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**NORMAN L. HATCH, JR.**
(1932-1991)

# GUIDEBOOK FOR FIELD TRIPS IN THE CONNECTICUT VALLEY REGION OF MASSACHUSETTS AND ADJACENT STATES

## VOLUME 2

PETER ROBINSON AND JOHN B. BRADY, EDITORS

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## EDITORS' COMMENTS

Our role as editors was limited to harassing authors into submission and making sure that requirements of space, format, and figure legibility were adhered to as much as possible. The trips and authors chosen reflect a combination of need to cover the region's current geologic research and of need to provide adequate trips for hundreds of participants. We did not "edit" in the traditional sense, nor did we even have time to read the various articles in detail. Thus, these volumes serve as a completely free and immediate forum for the authors' ideas, and there has been no attempt to balance with alternative viewpoints. We hope that scientists with different information and interpretations will try to have these published through ordinary channels. In the introduction in Volume 1, the term "craton X" was wrongly ascribed Philip Osberg. It was a term used by E-an Zen in 1983, and Robinson, as a native X patriot, feels chagrin at his error. We thank Edward S. Belt for calling attention to the beautiful lithographs in Edward Hitchcock's report and especially Walter P. Coombs for providing the following succinct description of the slabs illustrated in the guidebook covers.



## COVER ILLUSTRATIONS

Reproduced on the front and back covers of both volumes are selected lithographs from the classic 1858 work by Edward Hitchcock: "Ichnology of New England." Writing at a time when dinosaurs were known only from fragmentary skeletons, Hitchcock thought the fossil ichnites had been made by the three-toed feet of giant birds. In an ironic twist of fate, birds are now regarded as direct evolutionary descendants of bipedal, meat-eating dinosaurs. So Hitchcock was right after all!

**VOL. 1, FRONT COVER:** A slab from "the Lily Pond," one of several footprint quarries opened by Roswell Field on his farm in Northfield, Mass. The quarry site is now part of the Barton Cove recreation area that is operated by Northeast Utilities. Hitchcock identified five species of footprints on this slab, but there are probably only two, the larger of which may be called *Eubrontes* , and the smaller may be called *Grallator*. Both have traditionally been regarded as carnivorous dinosaurs footprints, but the size and abundance of *Eubrontes* suggests the trackmaker was an herbivore. *Grallator* prints are commonly linked to *Coelophysis*, a small carnivorous dinosaur known from a few skeletal fossils in the Connecticut Valley, and numerous skeletons from the Chinle Formation in New Mexico.

**VOL. 1, BACK COVER:** The slab in Fig. 1 comes from "the orchard" on Roswell Field's farm, and the slab in Fig. 2 comes from the Lily Pond. The slab in Fig. 3 comes from Smiths Ferry, a location along the Connecticut River in Easthampton, Mass. where a tow-rope raft-ferry was once in operation. Most of the tracks on these slabs are assigned to *Anchisauripus*, a track similar in morphology but intermediate in size between *Eubrontes* and *Grallator* prints, which are also present on these slabs. There is a minority theory that *Grallator* , *Anchisauripus*, and *Eubrontes* trackways were made by different age categories of one dinosaur. The diagonal structures in Fig. 1, which appear to be a consequence of slumping, were called *Helcura* by Hitchcock and were interpreted as turtle trails. The slab in Fig. 2 contains small quadrupedal tracks called *Batrachopus* . The unknown trackmaker, perhaps a crocodilian or pseudosuchian but not a dinosaur, had four toes on the pes and five on the manus, although the outermost finger rarely leaves an imprint.

**VOL. 2, FRONT COVER:** This famous specimen was for 60 years part of a sidewalk in Middletown, Connecticut, near where the rock was quarried. The tracks, assigned to *Grallator* , were discovered on the underside when the sidewalk was torn up. Hitchcock regarded this specimen as "the gem of the collection." Pads and claws can easily be seen on the footprints, which are preserved in high raised relief on the underside of sediment that covered and filled the original trackway. There are numerous mudcrack fillings some of which emanate from the tips of the toes. Apparently the original surface had developed a crust that was broken as the dinosaurs crossed. One might say that this slab records "dinoturbation" on the bottom and "homoturbation" on the top.

