomorphous after andalusite. The veins are believed to have formed by melt segregation at or near peak temperature conditions, giving rise to the coarse-grained nature of feldspar and garnets within the veins. M.J. Holdaway has suggested that aqueous fluid may have been largely carried away during earlier stages of melting, perhaps at the breakdown of muscovite and that the present texture is due to fluid-absent melting during breakdown of biotite in the assemblage; quartz-sillimanite-K-feldspar-garnet-cordierite-biotite.

- Return to vehicle; retrace route back to Route 32.
- 46.5 Turn right (north) onto Route 32.

- 46.7 Intersection with Route 32a; proceed straight ahead on Route 32a (north).
- 46.8 Village of Gilbertville.
- 49.8 Town of Hardwick; turn left (west) onto Greenwich Rd.
- 52.5 Gate 43- Quabbin Reserve; pass through gate and continue ahead (NOTE: gate is locked; permission of access for this trip obtained through the Massachusetts Water Resources Commission, Quabbin Division; James Holeva, acting Superintendent). Proceed straight ahead (west) on road to intake works.
- 54.5 Turn right (north) onto Baffle Dam Rd. and drive across south baffle dam.
- 54.7 Park on left and walk northwest along the southwest shore of baffle dam island.

#### **STOP 6 BAFFLE DAM PELITIC MIGMATITES**

This stop provides an example of the development of stromatic migmatite within the Ordovician Partridge Formation in the sillimanite-muscovite-orthoclase zone (IV) of Acadian metamorphism. Since the migmatitic layers are isoclinally folded, they appear to have formed prior to or early in the Acadian nappe-stage deformational history. Subordinate rock-types at this locality include massive and layered amphibolites and well bedded garnet-rich quartzose granulite. This may be a calcareous variety of typical coticule (Robinson et al., 1989). Point counted modes reported by Tracy (1978) indicate that most samples are dominated by quartz-biotite-sodic plagioclase (An 20-32) with lesser amounts of muscovite, sillimanite, garnet, K-feldspar and trace amounts of graphite, pyrrhotite and Fe-Ti oxides. Most schist samples contain coexisting muscovite and K-feldspar. Garnets in this vicinity are more magnesian than in other localities which include garnet-bearing Partridge sulfidic schist. Tracy (1978) and Robinson et al. (1989) suggested that the presence of K-feldspar in these rocks indicates onset of the biotite dehydration reaction and the formation of garnet and K-feldspar. Water liberated through this reaction and muscovite breakdown probably allowed melting, with water dissolved in the resultant melts. The relatively limited amount of melt coupled with the presence of phases critical to melt production within the restitic zones, suggests that reaction progress related to melting may have been interrupted. Robinson et al. (1989) suggested that limited amounts of available aqueous fluid in reaction zones was the critical terminating factor in melt reaction progress. Since only water-saturated melting could have occurred for the P-T conditions documented here (650-675°C at 6 to 7 kbars), melt reaction progress was probably dependent upon the availability of fluids produced through local breakdown of hydrous phases.

Return to vehicle; retrace route to Hardwick and Route 32a.

- 54.9 Turn left (east) onto intake works road and proceed east toward gate 43.
- 56.9 Proceed through gate 43 and continue east on Greenwich Rd.
- 59.6 Town of Hardwick; turn right (south) onto Route 32a.
- 61.6 Village of Gilbertville
- 61.7 Intersection with Route 32; proceed straight ahead on Route 32 (south).
- 61.9 New Braintree Rd. on left (Stop 5); continue straight ahead on Route 32.
- 65.0 Intersection with Route 9; turn right (west) onto Route 9.
- 66.2 Bridge over Ware River.
- 66.6 Junction of Routes 9 and 32 in center of Ware; Bear to right and continue west on Route 9.
- 69.1 Crest of Brimstone Hill; park on left side of road and walk across to north side of roadcut-  
WATCH FOR CARS AND STAY AWAY FROM ROAD!

#### **STOP 7 BRIMSTONE HILL**

This outcrop is located within the Greenwich syncline, an infold of Ordovician cover rocks surrounded by the main body of Monson Gneiss (Robinson et al., 1986). The rocks at this stop are part of the Ordovician Partridge Formation, and consist of rusty-weathering schist and amphibolite with lesser amounts of metamorphosed felsic and intermediate volcanics, calc-silicate rocks, and coticule. Also within the outcrop are several generations of pegmatite. On the east end of the roadcut (right-hand side) is the Zone V assemblage; quartz-orthoclase-garnet-biotite-sillimanite-ilmenite-graphite-pyrrhotite (Tracy, 1976, 1978). The rock has a mylonitic texture and most of the K-feldspar appears to be concentrated within clear megacrysts with finer-grained mantles set in fine-grained, lined matrix of prismatic sillimanite. Garnets show only minor retrograde zoning along extreme edges of porphyroblasts and are indicative of post-growth, high temperature diffusional homogenization and retrograde Fe-Mg diffusional exchange. This type of post-growth chemical modification is very common within the Eastern Acadian high temperature, moderately low pressure regime, as described by Armstrong et al. (1992). Several of the amphibolite horizons at this stop are associated with tonalitic veins and pods, with the typically coarse mafic

minerals cummingtonite, biotite, hornblende  $\pm$  garnet, These segregations are considered to be partial melt segregations derived from the surrounding amphibolites (Robinson et al., 1986).

Return to vehicle; this is the end of Day 1 trip; the GSA field trip will now continue west on Route 9 to Interstate 91 south. Follow I-91 south to Hartford and get on to I-84 west. Proceed on I-84 west to exit 32, Route 10, Southington.

This is the starting point for the Day 2 roadlog.

## DAY TWO

0.0 Start trip at bottom of entrance ramp to I-84 WEST. Proceed up ramp onto I-84 west.

14.6 Take EXIT 18 West Main St. Bear left and follow signs for West Main St; at bottom of ramp turn right onto West Main St.

14.9 Turn right and park in A&P supermarket parking lot. Walk back up ramp toward I-84; stay on north (left) side of ramp and continue to roadcut.

### STOP 8 TONALITE, TRONDHJEMITE AND GRANULITE IN THE CORE OF THE WATERBURY DOME

Interpreted by Dietsch (1988) as the core of an antiformal "bow-tie" structure within the core of the Acadian age Waterbury dome, this outcrop is composed of rusty-weathering, sulfidic quartz-plagioclase-mica-garnet-kyanite gneiss and schistose gneiss, and blue-gray, medium-grained tonalitic and trondhjemite gneiss and granulite. Also within the outcrop are several horizons of rusty-weathering, fine to medium-grained sulfidic schist and calc-silicate pods. Barometric calculations by Dietsch (1988) using GRAIL (3 ilmenite + kyanite + 2 quartz = almandine + 3 rutile; Bohlen et al., 1983) yielded pressures of 8.8 kbars at  $T = 730^{\circ}\text{C}$  (derived from garnet-biotite thermometry). More recent thermobarometric estimates of  $730^{\circ}\text{C}$  at 10 to 11 kbars were obtained in this area by Miller (1990) on rutile-ilmenite bearing pelite from the overlying Silurian-Devonian Straits Schist. The 8.8 kbar value obtained by Dietsch for these rocks (structurally beneath the Straits) may thus be too low. Higher pressure values for such high temperature estimates (ca.  $730^{\circ}\text{C}$ ) are consistent with the presence of kyanite in these rocks. Of particular importance is the nature of the contacts between the tonalite and trondhjemite with the surrounding granulite gneiss and rusty schistose gneiss; The tonalite/trondhjemite layers are believed to be partial melt products formed from amphibolites situated at slightly deeper crustal levels (and slightly higher temperatures). It is interesting to note that the occurrence of tonalite/trondhjemite is restricted to gneiss units within the core of the dome; no such rock-types are found within the overlying Straits Schist.

Return to vehicles; turn right onto W. Main St after leaving the A&P parking lot.

15.6 Intersection with I-84 and Route 8. Go through first light and under the I-84 overpass to second set of lights (on the far side of the overpass).

15.7 Make a left onto Watertown Ave. at the light on far side of overpass. This will take you to Route 8 North; follow those signs for Route 8, Thomaston Torrington! Head north on Route 8.

22.8 Take Exit 38 for Thomaston; Reynold's Bridge outcrop (Stop 10) is the roadcut to the left on the far side of Route 8 south.

23.2 Bottom of ramp; make a left at the stop light; this will put you onto Routes 6 and 109 (combined). Follow signs for Black Rock State Park.

23.7 Bear right at light onto Route 109; Again; follow signs for Black Rock State Park.

24.5 Turn left at sign for parking area for Black Rock State Park. Park in parking area and proceed west (ahead) to access road over flood control dam. Outcrop of interest can be seen at the termination of access road along the opposite side of dry flood gate.

### STOP 9 STRAITS SCHIST AT BLACK ROCK DAM

The Straits Schist at this locality consists of quartz-muscovite-plagioclase-biotite-garnet-sillimanite $\pm$ kyanite-ilmenite-rutile. Notice the dominant foliation dipping north away from the core of the Waterbury dome, to the south. Also notice the pegmatite and granite dikes which occur both within the foliation and cross-cut it. In several places the dikes appear to be folded by the axial planar dominant fabric, which may indicate that the dominant foliation is composite with younger dome-stage crenulate cleavage (visible on the far left side of the outcrop) or that the dikes actually intruded the Straits during waning stages of dominant foliation development. The presence of

abundant "rose" structures of vein quartz and the general coarse-grained texture of the high grade Straits Schist suggest to us that the rock is on the verge of melting. Discussion will focus on the P-T conditions of melt formation in this area, and the contrasting melt reactions and controlling P-T conditions between the western and eastern Acadian metamorphic regimes.

Return to vehicles; retrace route back toward Route 8 and the Reynold's Bridge locality.

24.5 Turn right out of parking lot onto Route 109

25.3 Bear left at light onto Route 6/109.

25.7 Turn right onto Route 8 south. Park at bottom of on ramp and walk up ramp to look at roadcut on right.

### **STOP 10 REYNOLDS BRIDGE GNEISS**

This roadcut is located within the Collinsville Formation and has been interpreted to be part of the Ordovician arc assemblage, accreted to the Laurentian continent during the Taconian orogeny (Stanley and Radcliffe, 1985). The felsic gneiss and its included and boudinaged and folded amphibolites have been interpreted by us as recrystallized layered volcanics. The deformation is considered to be more complex than that found within the Siluro-Devonian (?) Straits Schist and may be a composite of Taconian (Ordovician) and Acadian (Devonian) deformational events. We will discuss several alternative models for the protoliths and petrogenesis of the felsic and mafic rocks, including origin of the amphibolites as cross-cutting dikes within a felsic intrusive host, and the possible role of post-intrusion partial melting for formation of the different rock types.

Return to vehicles; continue south on Route 8.

27.0 Optional Stop. Roadcut on right is Straits Schist with boudinaged marble horizons, isoclinally folded (and subsequently refolded about upright dome-stage axial surfaces), layers of quartzite and coticule, multiple generations of pegmatite, and 'rose' structures of vein quartz.

32.1 I-84 exit 19; continue south on Route 8.

40.3 Exit 25; continue south on Route 8

51.3 Exit 14; intersection with Route 34 in Shelton. Take exit and get onto Route 34 east (for New Haven) at bottom of ramp. Proceed east on route 34 9.0 miles to junction with I-95, take I-95 N towards Rhode Island for 10.2 miles.

70.5 Take Exit 57, Guilford.

70.8 Turn right on Route 1 at end of off-ramp.

71.8 At 4th traffic light, turn right onto Route 77 south.

72.3 At stop sign, turn right along side of green then turn immediately left onto Whitfield Street to continue on Route 77 south

72.6 Turn right on route 146 signed for "Leetes Island".

73.1 Bear left to continue on Route 146.

73.9 Turn right onto Sam Hill Road.

74.4 Turn left onto Three Corners Road, proceed to top of hill.

74.8. Park in circle, walk up driveway of # 497.

### **STOP 11 GRANITE - MIGMATITE RELATIONS, STONY CREEK**

This outcrop provides a good sample of the Bishop's Hill Gneiss sequence of Stockman and McLellan (1987), which has some geochemical affinity with rocks of the Bronson Hill Anticlinorium seen earlier on this trip. Although this outcrop is clearly migmatitic, there is relatively little in-situ anatexis (due to the relative lack of low-melting compositions in these rocks) and most of the structural complexity in the outcrop can be attributed to syndeformational intrusion by the Stony Creek Granite. Structures in this outcrop, including asymmetric pull-aparts and shear zones, have been interpreted by Stockman and McLellan (1990) as forming during layer-parallel flattening, either associated with granite emplacement or part of a larger regional shear event. We will discuss the difficulties of distinguishing in-situ and intruded melts and will examine evidence suggesting that the internal deformation seen in some gneissic enclaves is related to melt generation processes.

Return to vehicles; retrace route back down Three Corners Road and Sam Hill Road to Route 146.

75.1 Turn right on Route 146 under railroad bridge.

- 75.8 Bear right towards Branford
- 77.4 Turn left on Old Quarry Road.
- 78.0 Entrance to Old Quarry Association property. Bus will drop participants here; private cars can proceed beyond this point. Continue round circle and take second opening signed for "Hall, #133". After 0.3 miles cross bridge onto Narrows Island, follow road 0.1 mile to end of island.

### ***STOP 12 SHEAR ZONES AND DEHYDRATION MELTING, NARROWS ISLAND***

In contrast to STOP 11, these gneisses belong to the Thimble Island Migmatite Complex of Stockman and McLellan (1987), a suite of predominantly felsic gneisses with some geochemical affinity to Avalon rocks as seen further east in Connecticut and in Rhode Island. They are cut by Stony Creek Granite similar to that seen at STOP 11; granitic material related to this appears in several sets of shear zones and we will discuss the temporal/causal relationships between melt formation and shear zone development. In-situ anatexis is relatively limited at this outcrop, although local mafic units show bi+ Ksp + gt clots interpreted as forming by dehydration melting.