**VOL. 2, BACK COVER:** The upper slab comes from Turner's Falls, the lower from the Lily Pond. The largest footprints on the upper slab are assigned to *Eubrontes*, the slightly smaller tracks are referred to *Anchisauripus*, and the numerous small prints are called *Grallator*. If this were the only specimen, it would be easy to distinguish among these three taxa. Unfortunately, tridactyl prints on other specimens are intermediate in size. On the lower slab, at the upper left, there is a series of three tracks with a tail drag mark. Among Connecticut Valley trackways, tail drags may be present under three circumstances: 1) certain footprint types commonly (always?) have a tail drag; 2) trackways that lead to a "sitting print" sometimes have a tail drag; and 3) tracks of small dinosaurs that were walking through a fairly thick layer of soft, water-logged sediment sometimes have a tail drag.

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# TECTONIC-COVER BASEMENT RELATIONS AND METAMORPHIC CONDITIONS OF FORMATION OF THE SADAWGA, RAYPONDA AND ATHENS DOMES, SOUTHERN VERMONT

by
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## INTRODUCTION

This trip will examine the structural and metamorphic evolution of Acadian interference domes and basins east of the Green Mountain massif (cored by Middle Proterozoic rocks) of southern Vermont (Fig. 1). Detailed 1:24,000 mapping by Armstrong and Ratcliffe and by Ratcliffe and others (1988) identified in figure 2 has shown that a regional set of multiply interfering folds, schistosities and cleavages can be used to explain the geometry of the Acadian structures. Crystalloblastic structures and garnet-inclusion fabrics indicate that peak metamorphic conditions developed during the latest stages of the dome evolution. Garnets grew later than much of the deformation and include many interference structures still present in the schist matrix, giving rise to inclusion trails that simulate synrotational growth fabrics. Thermobarometric determinations of syn-dome fabrics by Armstrong indicate a monotonic pressure gradient from 6.7 kbar (24 km) in the west to 9 kbar (32 km) in the east that is not accounted for by the low amplitude of the domes, which is less than 3 km. These data indicate that the Green Mountain massif and the Rayponda-Sadawga and Athens domes actually formed at different depths in the easterly thickening tectonic wedge. $^{40}Ar/^{39}Ar$ data (Sutter and others, 1985) and unpublished data of Laird and Sutter (reported in Karabinos and Laird, 1988, p. 289) indicate the last and major metamorphism in this area was Acadian.

These results and conclusions differ from interpretations of the Acadian history of Rosenfeld (1968) and of Rosenfeld and others (1988) which was derived in part from his study of syntectonic inclusion geometries of garnets. According to them, the Acadian deformation involved Stage I - westward-directed nappe-stage folds in Middle Proterozoic basement through Silurian and Devonian rocks that was followed by Stage 2- diapiric uprise of domes and late-stage eastward-directed backfolding.

We conclude, however, that the inclusion fabrics within the garnets and structures in the matrix can be explained by (1) static growth across preexisting microfolds of schistosities, (2) growth during development of late crenulation cleavage, and (3) by kinking of matrix schistosity outside the garnets by post-garnet deformation. Moreover, we also find that the matrix outside the garnets preserves cleavages, folds and multiple overprint fabric far more comprehensive and significant than those included within the garnets themselves. In this matter we agree in part with the interpretations of Bell and Johnson (1989) and Hayward (1992) that spiral inclusion fabrics of garnets in and around the Athens dome do not require simultaneous growth and rotation. But we strongly disagree with them that the garnets retain a record more complete than that found in the matrix, principally because the garnet grew too late in the tectonic evolution to record all the complexity found in the rocks themselves.