- Return to vehicles; retrace route back down Old Quarry Road towards Route 146.
- 78.5 Bear left at "Do Not Enter" sign.
- 78.6 Turn left onto Route 146.
- 78.7 USE CAUTION AT BRIDGE
- 78.8 Turn hard left onto Thimble Islands Road
- 81.0 Bear left on Thimble Islands Road following signs to "Town Dock".
- 81.3 At "Stony Creek Marine and Tackle" bear left to stay on Thimble Islands Road.
- 81.7 Junction of Thimble Islands Road and Flying Point Road. Bus will drop participants here. Turn left onto Prospect Hill Road, enter yard of grey house.

***STOP 13 ANATEXIS AND STRUCTURAL DISRUPTION, STONY CREEK*** This outcrop shows the same gneissic sequence as seen at STOP 12, but here the amount of anatexis is greater as is the amount of intrusive material. The overall structure is correspondingly more complex and it is possible to walk along layers and trace the gradual development of shear zones in parallel with increasing melt percentage. As melt increases the internal foliations within blocks become rotated relative to one another, suggesting that either increasing melting leads to disaggregation of the original stratigraphy or that melts form preferentially in zones of increased strain. We will discuss the development of these structures in terms of a model for the kinetics of melt segregation in which high strain rates can be achieved as melting progresses by switching from matrix-compaction limited segregation mechanisms to melt-flow limited segregation. This outcrop also shows two distinct types of intrusive granite, one related to the Stony Creek Granite seen earlier and a second garnet-rich unit which occurs both as large cross-cutting dikes and as diffuse patches which cross-cut and/or grade into Stony Creek Granite type material. We will discuss whether these garnets crystallized as liquidus phases or are restitic from disaggregated restite schlieren.

Return to vehicles, retrace route along Thimble Islands Road. This is the end of Day 2 trip; the GSA trip will now continue straight at "Stop" sign to junction with I-95.. Turn left on I-95 South towards New Haven. After 1.0 mile take exit 55 for Branford Motor Inn.

**Day 3 road log begins at Branford Motor Inn.**

### ***DAY THREE***

- 0.0 Turn left onto Route 1
- 0.2 Take left onto I-95N towards New London
- 1.2 Take Exit 56 for Route 146.
- 2.8 Turn left onto Route 146 at Stony Creek Package Store.
- 3.4 Turn right into Old Quarry Road again.
- 3.6 Turn left onto Andrews Road. Bus will drop participants here.; continue 0.2 miles down Andrews Road to beach at Little Harbour.

### ***STOP 14 MIGMATITE GEOMETRY AND MELT SEGREGATION***

Once again we see the gneisses of the Thimble Island Migmatite Complex, here including a unit containing  $mt+Ksp$  clots which may be related to biotite dehydration. Large masses of intrusive melt here are associated with regions of structural weakness such as fold hinges and boudin necks. The structural style is very similar to that seen at STOP 13, with early stromatic leucosomes succeeded by later shear zone leucosomes. We suggest that in these outcrops the transition from wet melting to dry melting (controlled by the P-T-t path of the terrain) can trigger changes in melt segregation mechanism such that early stromatic leucosomes form by filter-pressing whereas later dry-melting leucosomes develop shear zone geometry controlled by hydraulic failure.

Return to vehicles, retrace route to Route 146.

- 3.7 Turn right onto Route 146.
- 4.1 Turn right onto Shell Beach Road.
- 4.5 Pass gate house
- 4.7 Bear right onto Joshua Point Road.; bus will drop participants here. Walk 0.2 mile down Joshua Point Road to #234.

### **STOP 15 CONTROLS ON ANATEXIS, JOSHUA POINT**

This is the most lithologically and structurally complex area we will see in this portion of the Connecticut coast. The presence of  $gt+sill+bi+mt$  assemblages in this outcrop allows us to constrain metamorphic conditions to 730°C, 6.7 kbar. A common feature of this outcrop is the occurrence of large (up to 3 inch thick) selvages of sillimanite-rich material adjacent to garnet-bearing leucosome, attributed to dehydration melting of original biotite-bearing metasedimentary enclaves in the predominantly metaigneous suite. In addition to these garnetiferous leucosomes the area shows several types of intrusive garnet granite, cut by very coarse-grained  $bi+Ksp$ -rich pegmatites attributed to the Blackhall suite of Lundgren (1966) in the Lyme area. This raises the question of the areal distribution of pegmatites of this type and the possibility that Narragansett Pier - type plutons may exist at depth along the entire Connecticut coast serving as a source for pegmatites and a heat source for anatexis. Structural style in this outcrop is very variable, ranging from highly sheared to blocky, and correlates with lithology and in turn with amount of melt.

Return to vehicles

- 5.2 Turn right onto route 146.
- 6.5 Turn left to continue on Route 146 towards Guilford.
- 7.8 Follow route 146 back through Guilford (left on Whitfield Street and immediately right onto Boston Street).
- 8.9 Turn left on Soundview Road.
- 10.0 Turn right onto Route 1 and immediately right onto I-95N towards Rhode Island.
- 28.3 Take Exit 70 for Old Lyme immediately after Connecticut River bridge; travel east along Old Shore Road.
- 33.2 Turn right onto Billow Road.
- 33.3 Bus will drop passengers at "Stop" sign; walk 0.1 mile to beachfront.

### **STOP 16 POLYPHASE MELTING, OLD LYME SHORES**

These include the migmatites described by Lundgren (1966). This outcrop shows a much wider range of bulk compositions than we have seen so far, including metabasite units in which leucosomes carry garnet selvages and calc-silicate pods. The overall structure is one in which metabasite units occur as megaboudins within felsic and intermediate units similar to those seen at STOPS 11-15, and internal layering within the felsic units is progressively disrupted by increasing amounts of melt.

These migmatites provide an example of structurally complex migmatites produced during a single orogenic cycle. Stromatic leucosomes of biotite-garnet composition show low-angle discordance to a metamorphic foliation which in turn is axial planar to isoclinal folds. Radiometric dating by Zartman et al. (1988) has shown that rocks as young as 300 my. are deformed by these folds, thus these leucosomes have a maximum age of 300 m.y. The leucosomes have hypidiomorphic textures and miarolitic cavities which indicates that they crystallized after the last penetrative deformation; the presence of miarolitic cavities in rocks which record metamorphic conditions of c. 735°C, 7 kbar suggests very rapid decompression near the peak of metamorphism, perhaps related to gravitational collapse of an overthickened orogen. The stromatic leucosomes are in turn cross-cut by numerous ductile shear zones which contain leucosomes up to 20 cm. wide of garnetiferous leucogranite composition similar to that seen at STOP 15. Dehydration melting to produce such leucosomes would be favored by rapid decompression. Granitic pegmatites cutting both of these suites of migmatites have been dated at 268 m.y. (Wintsch and Aleinikoff, 1987). Thus the very complex geometry seen at this outcrop formed over a time period of less than 32 m.y. during

progressive melting and deformation. This suggests a complex interplay between melting, deformation, melt segregation and migmatite geometry related to the P-T-t evolution of the area.

- Return to vehicles; retrace route to Route 156, Old Shore Road.
- 33.4** Turn right onto Old Shore Road east
- 35.0** Bear right on Old shore Road following signs for I-95.
- 35.8** Junction with I-95. Turn left for I-95N for Rhode Island and Boston.

***END OF TRIP***



Chapter N

# Ring Dikes and Plutons: A Deeper View of Calderas as Illustrated by the White Mountain Igneous Province, New Hampshire

By John W. Creasy and G. Nelson Eby

Field Trip Guidebook for the Northeastern United States:  
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Massachusetts, Amherst, Massachusetts



CREASY AND EBY

**RING DIKES AND PLUTONS: A DEEPER VIEW OF CALDERAS  
AS ILLUSTRATED BY THE WHITE MOUNTAIN IGNEOUS PROVINCE, NEW HAMPSHIRE**

by

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**THE WHITE MOUNTAIN IGNEOUS PROVINCE**

The White Mountain igneous province (magma series of Billings, 1934) consists of plutons, ring complexes, and volcanics emplaced along a NNW trend across New England (Figure 1). Four petrographic associations are recognized (Creasy, 1974; Eby, 1987): (1) alkali syenite-quartz syenite-granite; (2) subaluminous biotite granite; (3) gabbro-diorite-monzonite and; (4) syenite-nepheline syenite. The igneous activity is largely confined to two periods, 200-165 Ma and 130-110 Ma (Eby and others, 1992). These two major periods of igneous activity are related by McHone and Butler (1984) to the opening of the North Atlantic Ocean. The reader is referred to Eby (1987) for an overview of the White Mountain igneous province.

The older White Mountain igneous province is dominated by silica-oversaturated subaluminous to peralkaline rocks of association 1, including the White Mountain batholith. Two minor nepheline-bearing intrusions occur at Red Hill, New Hampshire, and Rattlesnake Mountain, Maine. These two occurrences together with nepheline-bearing intrusions of the younger White Mountain province define a narrow zone that strikes at high angle to the NNW trend of the overall province (Figure 1; Creasy, 1989). To the north of this zone are found the large composite plutons and batholith of the older province; to the south only a few small scattered plutons of this age are present. In contrast, nearly all plutons of the younger White Mountain province are found to the south of this zone.

The Monteregian Hills and younger White Mountain igneous provinces represent the last period of igneous activity in New England (130-100 Ma). The bulk of the magmatism occurred *ca.* 125 Ma, but younger ages have been obtained for Little Rattlesnake (114 Ma, Foland and Faul, 1977) and Cuttingsville (100 Ma, Armstrong and Stump, 1971). Plutons emplaced to the west of Logan's line consist largely of mafic alkaline suites, many of which are nepheline normative. To the east of Logan's line, felsic rocks are much more important components of the intrusions and silica-undersaturated rocks are not found. Some of these younger plutons show ring-like structures (Ossipee and Pawtuckaway) while others appear to be small plugs (e.g. Little Rattlesnake, Ascutney, and Tripyramid). In most cases the most evolved rocks are syenites and quartz syenites, but biotite granites are found at Ossipee and Merrymeeting Lake.

This field excursion illustrates two contrasting examples of the White Mountain igneous province. Day 1 and Day 2 are devoted to the **White Mountain batholith** [associations 1 and 2 above] which comprises about 50% of the total areal extent of the older White Mountain igneous province. The **Mount Pawtuckaway** ring dike complex [association 3 above] of the younger White Mountain igneous province is the subject of Day 3.

**THE WHITE MOUNTAIN BATHOLITH**

**INTRODUCTION**

The White Mountain batholith (Figure 2; see also Hatch and Moench, 1984) is a composite of several overlapping centers of felsic magmatism. Individual centers are strikingly defined by composite ring dikes of porphyritic quartz syenite. Thick sections of rhyolitic crystal tuffs, breccias, and subvolcanic granite porphyry are partially circumscribed by the ring dikes. A mosaic of subalkaline to peralkaline silica-oversaturated plutons intrude these centers and provide areal continuity to the batholith. Distribution of porphyritic, miarolitic, and aplitic textures indicate that the roofs of several plutons are partially intact.

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The geology of the White Mountain batholith is described by Billings (1928), Billings and Williams (1935), Creasy (1974), Davie (1975), Eby and others (1992), Fitzgerald (1986), Henderson and others (1977), Moke (1946), Osberg and others (1978), Parnell (1975), Smith and others (1939), Wilson (1969) and, Wood (1975). Granites, quartz syenites, and syenites account for about 97% of the 1,000 km<sup>2</sup> area of the batholith; volcanic rocks of similar composition account for the remainder. Pink, medium-grained subalkaline biotite granite (the Conway Granite) and a green, medium-grained subalkaline to peralkaline amphibole granite (the Mount Osceola Granite) comprise 80% of the batholith. Medium-grained sub-alkaline to peralkaline amphibole syenites and quartz syenites are widely distributed and are similar in occurrence, texture, and mineralogy to the Mt. Osceola. Distinctive porphyritic quartz syenite occurs in ring dikes in the western (the Mount Garfield) and the eastern (the Albany) halves of the batholith. Fine-grained syenite occurs in isolated outcrops spatially associated with the ring dikes.

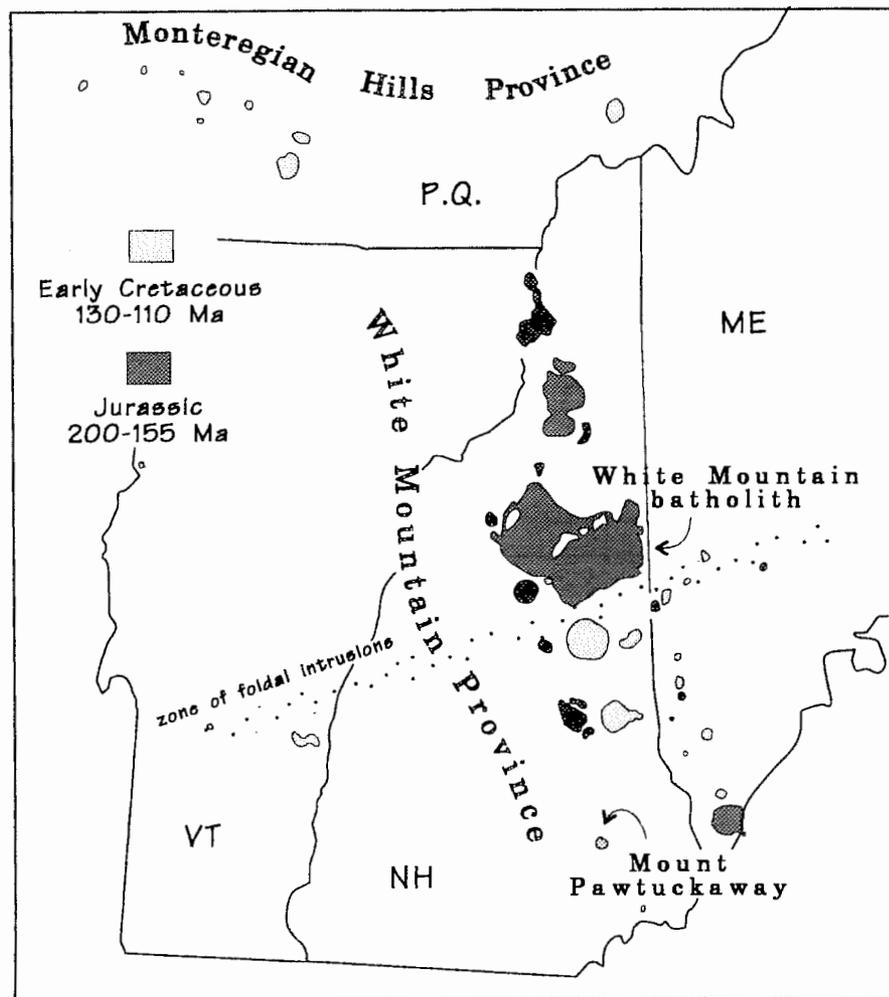


Figure 1. The White Mountain igneous province showing the location of the White Mountain batholith and Mount Pawtuckaway.

Volcanic rocks (the Moat Volcanics), chiefly trachyte, tuff, breccia, and alkali rhyolite and comendite, are found in the eastern portion of the batholith. Only minor occurrences of such lithologies are present within the ring dikes of the western batholith. Several units of granite porphyry (grouped as the Mount Lafayette unit) occurring in the western batholith differ little in texture or mineralogy from the comendites of the eastern batholith. We show these rocks as volcanics (Figure 2) although definitive volcanic textures are generally

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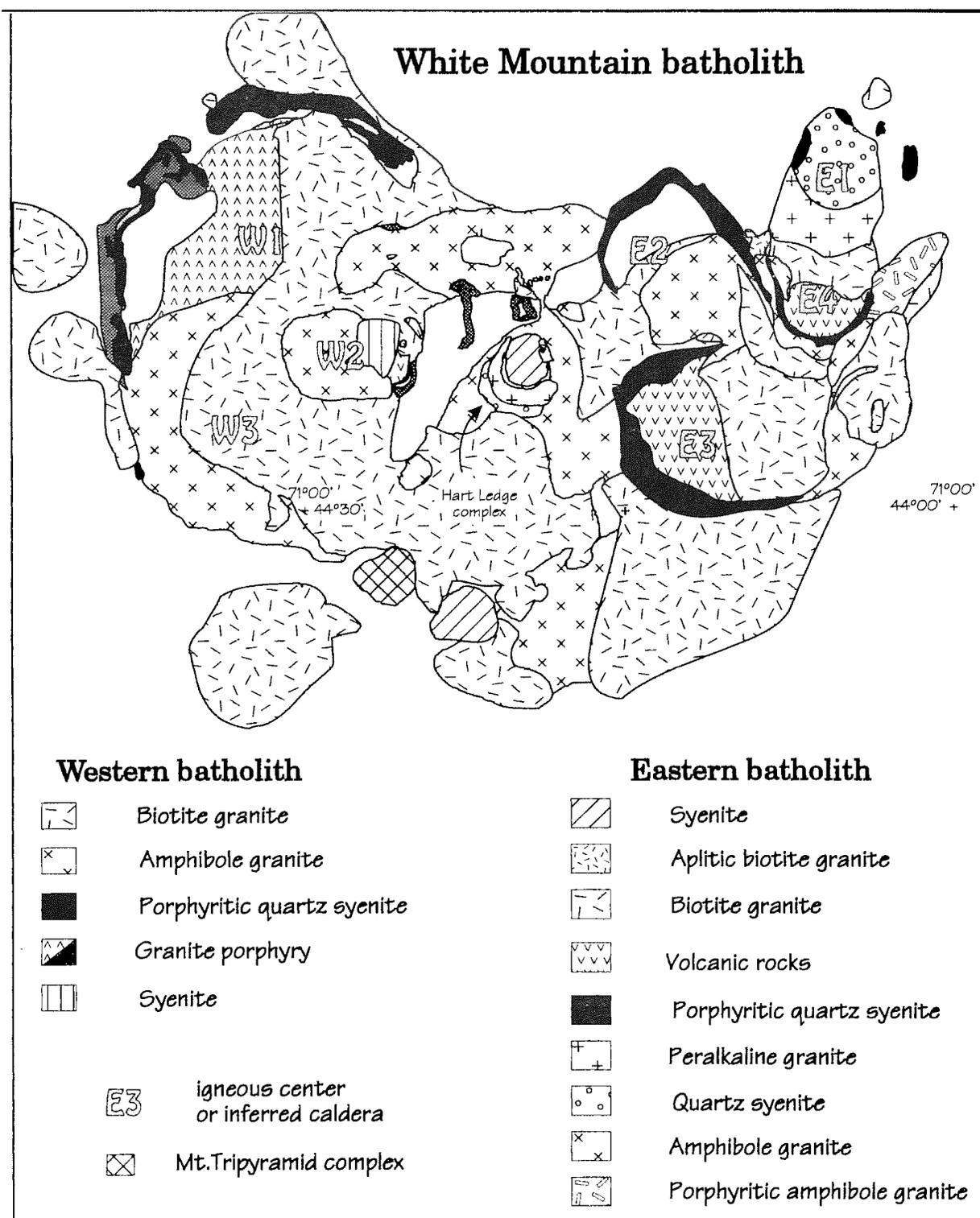


Figure 2. Geologic map of the White Mountain batholith (modified from Creasy, 1974; Osberg and others, 1976).

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lacking. Historically, the granite porphyry is treated as intrusive and not included within the Moat Volcanics (Williams and Billings, 1935; Eby and others, 1992).

Gabbro, diorite, and monzonite are present in the Mt. Tripyramid complex (Figure 2), a member of the younger White Mountain igneous province, that is spatially associated with the White Mountain batholith.

### EMPLACEMENT OF THE WHITE MOUNTAIN BATHOLITH

Emplacement of the batholith occurred in middle and late Jurassic time, 201-155 Ma (Eby and others, 1992). The western half of the White Mountain batholith, exposed in the Franconia and Crawford Notch 15' quadrangles, contains three igneous centers (W1-3, Figure 2), the largest of which is 20 km in diameter. Igneous activity commenced in the western batholith with emplacement of the porphyritic quartz syenite (201 Ma and 193 Ma) and the quartz porphyry (195 Ma) of center W1 and the syenite and trachyte of center W2 (193 Ma). Subsequent intrusions of amphibole granite (187 Ma) [possible W3 (?)] and biotite granite (181 Ma) were widespread across the entire area of the batholith. Intrusion of peralkaline granite (177 Ma) in the eastern part of the pluton is considered an extension of the amphibole granite (Mount Osceola) event.

The eastern portion of the White Mountain batholith, exposed in the North Conway and Crawford Notch 15' quadrangles, has at least four magmatic centers (Figure 3). Two centers with thick pyroclastic successions are interpreted as calderas (Noble and Billings, 1967; Fitzgerald, 1987; Fitzgerald and Creasy, 1988). Other centers where ring dikes or crescent-shaped intrusions are associated with epizonal plutons define more deeply eroded calderas. Caldera development here post-dates similar events in the western batholith by about 10-20 Ma. Dated units include the ring dike of center E2 (179 Ma); the Moat Mtn volcanic sequence (173-168 Ma) and ring dike (170 Ma) of center E3; and plutons of biotite granite (171 Ma and 155 Ma). We interpret the White Mountain batholith as a sub-horizontal slice through a caldera field cut about 1.5 km thick and 1-2 km below the original landsurface. This excursion illustrates the field characteristics and structural relations of plutons, ring dikes, and volcanics that constitute the eastern half of the White Mountain batholith (Figure 3).

### PLUTONS

Rocks forming plutons within the White Mountain batholith (Figures 2 and 3) are divided into two groups: (1) amphibole-bearing granites, quartz syenites and syenites and; (2) biotite granites.

#### **Amphibole-bearing Granites, Quartz Syenites, and Syenites**

**Mount Osceola Granite.** The Mt. Osceola Granite, a green amphibole  $\pm$  biotite granite, is the *oldest* member of the White Mountain magma series exposed in the North Conway quadrangle (Osberg and others, 1978; Eby and others, 1992). The number and original extent of plutons of the Mt. Osceola Granite within the North Conway quadrangle is not fully certain due to the complexity and abundance of younger rocks. A whole-rock Rb-Sr isochron for samples from both eastern and western portions of the batholith yields an age of 186 m.y. (Eby and others, 1992) and indicates synchronous intrusion over a broad area. [This age places the Mt. Osceola as the *youngest* member associated with the large magmatic center that forms the western portion of the batholith.]

The Mt. Osceola Granite is a medium- to coarse-grained hypersolvus granite that is dark green where fresh. It consists of an interlocking network of anhedral to subhedral micropertthite 3-10 mm in diameter enclosing rounded grains of smokey quartz. Ferrohastingsite and locally annite are interstitial, late crystallized minerals. Fayalite ( $\text{Fe}_{95.99}$ , Creasy, 1974) and sodic ferrohedenbergite (typical analysis  $\text{Na}_3\text{Ca}_{40}\text{Fe}_{56}\text{Mg}_1$ ) are frequently present in accessory amounts and encased by reaction rims of ferrohastingsite. Characteristic accessories include allanite, sphene, zircon, fluorite, ilmenite, and monazite. Locally the Mt. Osceola is weakly peralkaline with ferrichterite or riebeckite rimming ferrohastingsite. Miarolitic cavities may be locally abundant (one percent of outcrop area) and large (six to eight square centimeters). Pegmatite pods five to twenty centimeters across are

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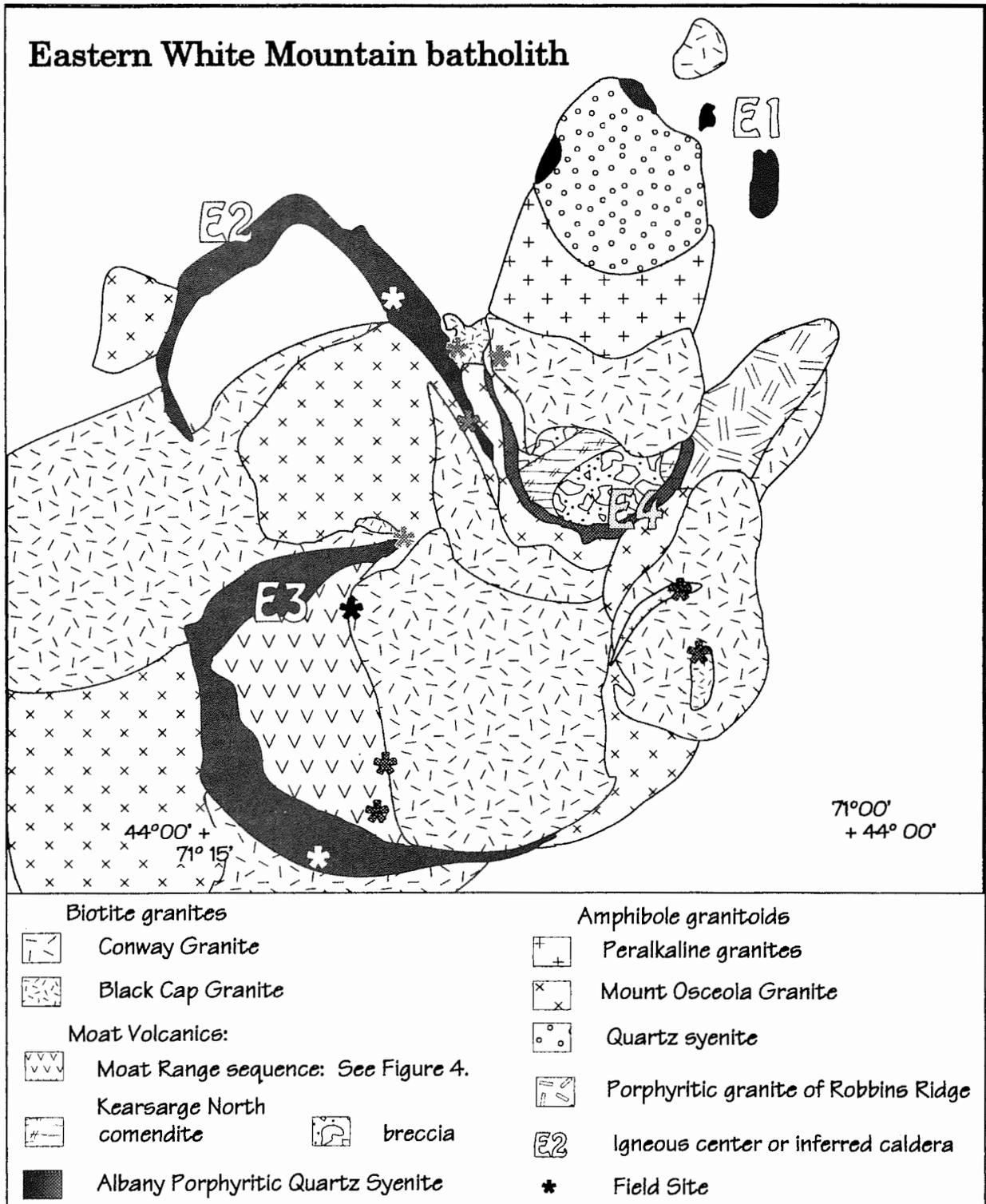


Figure 3. Geologic map of the eastern half of the White Mountain batholith (Billings, 1928, modified by Osberg and others, 1978, Creasy, 1986).