## PREVIOUS WORK

Unpublished geologic mapping in this area conducted by J.L. Rosenfeld (1954), J.B. Thompson, Jr. and D.E. Wilhelm was used to compile the 1961 1:250,000 bedrock geologic map of Vermont (see Doll and others, 1961, and references contained therein). Rosenfeld and others (1988) presented a regional geologic map covering part of the Townshend area. Detailed geologic maps of this area include the Jamaica area (Karabinos, 1984a), the Wilmington 15-minute quadrangle Skehan (1961) and the Brattleboro 15 minute quadrangle of Hepburn and others (1984). Ratcliffe and others (1988) presented a regional geologic map that compiles mapping by Ratcliffe and William Burton in the Jamaica, Peru, Londonderry, Woodford (Burton, 1991),

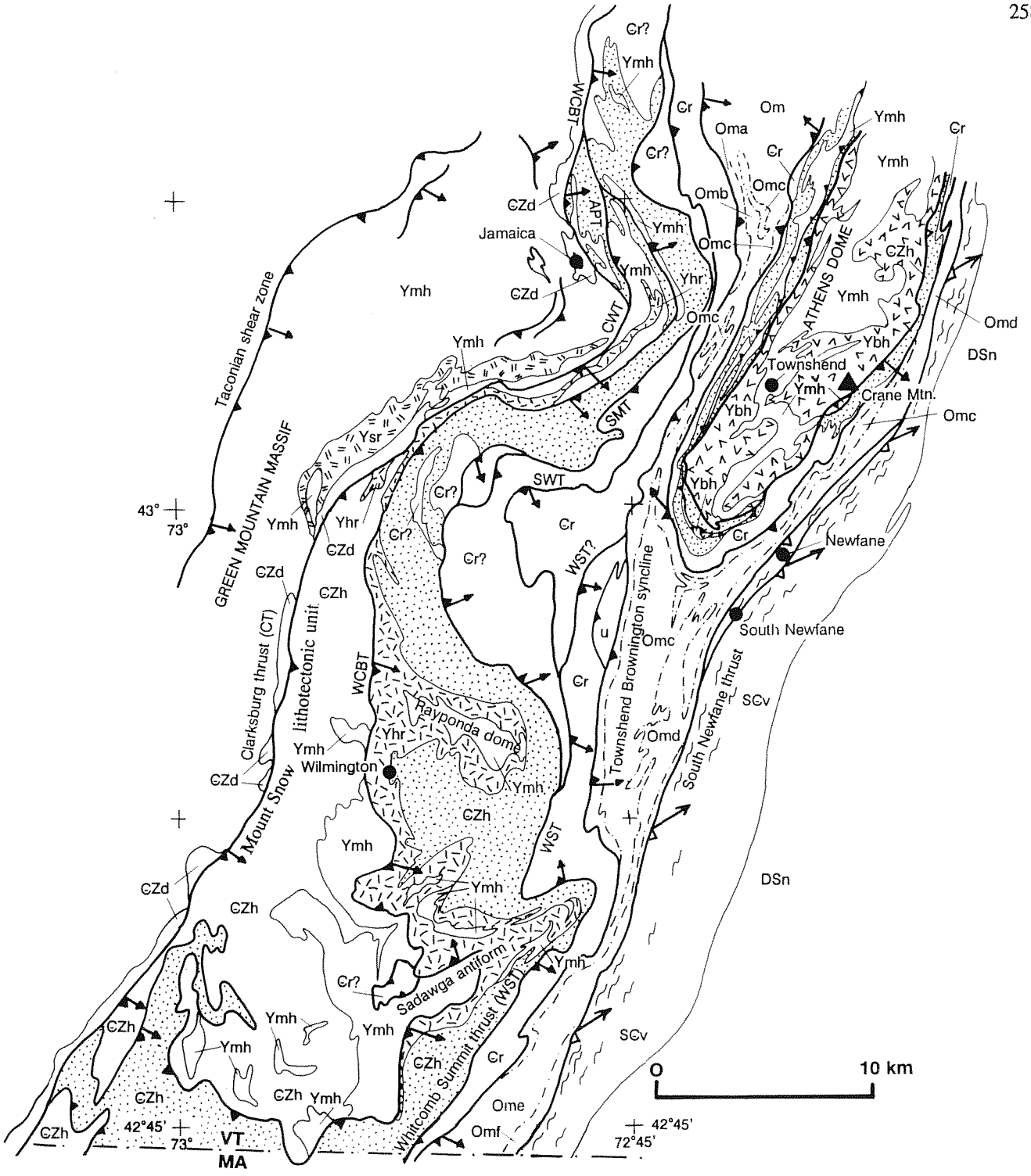
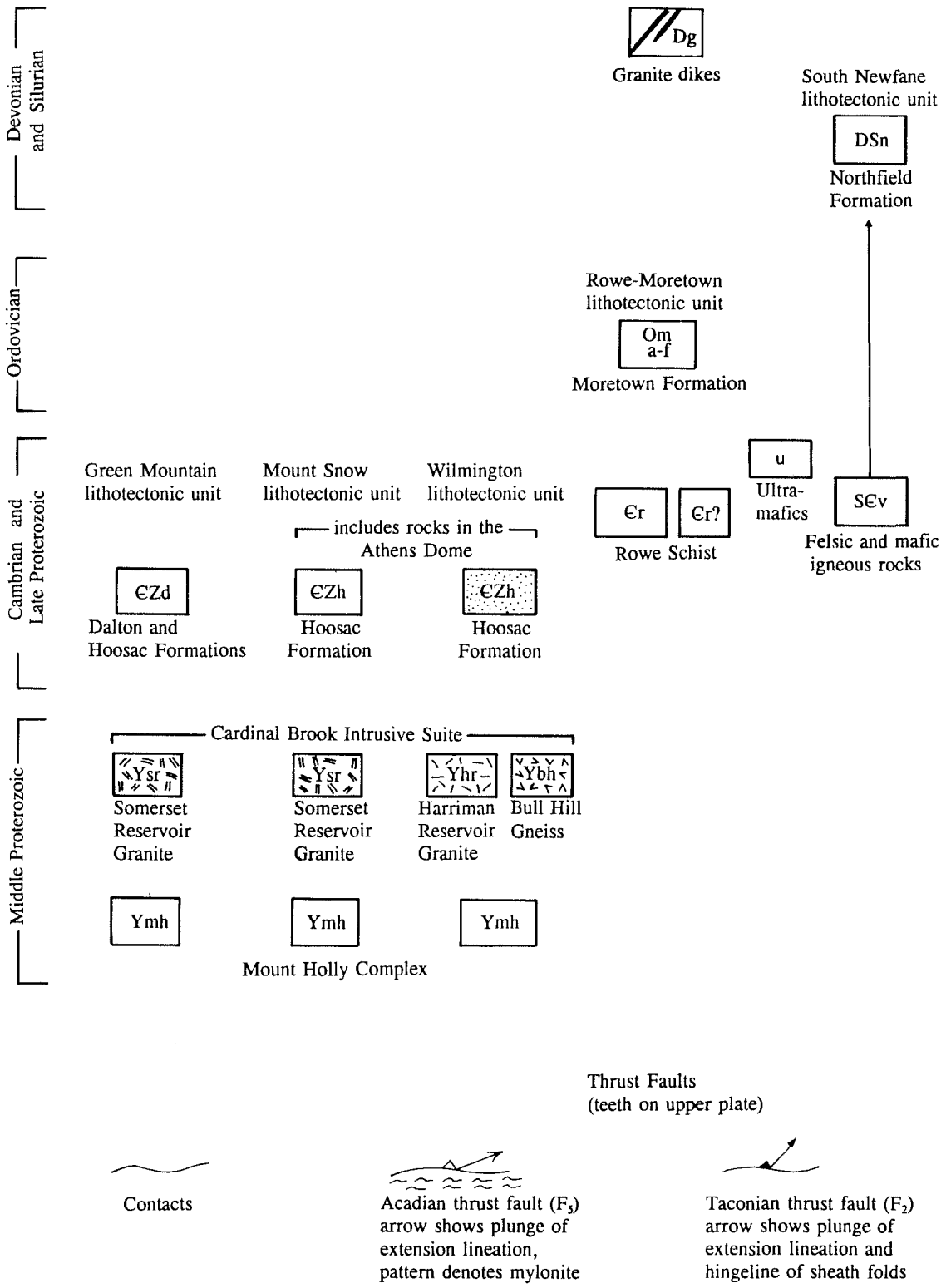