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abundant in many exposures of the Mt. Osceola Granite, locally forming up to two to three percent of the outcrop. Aplite dikes, quartz veins, and fractures are abundant in all large exposures of the Mt. Osceola Granite. The aplitic dikes rarely exceed ten centimeters in width although they may be traced continuously for a hundred meters. Veins of quartz range from two to five centimeters in width and commonly have open cores into which project well-formed crystals of quartz.

Protracted crystallization of feldspar progressively depleted the melt in Al and Ca and progressively enriched the melt in volatiles. Temperatures, total pressure, and water contents necessary for the stabilization of hydrous mafic phases were obtained after about 90% solidification. Compositions of interstitial amphiboles range from Al-poor ferrohastingsite → ferrichterite → riebeckite. The occurrence of several amphiboles within samples deemed in hand specimen to be Mt. Osceola Granite suggest that the degree of fractionation of the crystallizing magma was locally variable. A fluid phase of sufficient volume to produce deuteric alteration and form pegmatitic pods was present during the final stages of crystallization.

**Peralkaline Granite.** Riebeckite-arfvedsonite granite forms an arcuate dike and small pluton intruding the Conway Granite of the Green Hills pluton. Peralkaline granites also form larger areas of outcrop in the eastern (e.g. on North Doublehead, Parnell, 1975) and central (e.g. Hart Ledge area, Henderson and others, 1977) batholith that appear to be young plutons spatially and genetically associated with Mt. Osceola Granite. Contacts between the peralkaline granites and the Mount Osceola are commonly gradational.

The peralkaline granites are composed of subhedral grains of white micropertite (5-10 mm) and clear quartz (2-6 mm), blocky interstitial grains of riebeckite-arfvedsonite (<10 mm), and flakes and aggregates of interstitial biotite. Characteristic of this rock are abundant radiating arrays of golden colored astrophyllite. Fluorite, ilmenite, sphene, and apatite are common accessory minerals. Near contacts, miarolitic pods and cavities are developed on a cm-scale; here prismatic riebeckite crystals are found up to 5 cm in length. One small body within the Hart Ledge complex (Wood, 1975) contains ferrichterite (7%) in place of riebeckite-arfvedsonite; fayalite and ferrohedenbergite are accessories.

**Alkali feldspar Quartz Syenite.** Quartz syenite forms a small pluton within magmatic center E1 (Figure 2) and two arcuate bodies associated with the Hart Ledge complex of the central batholith (Figure 2; Henderson and others, 1977). The Hart Ledge complex is the youngest igneous activity in the central portion of the batholith, 169-162 Ma (Eby and others, 1992).

The quartz syenite is composed of tabular subhedral crystals of micropertite (1-4 mm) Anhedral quartz (< 2 mm) is interstitial to and never included within these crystals of micropertite. Rounded grains of sodic ferrohedenbergite averaging 0.5 mm are present in all specimens though in variable amounts; commonly, these grains are enclosed within the micropertite. Ferrichterite, the most abundant mafic mineral, forms interstitial grains and reaction rims on sodic ferrohedenbergite. That a vapor phase may have formed is suggested by the occurrence of riebeckite. Riebeckite forms very thin rims on ferrichterite, coats fractures within the ferrichterite and penetrates pyroxene within. Further, tufts of acicular needles of riebeckite grown on a substrate of ferrichterite project into grains of quartz. These needles commonly less than one micron in diameter seems to dictate growth from a vapor phase. Ilmenite, allanite, zircon, and sphene are common accessories.

**Alkali-feldspar Syenite.** Syenite is an uncommon plutonic member of the White Mountain magma series; only two are described from the White Mountain batholith. The syenite occurring in the central portion of the batholith (Figure 2; Wood, 1977; Henderson and others, 1977) is part of the Hart Ledge complex. The syenite is a coarse rock, dark green where fresh; blocky micropertite and ferrohastingsite (about 10%) account for ninety-five percent of the hand specimen; fayalite and ferrohedenbergite are minor accessories. The syenite contains miarolitic pods of coarse prismatic ferrohastingsite and irregularly-shaped quartz.

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The syenite is of interest because of the spatial and genetic relationship to the widespread Mt. Osceola Granite and to the peralkaline granites and quartz syenite. Significantly, REE data for the Hart Ledge complex (Creasy and others, 1979; Eby and others, 1992) show a positive europium anomaly for the syenite, but substantial negative anomalies for the quartz syenites and peralkaline granites. The syenite may represent the cumulus feldspar and the peralkaline granites may represent residual liquids derived from the crystallization of a Mount Osceola-type parental magma. No other analyzed rocks from the batholith show negative Europium anomalies.

The variation in composition of these amphiboles, consistent with principles of crystal fractionation, suggests the riebeckite granite is more strongly fractionated than the other two units. The Mt. Osceola Granite, the quartz syenite of Mt. Tremont, and the riebeckite granite, respectively, were derived from the same or similar magmas that had undergone increasingly greater degrees of crystal fractionation. The apparent variation in degree of fractionation may reflect the sequential evolution of a single parent magma or may result from the exposure, at current levels of erosion, similar magmas that had fractionated to differing degrees.

### **Biotite Granites**

**Conway Granite.** This pink sub-aluminous biotite granite is the most extensive unit in the North Conway quadrangle. Billings (1928) showed the Conway Granite as a single irregularly shaped pluton and as the youngest of the White Mountain magma series. More detailed mapping (Osberg and others, 1978) has recognized several distinct plutons of biotite granite on the basis of texture and outcrop geometry (Figure 3). Absolute ages for plutons in the North Conway quadrangle are within the range of 183-155 Ma (Eby and others, 1992). Field relations suggest that emplacement of these plutons was not synchronous across the quadrangle but related to individual magmatic centers. The Birch Hill pluton (Osberg and others, 1978) is the largest pluton. The Conway Granite of this pluton becomes finer grained, porphyritic, and miarolitic where it intrudes the Moat Volcanics. The Gardiner Brook pluton (Osberg and others, 1978) intrudes Moat Volcanics on Mount Kearsarge and is associated with the magmatic center defined by ring IV (Figure 3). The Conway Granite of this pluton shatters Silurian metasedimentary rocks (Hatch and others, 1984) along the East Branch of the Saco River. A third pluton underlies most the Green Hills, the prominent north-south oriented ridge forming the east side of Mt. Washington Valley; this pluton is well exposed on Black Cap mountain.

The Conway Granite is a medium- to coarse-grained pink biotite two-feldspar granite. Values of microperthite:oligoclase range from 2:1 to 10:1 and average 4-5:1 (Creasy, 1974). Annite (Ann<sub>90</sub>) forms anhedral interstitial grains up to 5 mm in size. In contrast with other members of the White Mountain magma series, fayalite and ferrohedenbergite are absent. Subordinate amphibole is present in some samples. Zircon, allanite, apatite, sphene, and fluorite are common accessories. Near intrusive contacts, the Conway Granite shows a variety of textures that may grade into each other on the outcrop scale: porphyritic, aplitic, miarolitic, and pegmatitic. Miarolitic cavities are typically of mm-scale and bounded by euhedral crystals of quartz and feldspar. A zone of miarolitic cavities ranging up to several meters is present within the Conway Granite adjacent to the Moat volcanics on the east side of the Moat Range. This and similar occurrences of miarolitic cavities in the eastern batholith have produced many beautiful smokey quartz crystals. Weakly developed banding on the cm- to dm-scale resulting from variations in grain size and/or mineral concentrations is developed near some contacts. Lithic fragments of any type are sparse in the Conway Granite.

**Black Cap Granite.** The Black Cap Granite (Billings, 1928) is a fine-grained pink aplitic biotite granite that outcrops in two small areas in the North Conway quadrangle. It is composed of quartz, microperthite, subordinate oligoclase, and chloritized biotite. Accessories include zircon, magnetite, apatite, and fluorite. The Black Cap Granite is shattered and intruded by the Conway Granite (Green Hills pluton) on the flanks of Black Cap. Billings considered this rock an early lithologically distinct 'phase' of the Conway Granite. Osberg and others (1978) suggest that the Black Cap granite to be coeval with and a roof facies of the Conway Granite.

## CREASY AND EBY

Table 1. Modes from the eastern portion of the White Mountain batholith (Billings, 1928; Osberg and others, 1978; and Davie, 1975).

<u>Stop No.</u>	#1a	#1b	#2a	#2b	#2c
Quartz	35	22	10	16	20
Alkali feldspar	49	72	79	70	68
Plagioclase	14	3	0	0	0
Biotite	2	tr	1	0	tr
Amphibole	0	1	8	9	
Ferrohedenbergite	0	tr	8	2	1
Fayalite	0	2	1	tr	tr
Accessories	tr	tr	3	3	3

<u>Stop No.</u>	#5	#6	#7	#8	#9
Quartz	13	29	33	39	28
Alkali feldspar	74	48	49	54	47
Plagioclase	0	17	16	0	18
Biotite	1	6	2	1	7
Amphibole	12	0	0	5	0
Ferrohedenbergite	0	0	0	0	0
Fayalite	0	0	0	0	0
Accessories	tr	tr	tr	1	1

- 1a Conway Granite, Birch Hill pluton, Hurricane Mtn Road.  
 1b Mt. Osceola Granite, Rattlesnake Mtn, Redstone area.  
 2a,b,c Albany Porphyritic Quartz Syenite, three distinct types present within ring dike E3, Little Attitash Mtn.  
 5 Albany Porphyritic Quartz Syenite, Jackson Falls.  
 6 Black Cap Granite, Thorn Mtn.  
 7 Conway Granite, Gardiner Brook pluton, Burnt Knoll Brook.  
 8 riebeckite granite, North Doublehead.  
 9 Conway Granite, Green Hills pluton, Black Cap mountain.

## RING DIKES

Ring dikes of the eastern White Mountain batholith consist of porphyritic quartz syenite and subordinate porphyritic syenite. The porphyritic quartz syenite is similar in appearance across the batholith but occurrences in the eastern and western parts of the batholith are distinguished as the Albany and the Mount Garfield, respectively. Ring dikes define at least four magmatic centers in the eastern batholith (Figure 3). Ring dikes of centers E2 and E3 yield Rb/Sr ages of 179 and 170 Ma, respectively (Eby and others, 1992). These ring dikes are outwardly dipping at 40°- 80° with well developed chill margins adjacent to older rocks. The ring dikes are not seen to cut each other but relationships to other units indicate their emplacement was not simultaneous. Although similar in mineralogy, the ring dikes of different centers are distinctive in mineral chemistry and in texture.

## CREASY AND EBV

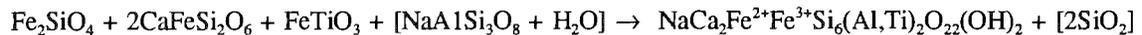
In fine structure individual ring dikes are themselves multiple intrusions. For example, at least four separate intrusions of porphyritic quartz syenite and syenite occur in ring E3 (Table 1; Figure 3). The increasing abundance of feldspar phenocrysts and of total quartz content with decreasing age suggests (Davie, 1975) successive differentiates of a subjacent magma body. Inclusions present in ring E2 document a similar structural relationship.

**Albany Porphyritic Quartz Syenite**

The Albany Porphyritic Quartz Syenite contains phenocrysts of microperthite (5-10 mm) and subordinate quartz (2-4 mm); minor phenocryst phases include ferrohedenbergite ( $\text{Ca}_{45}\text{Fe}_{45-40}\text{Mg}_{10-15}$ ), fayalite ( $\text{Fa}_{90-94}$ ) and ilmenite (Creasy, 1974). Variation of total phenocryst abundance (10-60%) and phenocrystic feldspar:quartz (5-10:1) is noted both within a ring dike (e.g. #2a, b, c of Table 1; Davie, 1975) and among different ring dikes. Quartz phenocrysts are subangular in chilled border zones but are rounded with seriate margins in coarser varieties.

The groundmass is uniform in grain size (<2-3 mm) within an intrusion (except near contacts) but shows variation among different intrusions. Minerals of the groundmass are anhedral quartz, alkali feldspar, subordinate oligoclase, and minor ferrohastingsite and annite. Ferrohastingsite occurs as anhedral interstitial grains and as rims on the ferrohedenbergite and fayalite. In both occurrences it may poikilitically enclose small quartz grains. Grunerite is found as reaction rims on fayalite and is surrounded by ferrohastingsite. Accessories include allanite, sphene, zircon, and fluorite. Secondary sericite, biotite and chlorite are commonly present.

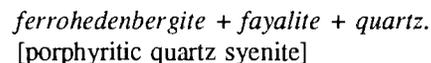
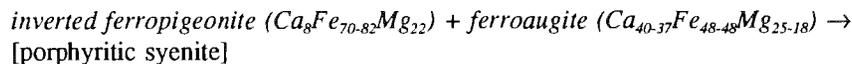
The mineralogic transition from the anhydrous phenocryst (intratelluric) assemblage to the hydrous groundmass (emplacement) assemblage is written as a simplified end-member reaction (Creasy, 1974):



where [ ] indicates probable melt components. The stabilization of the hydrous phase, ferrohastingsite, relative to fayalite and ferrohedenbergite requires a reduction in temperature or a combined reduction in temperature and total pressure. The textural and physicochemical changes coincide with emplacement of the ring dikes.