Figure 1. Generalized geologic map of the Sadawga-Rayponda dome and southern part of the Athens southern Vermont. Devonian granite dikes identified on figure 2. Contacts between units a-f in the Moretown Formation shown by dash-dotted lines.

CORRELATION OF MAP UNITS



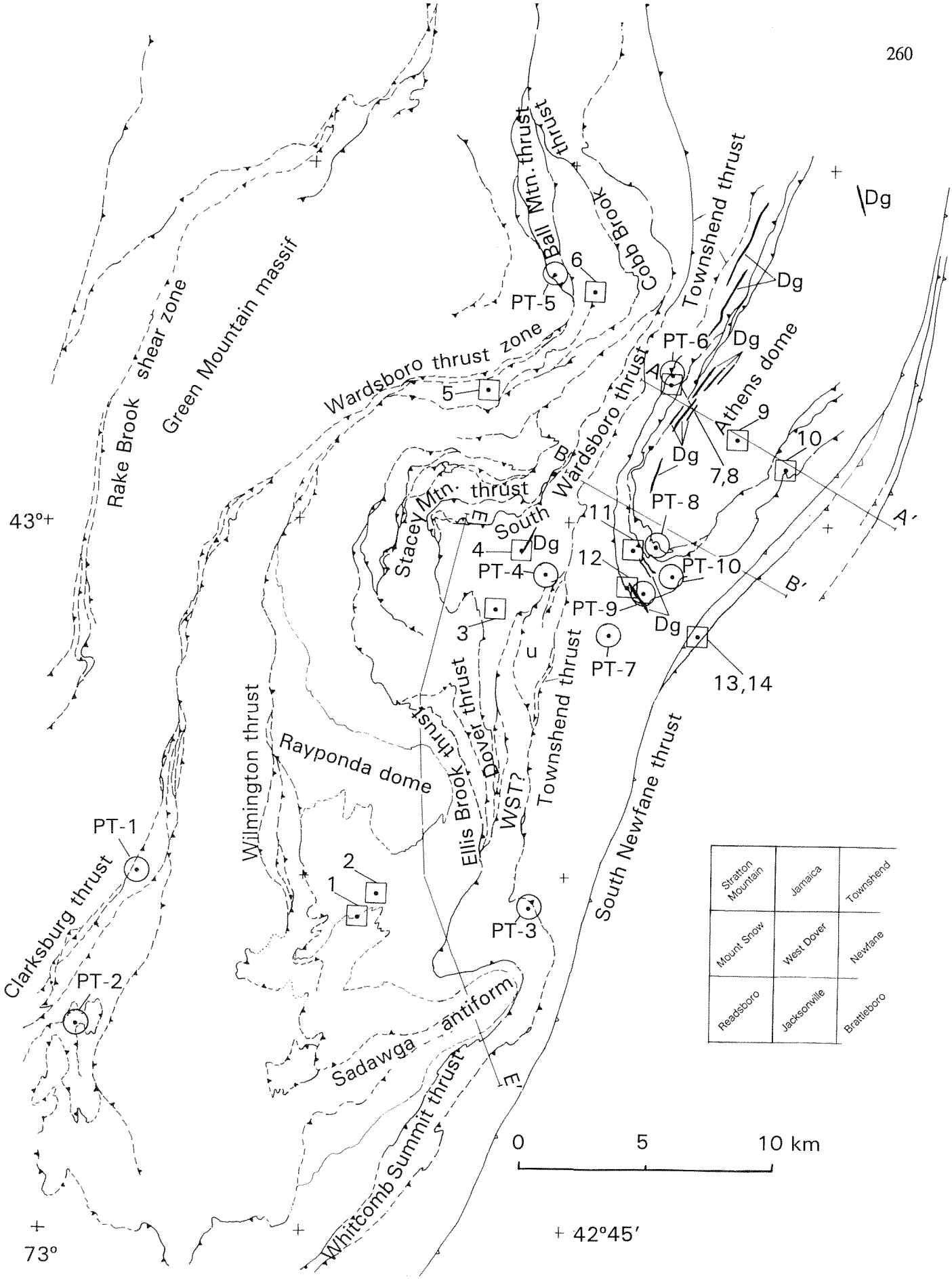



Figure 2. Tectonic map showing major faults and location of field trip stops (squares) and of pressure temperature determinations PT-1 to PT-10 in table 1. Dg refers to granite dikes not identified in figure 1.

Stratton Mountain (Ratcliffe and Burton, 1989), Mount Snow, and Readsboro (Ratcliffe, 1992) and Sunderland (Burton, in press) quadrangles. A generalized geologic map of the core of Athens dome and cover rocks based on mapping of Ratcliffe in 1988-1990 may be found in Ratcliffe (1991a). A compilation at 1:250,000 by Thompson, McLelland, and Rankin (1990) presents revision to the 1961 map of Doll and others (1961) of the Athens dome and Jamaica area. These revisions are based on unpublished manuscript maps of Rosenfeld and Thompson dated 1987. These or other unpublished source materials cited in Doll and others (1961) and in Thompson, McLelland and Rankin (1990) are not generally available and have not been examined. All geology shown here is the result of mapping by the authors, including remapping of the area previously published by Karabinos (1984a); much of the geology mapped by him has been confirmed although some modifications have been made.

Cover rocks east of and above the Hoosac Formation have been mapped by Doll and others (1961) as an eastward-topping section consisting of the Pinney Hollow, Ottauquechee, and Stowe Formations. In adjacent Massachusetts, Hatch and others (1966) and Chidester and others (1967) were unable to map this stratigraphy because of extensive lateral and vertical variation of rock types in this belt. They chose to abandon the Vermont nomenclature and adopted the name Rowe Schist to apply to all rocks between the top of the Hoosac Formation and the first beds of the Moretown Formation. This practice has been adopted here, and the names Pinney Hollow, Ottauquechee, and Stowe are not used for the same reasons cited by Hatch and others (1966), Stanley and Hatch (1988), and by Ratcliffe (1991b), namely the heterogeneous discontinuous nature of the rock units within this belt and the uncertainty of correlation of these belts to the type sections exposed 50 to 100 kilometers to the north.