**Porphyritic Syenite.**

The porphyritic syenite is the oldest and least evolved lithology present within ring dikes. In contrast to the porphyritic quartz syenite, total phenocryst abundance is commonly less than fifteen percent, reaction rims of ferrohastingsite are minor or lacking, and alkali feldspar lacks exsolution textures. A two-pyroxene assemblage proxies for the ferrohedenbergite typical of the Albany:



This reaction marks the boundary of the pyroxene 'forbidden zone' and bulk grain compositions cited above yield  $T = 825^\circ\text{C}$  (Lindsley, 1983). Compositions of orthopyroxene host and ferroaugite lamellae of the inverted ferropigeonite give temperatures of about  $700-750^\circ\text{C}$ .

## CREASY AND EBY

## VOLCANIC ROCKS

The Moat Volcanics (Billings, 1928) are exposed in the Moat Range to the west of North Conway (E3, Figure 3; Figure 4); a second major occurrence is on Kearsarge North (E4, Figure 3). The Moat Range sequence shows pronounced and laterally persistent layering (Billings, 1928; Noble and Billings, 1967) striking northwest and dipping 30°-40° to the northeast. Billings (1928) interpreted these exposures as an originally flat-lying sequence that subsided and rotated along a ring fracture (now the ring dike). Noble and Billings (1967) suggest syn- or post-subsidence accumulation of the sequence within a caldera. An intra-caldera setting is supported by the detailed mapping of Fitzgerald (1987; Figure 4). The original extent of the volcanics outside the caldera is a major question. The isotopic systematics (Eby and others, 1992) indicate that fractional crystallization of quartz syenitic magma with minor amounts of crustal contamination can produce the comendites of the Moat Range sequence.

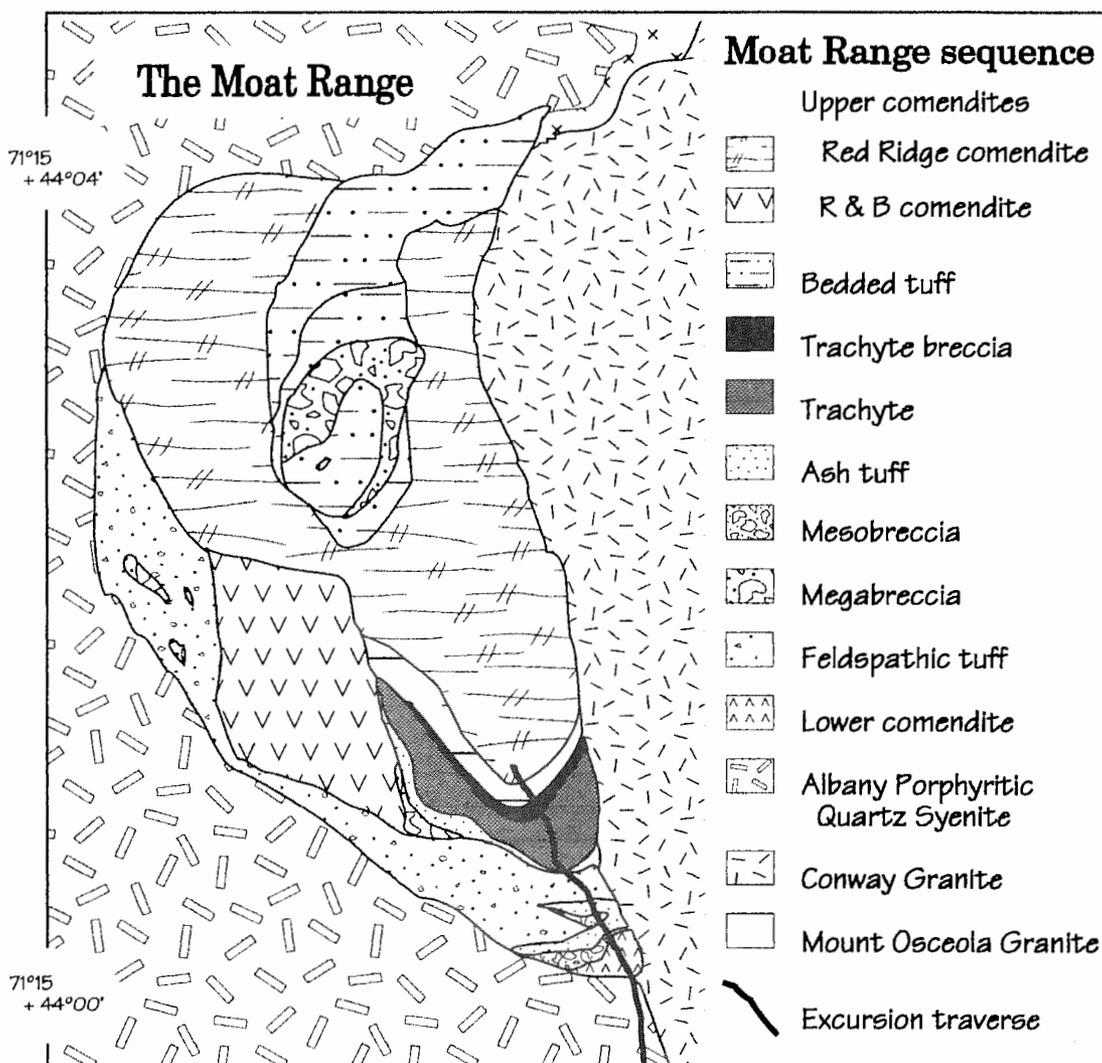


Figure 4. Geologic map of the Moat Range (Billings, 1928; Fitzgerald, 1987).

Five crystal-rich units dominate the 3.6 km of volcanic stratigraphy of the southern Moat Range (Fitzgerald and Creasy, 1988; Figure 4). These are (base to top): lower comendite (380 m), feldspathic welded tuff (425 m), the R & B comendite (1070 m), trachyte and trachyte breccia (300 m), and Red Ridge comendite (>1050 m).

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The lower comendite contains interbeds of polymict lapilli tuff and breccia that are thickest and coarsest adjacent to the enclosing ring dike. Internal structures (fiamme, oriented lithic clasts) developed within the lower two units but are lacking in the upper comendites. Several thin but persistent horizons of bedded ash tuff, lapilli tuff, and mesobreccia and megabreccia lie above and below the trachyte. Bedding generally parallels the surrounding ring dike (E3) and dips radially inward at 20°-40°.

The northern Moat Range is dominated by the upper comendite units which are generally homogeneous but show variation of abundance and proportion of phenocrysts and lithic fragments. Bedded tuffs here are quite similar to the comendites but exhibit bedding (averaging 2 m), eutaxitic textures, and are more lithic-rich. A section of chaotic megabreccia (60 m thick, clasts up to 4 m) within the bedded tuffs is interpreted as a lahar or debris flow (Fitzgerald, 1987). A polymict clast-supported mesobreccia is present on the summit of North Moat Mtn (3196'), highest point of the Moat Range.

Kearsarge North (E4, Figure 3) exposes a mix of comendite, feldspathic tuff, and meso- and megabreccias (Billings, 1928) similar to those found on North Moat. The comendites lie topographically below the more extensive breccias and are lithic-rich. As in the Moat Range sequence, the breccias on Kearsarge North are polymict (although in any horizon one or another clast lithology may predominate) and the matrix:clast ratio varies greatly.

The comendites are blue-gray to pink rocks contain variably abundant phenocrysts (1-3 mm) of quartz and microperthite and rare phenocrysts of biotite, ferrohastingsite, ferrohedenbergite, and riebeckite set in a matrix of quartz and alkali feldspar. Accessories include apatite, fluorite, zircon, and magnetite. Lithic fragments (mm- and cm-scale) constitute 1-5% of most comendite samples. Lithic types include hornfels, cogenetic volcanic rocks, porphyritic quartz syenite, and rarely cogenetic plutonic rocks. The trachyte consists of phenocrysts (2-3 mm) of pink alkali feldspar set in a dense (<0.1 mm) groundmass of alkali feldspar. Accessories include abundant hematite and minor zircon, magnetite, epidote and clinozoisite. The breccias contains angular to subrounded blocks ranging from a few centimeters to a meter in size. Lithic fragments include a variety of metamorphic rock lithologies, Paleozoic intrusive rocks, and cogenetic volcanic and hypabyssal rocks.

#### ITINERARY FOR DAY 1 AND DAY 2

All stops are located in the U.S.G.S. North Conway and Crawford Notch 15' Quadrangles. Access to Stops 2 and 3 is through the Ossipee Lake 15' Quadrangle. All excursion stops are located on the U.S.G.S. 7.5' topographic base in Figures 5 - 11.

**STOP 1A. CONWAY GRANITE AT THE REDSTONE PINK QUARRY (45 MINUTES)** (North Conway West Quadrangle). The coarse pink Conway Granite (Table 1, #1a) of the Birch Hill pluton is homogeneous in grain size and texture and was quarried extensively for building stone (Dale, 1923). Sparse inclusions, chiefly of porphyritic granite, are present locally. A few thin (3 cm) dikes of aplitic granite are seen on the main bench of the quarry. Samples from here have been used in numerous studies related to the high concentrations of U (20 ppm) and Th (70 ppm) in the Conway Granite (e.g. Adams and others, 1962; Birch and others, 1968; Osberg and others, 1978).

Exit the clearing adjacent to columnar pieces of granite. Follow path .1 mi to small quarry with derricks.

**STOP 1B. MT. OSCEOLA GRANITE AT THE REDSTONE GREEN QUARRY (30 MINUTES)** (North Conway West Quadrangle). The Mt. Osceola Granite (Table 1, #1b) was quarried here. The contact between this green ferrohastingsite granite and the Conway Granite of Stop 1a is reached by traversing around the east and north (top) sides of the quarry along a narrow foot trail. The Conway Granite is distinctly finer grained at the contact.

## CREASY AND EBY

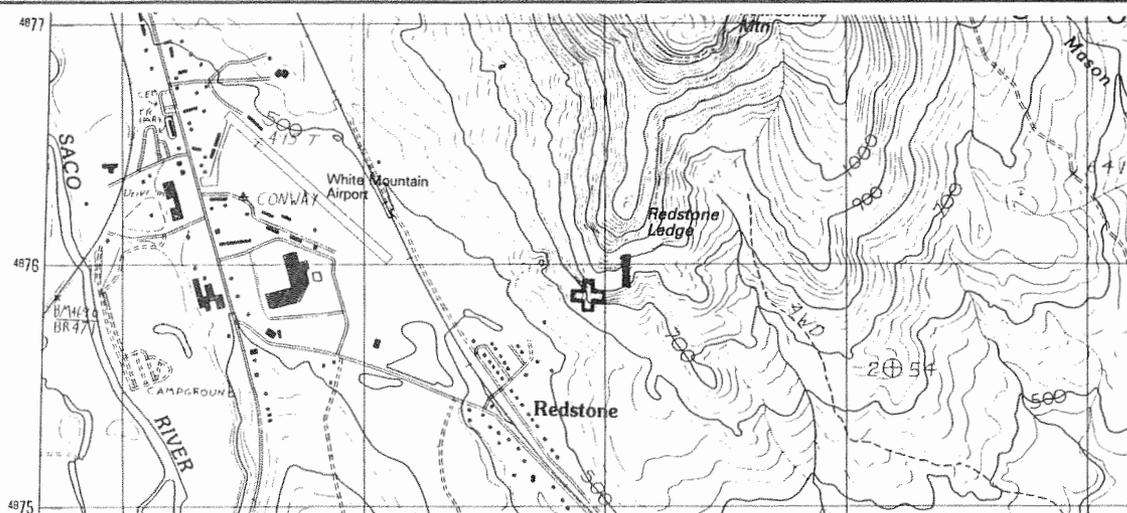


Figure 5. Location of STOP 1 (North Conway West Quadrangle). Grid spacing is 1 km.

**STOPS 2A, B, C. ALBANY PORPHYRITIC QUARTZ SYENITE OF RING DIKE E3 (1 HOUR)** (Silver Lake Quadrangle). The Kancamagus Highway (and adjacent Swift River) crosses ring dike E3 (Figure 3) at a low angle providing excellent outcrops of the Albany Porphyritic Quartz Syenite and the Porphyritic Syenite within the composite ring dike E3. This stop illustrates at least three Albany-type lithologies (Table 1, #2a, b, c) distinguished by the abundance and proportion of quartz and feldspar phenocrysts. Davie (1975) demonstrated the composite nature of this ring dike on the north margin (Attitash Ski Area, North Conway West Quadrangle) which provides excellent but less accessible exposures.

**STOP 3. ALBANY PORPHYRITIC QUARTZ SYENITE AT TYPE LOCALITY (20 MINUTES)** (North Conway West Quadrangle). This is the probable type locality for the Albany porphyritic quartz syenite (Hitchcock, 1877) and a calendar-quality photo op of the Albany covered bridge.

**STOP 4. TRAVERSE OF MOAT RANGE VOLCANIC SEQUENCE, SOUTH MOAT MTN (4 HOURS)** (North Conway West Quadrangle). The purpose of Stop 4 is to illustrate the varied lithologies of the Moat Range sequence. This involves a traverse of about 3.5 miles and 1200 vertical feet (up going, down returning) along and adjacent to the South Moat Trail. The traverse is shown on Figure 4.

**STOP 5. U.S.F.S. SMOKEY QUARTZ COLLECTING AREA (30 MINUTES)** (North Conway West Quadrangle). This is an accumulation of boulders of the miarolitic contact facies of the Conway Granite; bedrock is not exposed.

**STOP 6. CATHEDRAL LEDGE (20 MINUTES)** (North Conway West Quadrangle). The Conway Granite of the Birch Hill pluton is beautifully exposed on the summit of Cathedral Ledge (New Hampshire's Half Dome!). The excellent views of the Saco River and Mt. Washington Valley make this a convenient location for geologic orientation.

**STOP 7. CONTACT OF MOAT VOLCANICS AND CONWAY GRANITE IN LUCY BROOK (2 HOURS)** (North Conway West Quadrangle). At Diana's Baths (0.4 mi) the coarse Conway Granite becomes porphyritic and contains large (up to 3 m) segregations of dark material that Billings (1928) interpreted as partially assimilated inclusions. NO HAMMERS PLEASE! From here, follow the Moat Mtn Trail (1.4 mi) to the first trailside ledges (Moat Volcanics), then cut over to Lucy Brook (100 ft). The contact between the Moat Volcanics (upstream) and the Conway Granite (downstream) is very well exposed in the Brook. In the lowest stream exposures, the coarse Conway Granite grades into medium-grained porphyritic contact facies characterized

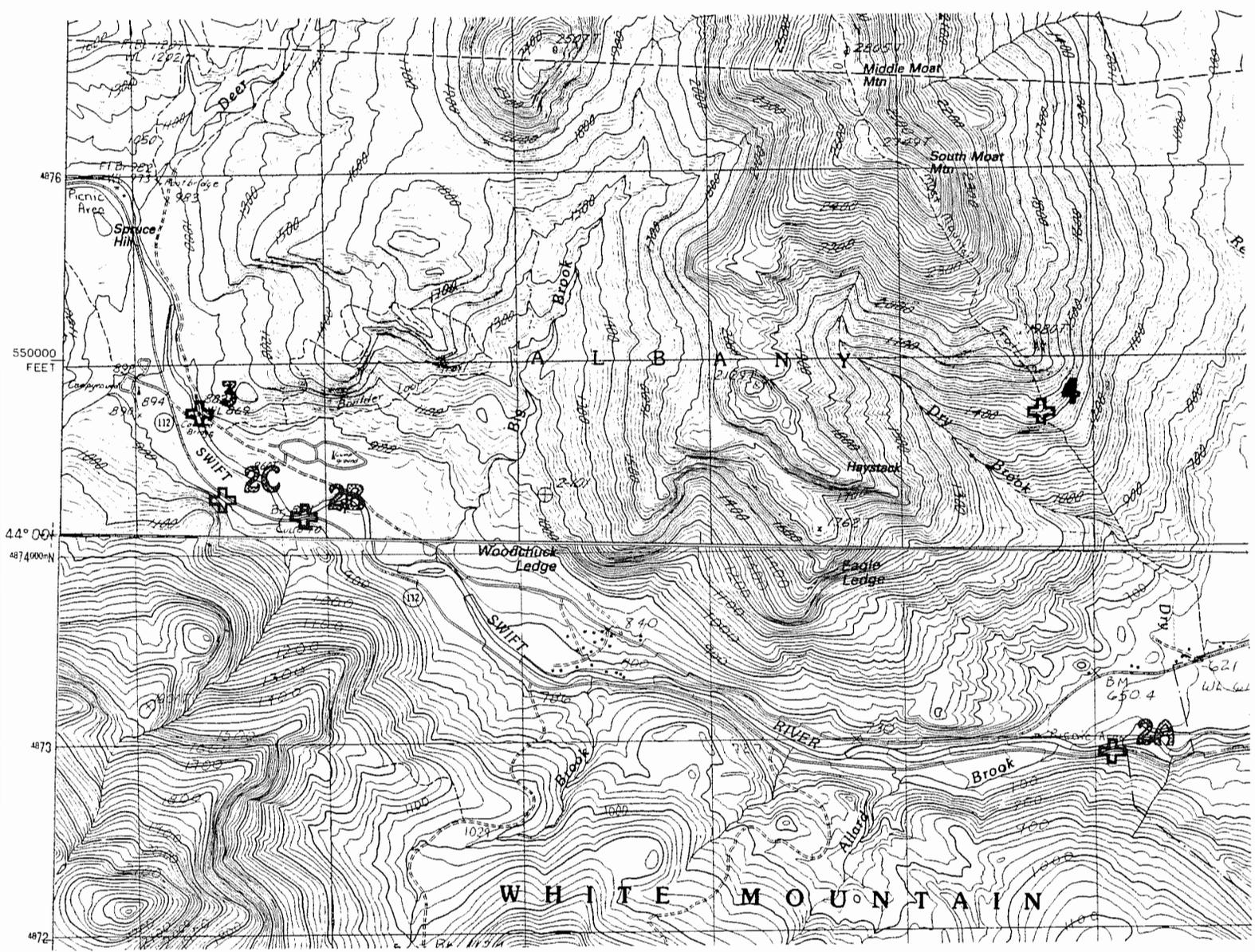


Figure 6. Location of STOPS 2, 3, and 4 (North Conway West Quadrangle). Grid spacing is 1 km.

## CREASY AND EBY

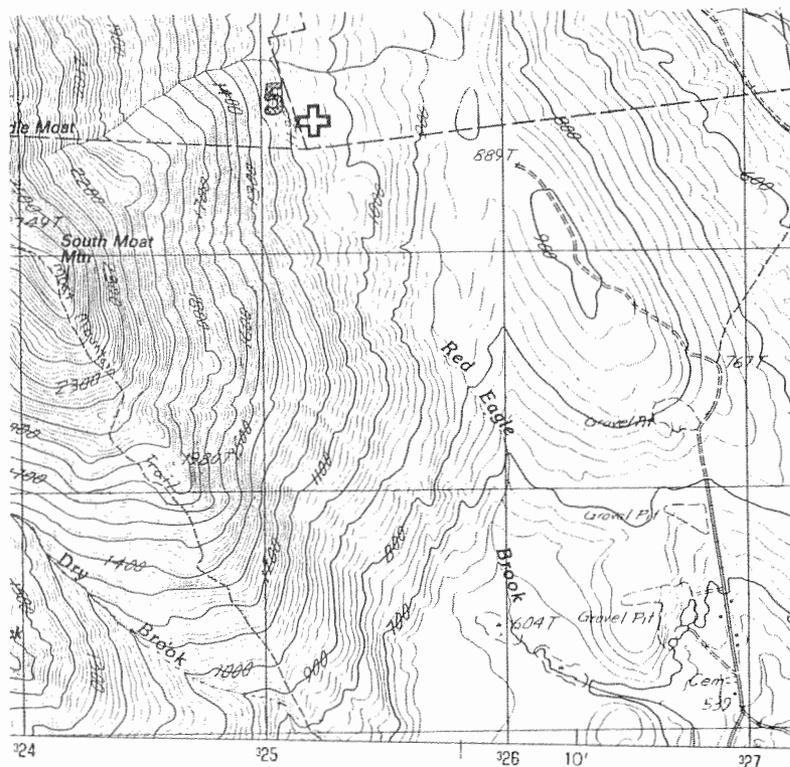


Figure 7. Location of STOP 5 (North Conway West Quadrangle). Grid spacing is 1 km.

by a heterogeneous and miarolitic texture. Large rounded inclusions of fine-grained biotite granite are included in the Conway Granite, perhaps representing an older contact facies. Blue-gray comendite of the Moat Volcanics is bleached and pink where cut by numerous fractures and thin (<2 cm) quartz veins. Eutaxitic texture is not well developed here; the rocks resemble quartz porphyry.

**STOP 8. MOUNT OSCEOLA GRANITE AT HUMPHREY'S LEDGE (20 MINUTES)** (North Conway West Quadrangle). The Mount Osceola Granite is exposed at the north end of Humphrey's Ledge. This is an excellent locality for collecting samples that contain fayalite and ferrohedenbergite.

**STOP 9. ALBANY PORPHYRITIC QUARTZ SYENITE OF RING DIKE E2 (1 HOUR)** (Jackson Quadrangle). Park on wide right (east) shoulder at Jackson Falls picnic area. Walk to broad exposures in Wildcat Brook. NO HAMMERS PLEASE! The Albany Porphyritic Quartz Syenite in composite ring dike E3. This is one of the best (and most scenic) localities to examine the Albany--about 2000 ft of continuous exposure is present between here and Jackson along Wildcat Brook. Just downstream of the iron bridge a screen of Siluro-Devonian gneisses and granite 110 ft wide is intruded by the Albany. Downstream from this screen, the Albany is relatively uniform in mineralogy and texture and contains a few small (2-5 cm) inclusions. Upstream, large (up to 1 m) inclusions of feldspar-poorer porphyritic syenite are enclosed by the Albany.

**STOP 10. INTRUSIVE RELATIONS AT THORN MTN (1 HOUR)** (Jackson Quadrangle). A 1 mi loop at traverses ski trails to and from the summit of Thorn Mtn. The ski trails on the north slope of Thorn Mtn expose pavement outcrops of the fine-grained and homogeneous Black Cap biotite granite (Table 1, #6). Just below (north of) the summit, the Black Cap intrudes a screen of country rocks consisting of Silurian meta-

## CREASY AND EBY

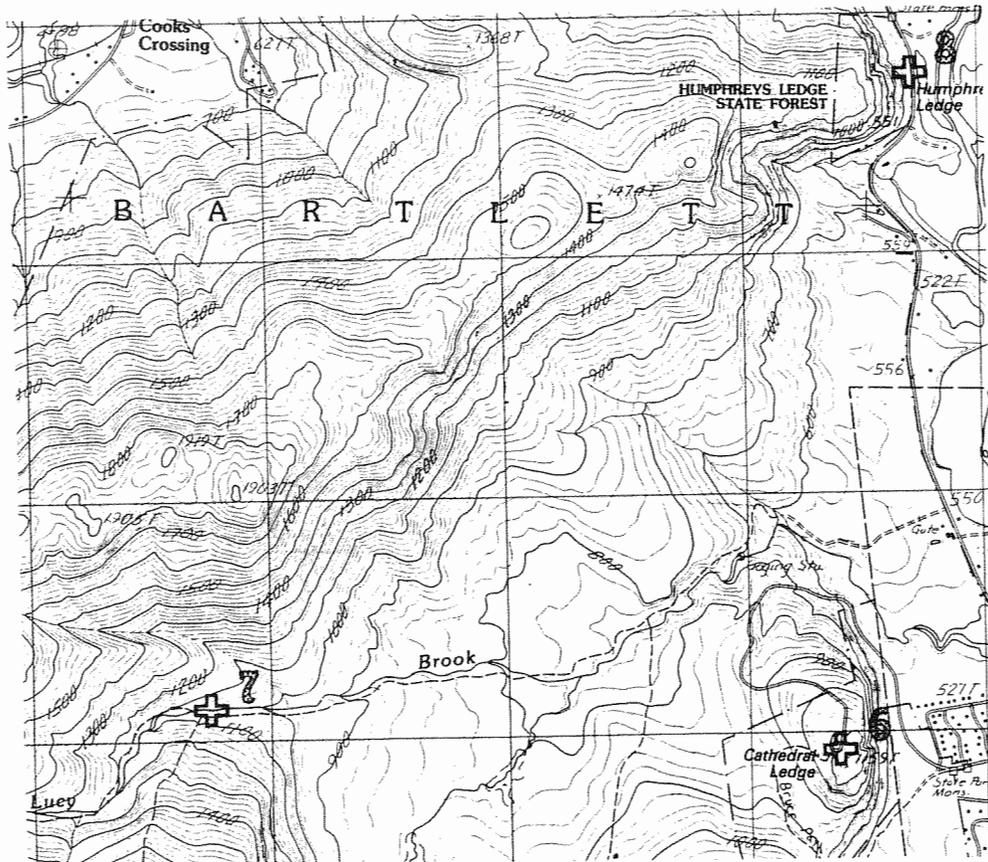


Figure 8. Location of STOPS 6, 7, and 8 (North Conway West Quadrangle). Grid spacing is 1 km.

morphic rocks and Devonian (?) granites and pegmatites. The south margin of the screen is shattered and intruded by a chill facies of Albany Porphyritic Quartz Syenite. At and south of the summit the Albany Porphyritic Quartz Syenite has coarser grained groundmass and may be a separate intrusion. The screen is cut out to the east; the Black Cap Granite is in contact with the chilled Albany but age relations are not evident here.

**STOP 11A. INCLUSIONS WITHIN RING DIKE E4 (20 MINUTES)** (North Conway West Quadrangle). The purpose of Stop 11 in total is to examine the internal and marginal aspects of the Albany Porphyritic Quartz Syenite within ring dike E4 exposed within the East Branch of the Saco River. Stop 11A is within the interior of the ring dike. Notable here are abundant small inclusions. The average azimuth of inclusions does not parallel the northwesterly trending contact of the Albany with the younger Conway Granite. They may parallel the original contact of the Albany with the country rock; subsequent intrusion of the Conway Granite resulted in a different orientation of the contact.

**STOP 11B. CHILL MARGIN OF RING DIKE E4 (1 HOUR)** (North Conway West Quadrangle). The Albany is here chilled against a screen of schists, grading over a distance of twenty feet from typical Albany to a dense brown porphyry through a reduction in grain size of the groundmass. The percentage of quartz phenocrysts increases (from 5 TO 15%) and their average size decreases towards the margin.