Many excellent studies of the metamorphic petrology and of metamorphic textures have been conducted in this area. Rosenfeld (1954, 1968) first recognized the possibility that Acadian metamorphism overprinted pre-Acadian metamorphic fabrics in Cambrian and Ordovician rocks. He noted the presence of texturally and chemically zoned garnets and proposed that "unconformity garnets" from the cover rocks included relict Taconian cores. Karabinos (1984a, b) studied these "unconformity garnets" and observed Mg/Fe and Mn/Fe variations which may indicate not only distinct overgrowth rims but a period of retrogression prior to rim (Acadian) overgrowth. Karabinos's (1984a, b) conclusions are generally supported by regional relationships and $^{40}Ar/^{39}Ar$ dating studies (Sutter and others, 1985; Laird and others, 1984; Ratcliffe and others, 1988; Burton and others, 1990, 1991) all of which provide evidence for Taconian metamorphism in a belt immediately west of the Jamaica area in the Green Mountain massif. However an alternative interpretation is that reversals in chemical zoning may result from changes in equilibria that involve garnet resportion during a single prograde (Acadian) metamorphism (Thompson and others, 1977; Armstrong and Tracy, 1991). Unfortunately, definitive proof of Taconian metamorphism within the area of the present trip is not available owing to the high grade overprint of Acadian age which is well supported by $^{40}Ar/^{39}Ar$ hornblende cooling ages of 376 to about 389 Ma (Laird and others, 1984; Sutter and others, 1985; Laird, 1991). The general conclusion of previous authors is accepted here, namely that Cambrian and Ordovician cover rocks of the Green Mountain massif and the area to the east were subjected to a medium-grade Taconian Barrovian metamorphism at about 455 Ma and subsequently overprinted by an Acadian garnet to kyanite event at about 389 Ma. In the area of this field trip, Taconian metamorphism probably was at least garnet-grade and perhaps higher based on the northward projection of Taconian metamorphic zones from the south (Sutter and others, 1985). The crustal position of these rocks between the end of the Taconian and beginning of the Acadian is uncertain; they may have been deeply buried through much of this time.