## CREASY AND EBY

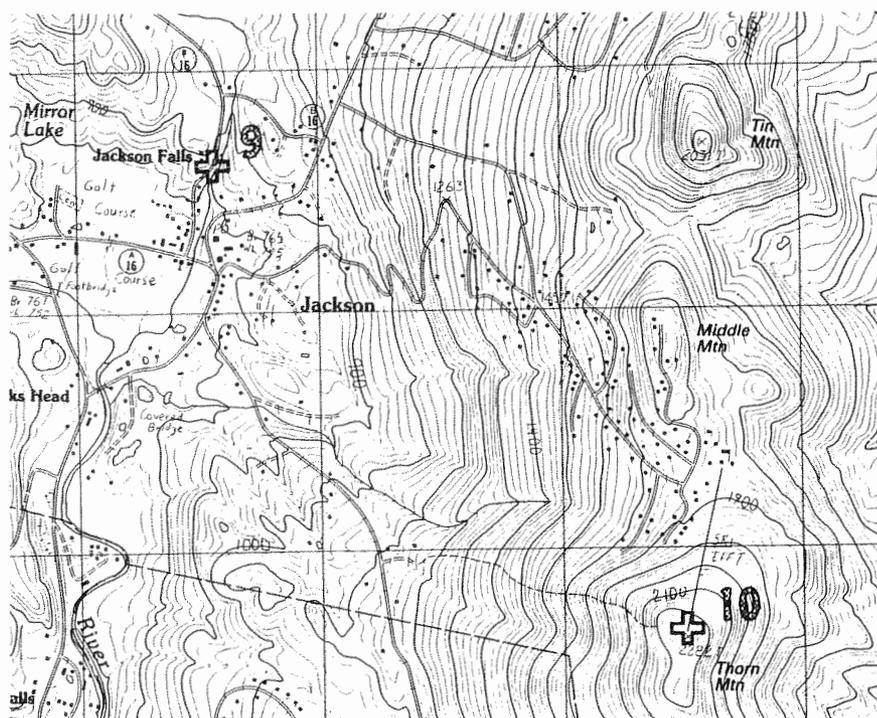


Figure 9. Location of STOPS 9 and 10 (Jackson Quadrangle). Grid spacing is 1 km.

**STOP 12. CONTACT OF THE GARDINER BROOK PLUTON. (30 MINUTES)** (North Conway West Quadrangle). Park on south side of bridge and cross bridge to outcrop on northeast side of road. The coarse-grained Conway Granite of the Gardiner Brook pluton (Table 1, #7) exposed at the north end of the long outcrop becomes porphyritic towards the Silurian schists and gneisses which occur downstream of the bridge (and to Stop 11B!). These country rocks are shattered and intruded by a fine-grained biotite granite (Black Cap Granite ?) which is intruded by the porphyritic Conway Granite.

**STOP 13. PERALKALINE GRANITE ON HURRICANE MTN (1 HOUR)** (North Conway East Quadrangle). Follow woods road to a point where a trail leaves left (about .25 mi); ascend via trail to open south-facing ledges (.15 mi) that expose riebeckite-arfvedsonite granite. Conway Granite is exposed around the rest of the mountain. Abundant miarolitic pods and cavities notable for their large prismatic crystals of riebeckite-arfvedsonite--check blast debris! This miarolitic contact facies is probably the top or margin of a larger subjacent pluton of peralkaline granite (c.f. South Doublehead Mtn). Exposures near the summit of Hurricane Mtn lack biotite and contain only scattered miarolitic pods and may lie below this contact zone. Astrophyllite is a characteristic accessory mineral in this granite.

**STOP 14: BLACK CAP GRANITE AND GREEN HILLS PLUTON (1.5 HOURS)** (North Conway East Quadrangle). The purpose of this traverse (about 2 miles, 600' of relief--up going, down returning) is to examine the relationship between the Black Cap and Conway granites. The medium-grained Conway Granite of the Green Hills pluton (Table 1, #9) is well exposed in the ledges as the trail ascends the north slope of Black Cap from the Hurricane Mtn road. Near and south of the summit, the Conway locally becomes sub-porphyritic. About .25 mi southeast of the summit (no trail) the exposure geometry of the fine-grained Black Cap Granite suggests it forms a sub-horizontal sheet--probably the roof facies of the pluton. The Black Cap Granite is shattered and intruded by the main intrusive pulse of the Conway Granite.

CREASY AND EBY

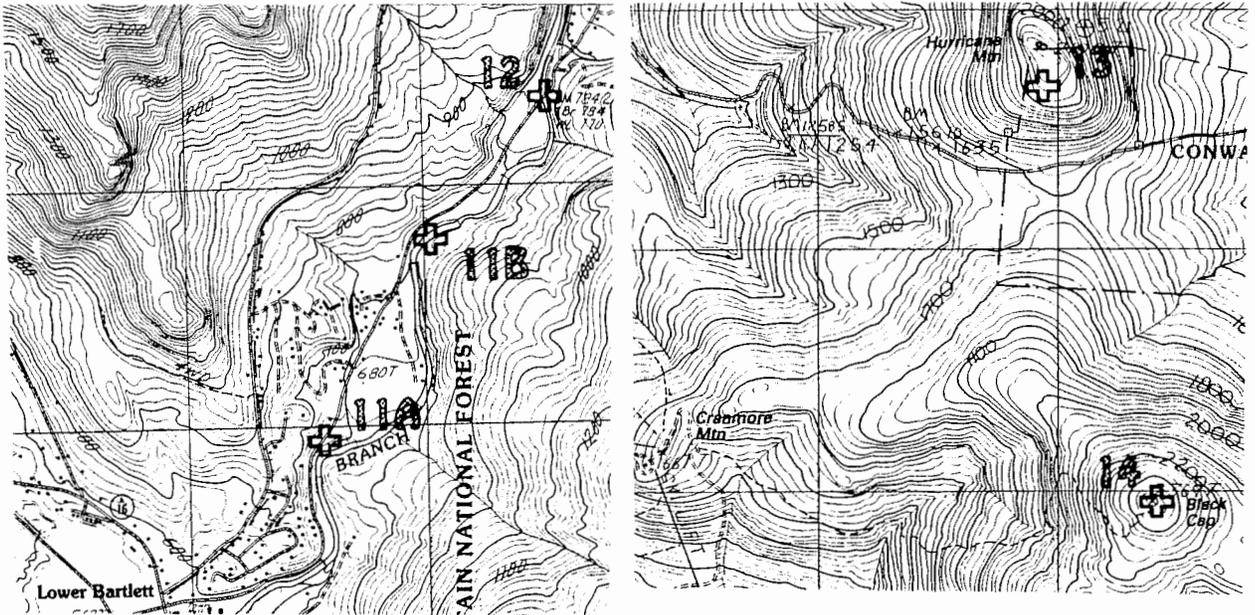


Figure 10. (Left) Location of STOPS 11 and 12 (North Conway West Quadrangle). (Right) Location of STOPS 13 and 14. Grid spacing is 1 km.

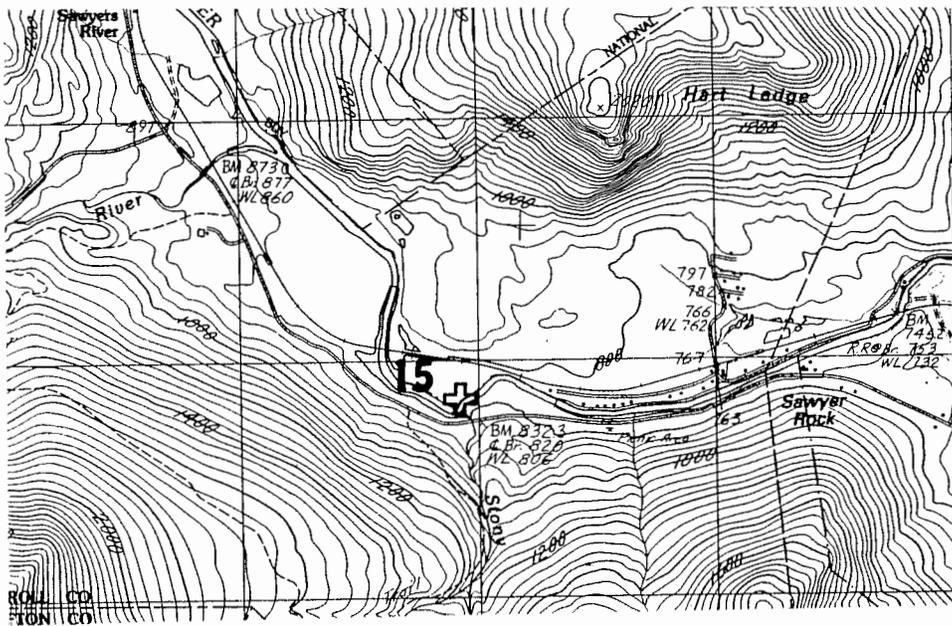


Figure 11. Location of STOP 15 (Bartlett Quadrangle. Grid spacing is 1 km.

## CREASY AND EBY

### STOP 15: SYENITE OF THE HART LEDGE COMPLEX. (30 MINUTES) (Bartlett Quadrangle).

Coarse syenite of the Hart Ledge body is found at the juncture of Stony Brook and the Saco River. Samples from this locality have a negative Europium anomaly (Creasy and others, 1979), most unusual among felsic rocks of the White Mountain batholith. The purpose of this stop is to evaluate possible field evidence for a cumulus origin. A vague cm-scale mineral layering is present but is it convincing? The mirolitic cavities or pegmatitic clots dimpling the surface of this outcrop contain large crystals of ferrohastingsite and quartz. Abundant dikes of syenite pegmatite and aplite, many of them compound, are well exposed.

## MOUNT PAWTUCKAWAY

### INTRODUCTION

Mount Pawtuckaway, a member of the younger White Mountain igneous province, is located in Rockingham County, New Hampshire. The complex is centrally located on the 15' Mt. Pawtuckaway quadrangle and falls within the boundaries of Pawtuckaway State Park. The complex has a surface exposure of approximately 8 km<sup>2</sup>, is roughly circular in outline, and the maximum relief is on the order of 200 m. The mafic rocks, which are easily eroded, underlie the lowlands while the more resistant monzonites and syenites form ridges. The low ridge along the southern edge of the complex is underlain by gabbros and troctolites. The Pawtuckaway magmas were intruded into the Precambrian Massabesic Gneiss Complex.

The earliest studies of the Mount Pawtuckaway complex described the rocks as largely syenites and camptonites. Roy and Freedman (1944) were the first to completely map the complex and a further modification of the geology is found in Freedman (1950). Shearer (1976) did a geochemical study of the major units and Richards (1990) undertook a structural, petrographic and geophysical study. A series of senior projects (J. Dadoly, M. Kick, M. Lambert, and J. Plunkett) conducted at the University of Massachusetts, Lowell, have dealt with various aspects of the geology and geochemistry of the complex. The data of all these investigators has been used to construct the current version of the Mount Pawtuckaway geologic map (Figure 5).

### GENERAL GEOLOGY

Modeling of geophysical data (Richards, 1990) indicates that the pluton is a plug-like structure extending to a depth of approximately 3 km. The units have steep contacts, and a body of high magnetic susceptibility occurs at depth. Field relations indicate that the mafic rocks were emplaced prior to the felsic rocks (see Figure 5 for the locations of the various units). Based on apatite fission-track data Doherty and Lyons (1980) estimated that the rocks currently exposed at the surface were originally at a depth 3.0 to 3.6 km. A K-Ar biotite age of  $124 \pm 2$  Ma (corrected to new decay constants) has been determined for the coarse-grained monzonite (Foland and others, 1971).

The earliest mafic rocks are pyroxenites which are preserved as blocks in the foliated diorite and medium-grained monzonite. Given the large size of some of these blocks, it is unlikely that they have undergone any significant upward transport, and they most likely occur close to their original level of emplacement. An arcuate body of gabbro occurs along the southern margin of the complex. This unit is distinguished by the anorthite content of its plagioclases ( $An_{60}$  to  $An_{46}$ ), the essential absence of apatite (which is a common accessory in the other mafic units), and its distinctive trace element geochemistry. Troctolites occur within this unit. Locally the gabbro does show a foliation dipping inward at about 50°. The position of the gabbro in the sequence of mafic rock emplacement is ambiguous since it is not intruded by any of the other units. Several varieties of diorite are found: coarse-grained hornblende diorite, medium-grained foliated diorite and fine-grained diorite. The coarse-grained hornblende diorite is largely confined to the western portion of the complex. Locally this unit does show

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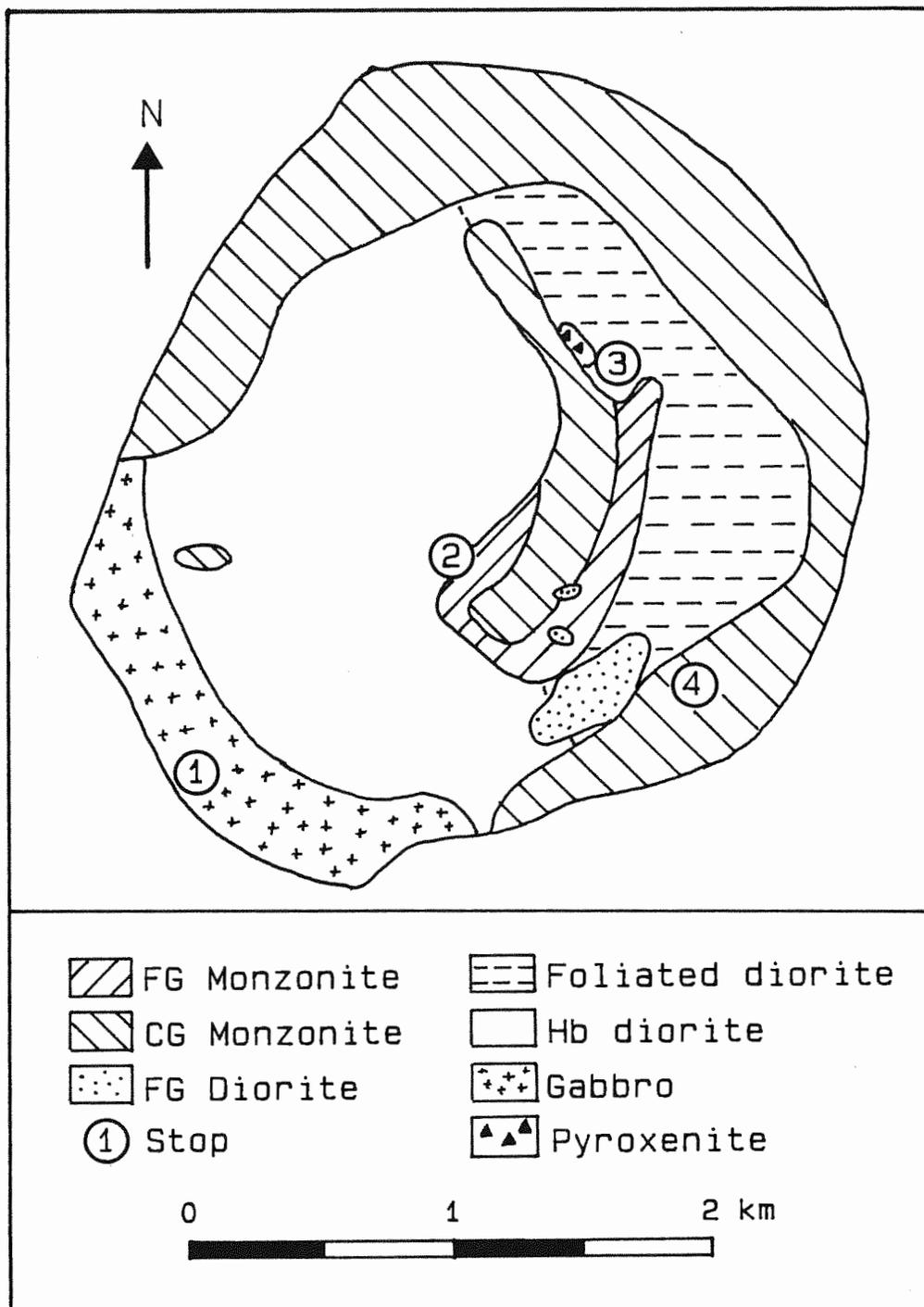


Figure 12. Geologic map of Mount Pawtuckaway from Roy and Freedman (1944), modified on the basis of additional mapping by Richards (1990), Kick (1992, unpublished), and Eby (1993, unpublished).

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a weak foliation. The medium-grained foliated diorite is differentiated on the basis of its generally higher pyroxene/amphibole ratio and the presence of foliation due to the alignment of plagioclase laths. These rocks are largely confined to the eastern portion of the complex. The fine-grained diorite is a distinctive unit texturally and shows no foliation. Because of the absence of foliation it is inferred to be the last mafic unit emplaced in the complex.

The felsic rocks are monzonites and syenites. The outer ring of coarse-grained monzonite grades inward to coarse-grained syenite. An arcuate unit of coarse- to medium-grained monzonite occurs within the complex. The fine-grained monzonite partially surrounds this arcuate structure and is found as inclusions in the coarser-grained monzonite. Dikes of what appear petrographically to be fine-grained monzonite cut the outer coarse-grained monzonite. Most of these units are cut by mafic and felsic dikes which represent the last stages of igneous activity at Mount Pawtuckaway.

The pyroxenites are apparently the earliest rocks to be emplaced at Mount Pawtuckaway. The arcuate gabbro which outcrops at the southern edge of the complex may be related to the pyroxenite. In terms of both texture and geochemistry, the pyroxenite and gabbro can be interpreted as cumulus rocks. The initial phase of magmatic activity at Mount Pawtuckaway, therefore, may have consisted of the precipitation, from the wall inward, of minerals from a convecting mafic magma. The coarse- and medium-grained diorites were the next units to be emplaced. The foliation in these units dips towards the center of the complex and the dip increases as one moves inward (Kick, 1992, unpublished data), suggesting a funnel-shaped intrusion. The fine-grained diorite does not show any foliation, and does occur as inclusions in the medium-grained monzonite, thus it must be the last of the mafic units emplaced, but must predate at least some of the felsic units. The fine-grained monzonite occurs as inclusions in the medium- to coarse-grained monzonite of the central arcuate unit. If the fine-grained monzonite dikes in the outer coarse-grained monzonite are related to the central fine-grained monzonite then there are at least two periods of coarse-grained monzonite emplacement.

The sequence of events as deduced from the field relations indicate that a number of different magmas must have been involved in the formation of the Mount Pawtuckaway intrusion. The initial magmas were mafic. Later magmas were more felsic in composition. All of the felsic rocks are broadly similar in chemical composition, and their textural differences may be due to the water pressure at the time of crystallization. There is also evidence for several periods of subsidence. The first formed the outer ring of coarse-grained monzonite. This was followed by the emplacement of the fine-grained monzonite which, on textural grounds, may represent a magma which vented to the surface. Subsequently the central arcuate coarse- to medium-grained monzonite was emplaced. The last period of igneous activity is represented by the emplacement of mafic and felsic dikes which cut all of the other units.

## PETROGRAPHY

### Mafic Units

**Pyroxenite.** The pyroxenites are coarse-grained and largely composed of cumulus olivine and augite with interstitial labradorite and opaque minerals. The augites show a pink tint and are spotted and rimmed by red-brown amphibole. The augites contain minute opaque inclusions which are oriented parallel to crystallographic directions. The opaques are titanomagnetite intergrown with hercynite. Apatite occurs in trace amounts.

**Gabbro.** The gabbros are medium- to coarse-grained and locally show a well-developed foliation due to the alignment of plagioclase laths. Plagioclase ( $An_{60}$  to  $An_{46}$ ) and a light pink augite are the major minerals. Olivine is locally abundant. The augites contain oriented minute opaque inclusions and are rimmed and spotted by red-brown amphibole.

**Hornblende diorite.** The grain size is variable from medium-fine-grained to coarse-grained, and locally foliation can be found. The plagioclase is generally andesine, but can be zoned to oligoclase. The pyroxenes are generally light green, but pink cores are not uncommon. The pyroxenes are often extensively replaced by

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reddish-brown hornblende and green hastingsite. Red-brown biotite occurs both as separate grains and replacing pyroxene and amphibole. Apatite is a common accessory ranging in modal abundance from 1.6 to 3.5%. Olivine, extensively altered, is occasionally found.

**Foliated diorite.** The grain size is variable from fine- to coarse-grained, and most specimens show a foliation due to the alignment of plagioclase grains. The plagioclase is generally andesine, but may be zoned to oligoclase and occasionally labradorite cores are found. The pyroxenes are light pink and light green in color, and where the two varieties occur together the light pink pyroxene constitutes the core. The pyroxenes are invariably partly replaced by red-brown to green amphiboles. The pyroxenes contain oriented minute opaque inclusions, and the preservation of these inclusions in the amphiboles indicates the prior existence of pyroxene. Olivine is found in most specimens and locally is an important accessory. The biotites are straw brown to red brown and generally occur as large flakes. Apatite is an important accessory ranging in modal abundance from 1.0 to 2.6%.

**Fine-grained diorite.** The rock consists of a felted matrix of plagioclase ( $An_{41}$  to  $An_{26}$ ), hornblende, minor biotite and trace apatite and opaques with minor small phenocrysts. The phenocrysts are plagioclase, some with alkali feldspar overgrowths, and hornblende. Aphanitic dark gray blebs of mafic minerals are also found.

Table 1. Representative Modes for the Mafic Rocks

	Pyroxenite	Gabbro		Hb diorite		Foliated diorite		Fine-gr. diorite
	MP73	MP81	MP83	MP1	MP50	MP9	MP75	K22
Plag	8	61	67	46	64	45	63	57
Oliv	18	5	21	-	-	7	1	<1
Pyroxene	51	10	<1	16	1	11	6	10
Amphibole	18	19	6	20	25	24	20	15
Bio	-	-	-	9	1	2	4	15
Opaques	5	5	6	6	5	8	4	2
Apatite	<1	-	<1	3	4	3	2	<1

### Felsic Units

**Coarse-grained Monzonite and Syenite.** The grain size varies from medium- to coarse-grained, and the monzonites and syenites are gradational into each other with changes in the K-feldspar/plagioclase ratio. The plagioclase is generally oligoclase and the alkali feldspars are microperthitic. The plagioclases are often rimmed by perthite. The pyroxenes are colorless to light green and are partly replaced by red-brown and dark green amphiboles. The biotites are reddish brown to straw brown and replace both pyroxene and amphibole. Quartz is interstitial, and some sections contain fayalitic olivine.

**Fine-grained monzonite.** The grain size varies from fine-grained to very-fine-grained. Some sections have phenocrysts of biotite and hornblende. The major minerals are oligoclase and microperthite. Quartz is a minor phase. The amphiboles are green to dark green and the biotites are reddish brown to straw brown. Pyrrhotite and apatite occur as accessories.

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Table 2. Representative Modes for the Felsic Rocks

	Coarse-grained monzonites & syenites				Fine-grained monzonite
	MP8	MP12	MP15	MP49	R&F <sup>1</sup>
Plag	19	31	24	8	40
K-feldspar	46	34	54	84	35
Quartz	-	<1	2	3	2
Olivine	1	1	-	-	-
Pyroxene	12	14	4	<1	4
Amphibole	<1	6	11	3	7
Biotite	15	8	4	-	9
Opaque	5	5	1	2	3
Apatite	1	1	-	-	-

<sup>1</sup>Roy and Freedman (1944)

## ROAD LOG

## Mileage

- 0.0 Entrance road to Pawtuckaway Mountains and fire tower. The entrance is marked by a small brown sign on the east side of NH Route 107, 3.2 miles north of the juncture of Routes 107 and 27.
- 2.0 Park by small farm cemetery on south side of road.

**STOP 1. GABBRO AT MELOON HILL (1 HOUR)** (Mt. Pawtuckaway Quadrangle). Follow logging road departing from west side of cemetery 300 feet to large flat outcrops of gabbro. Take left hand fork and continue another 1000 feet and then proceed southwesterly to SW side of Meloon Hill. Outcrop is essentially continuous along this side of the hill. The gabbro is medium-grained and locally shows a well-developed foliation. The foliation strikes parallel to the contact and dips steeply towards the center of the intrusion. This unit is cut by several fine-grained mafic dikes. Return to vehicles.

- 2.3 Turn left onto loop road.
- 2.9 Walk east 200 feet along logging road. Small quarry located just to the north of the road.

**STOP 2. MIDDLE MOUNTAIN TRAVERSE (2 HOUR)** (Mt. Pawtuckaway Quadrangle). At the western end of the quarry hornblende diorite has been engulfed by fine-grained bluish-gray monzodiorite. Isolated outcrops of this monzodiorite are found in the immediate area. Proceeding eastward in the quarry outcrops of hornblende diorite are observed. These outcrops are cut by both felsite and fine-grained monzonite dikes.

Proceed southeastward from the quarry up Middle Mountain. A series of outcrops provide almost 100% exposure of the fine-grained monzonite. **CAUTION:** This rock is very brittle and fragments come off the outcrop like shrapnel. Do not wound yourself or a fellow geologist. There are slight variations in grain size throughout this unit, but they do not appear to be correlated with distance from the contact. At the top of Middle Mountain outcrops of coarse-grained monzonite are found. Proceed a short distance eastward through

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this unit. In this area the outcrops are deeply weathered and fresh pieces are difficult to obtain. Inclusions of fine-grained monzonite are found in some outcrops of coarse-grained monzonite. Return to vehicles.

- 3.6 Park at the intersection of the loop road and Round Pond road. During times of heavy rainfall the road may not be passable between Stop 2 and Stop 3.

**STOP 3. ROUND POND ROAD TRAVERSE (2 HOUR)** (Mt. Pawtuckaway Quadrangle). Walk back (west) along the road several hundred feet to a road leading north into a primitive picnic area. Outcrops of coarse-grained monzonite are found on either side of the road. Diorite and fine-grained monzonite inclusions are found in the coarse-grained monzonite. On the east side of the road are several outcrops of fine-grained monzonite which contain blebs of coarse-grained monzonite. In thin section no sharp boundaries are observed between the two types of monzonite.

Return to the intersection and continue on the Round Pond road in an easterly direction. Outcrops of pyroxenite are found along the road approximately 400 feet from the intersection. Follow the ridge line northward about 300 feet to an outcrop of large pyroxenite blocks in medium-grained monzonite.

Return to Round Pond road and continue eastward onto an abandoned road. Outcrops of foliated diorite are found in and on both sides of the road. Both fine- and medium-grained varieties of the foliated diorite are observed. Where the two varieties are in contact, the fine-grained diorite appears to intrude the medium-grained diorite.

Continue eastward to Round Pond. Outcrops of coarse-grained monzonite north of the road and just west of the brook carry inclusions of fine-grained diorite. A mafic dike cutting the monzonite is exposed in the stream bed. Return to vehicles.

- 4.5 Continue southward on loop road to parking area for fire tower trail.

**STOP 4. SOUTH MOUNTAIN TRAVERSE (2 HOUR)** (Mt. Pawtuckaway Quadrangle). Proceed up trail to top of South Mountain. Excellent exposures of the coarse-grained monzonite are found along the upper portion of the trail. Several fine-grained monzonite dikes cut this unit. A number of mafic dikes are exposed in the immediate area of the fire tower. On a trail going southward from the fire tower is an exposure showing mixing between felsic and mafic magmas. Return to vehicles. Continue on loop road to juncture. Turn right and proceed back to Route 107.

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