In addition to the local studies, the unpublished work of Nisbet (1976) on the northern closure of the Chester dome contributed to our understanding of the southern domes. He concluded from study of overprinted foliations and folds that three stages of noncoaxial and noncoplanar folding affected the basement and cover there. He further concluded that the reverse asymmetry of "dome" stage folds can be explained by a late foliation transection of already folded rocks. We agree in general with Nisbet's analysis.

**The Garnet Inclusion Problem**

Recent papers, Bell and Johnson (1989), Hayward (1992), to mention a few, have sparked a world-wide debate regarding the significance of inclusion fabrics in porphyroblasts and whether or not spiral or spiral-like inclusions trails are evidence for syntectonic growth and rotation. Some of the data used by Hayward in particular come from the area of this field trip. Hayward (1992) sampled many of the localities Rosenfeld (1968) used to establish rotation senses to support the nappe-stage and dome stage model. Simply put, Bell and Johnson (1989) and Hayward (1992) proposed that (1) garnets do not provide evidence for synrotational growth because internal trails are not continuous but markedly discontinuous owing to multiple subhorizontal and subvertical internal discontinuities; (2) these discontinuities represent successive foliations produced by alternating horizontal compression (crustal thickening) and horizontal extension (crustal collapse) in accordion style deformation. The garnet retains this record whereas the schist matrix is continually transposed, flattened or reactivated so as to destroy the fabrics from older events; (3) these internal foliations have maintained their spatial orientation through these successive events because the garnet porphyroblasts are irrotational in the deforming schist matrix as adduced from the preponderance of subvertical and subhorizontal discontinuities in the garnet over wide areas (i.e., the entire area of the Chester and Athens domes; Hayward, 1992).

The boldness of these propositions is breathtaking. If accepted, they destroy the usefulness of rotational garnet fabrics and obviate the necessity of geologic mapping because the record of the deformational history is contained and best seen only in the study of the garnets themselves. The arguments for or against these positions have sparked a series of discussions reminiscent of the Copernican dialogues (Passchier and others, 1992; Bell and others, 1992).

Are there any field and geochemical observations that can help resolve some of these differences? We believe there are.

Structural observations: (1) multiply folded foliation surfaces ($S_2$, $S_3$, $S_4$, $S_5$) can be mapped as shown in figures 4 and 5; (We have shown that the $S_2$, $S_3$, and $S_4$ surfaces are not constant in space but are folded in successive shortening events.); (2) garnet statically overgrows microfolds of folded schistosity seen in outcrop to have been produced by the axial surface of $S_3$ crossing $S_2$, and by $S_4$ crossing $S_2$ and $S_3$. The dominant apparent-rotation sense in outcrop is the result of the preserved asymmetry resulting from short-limb and long-limb bias of the overprint relations.

Geochemical observations: (1) garnets show smooth and continuous element distribution patterns without discontinuities; (2) composition of matrix and included phases and changes in garnet Fe-Mg, Mn-Ca concentrations fit predicted mass-balance equations for a single prograde metamorphic history.

We believe our observations indicate that the garnets completed growth during the $F_4$ and $F_5$ fold events. At this time, $S_2$ foliation was steeply dipping east or west in tightly appressed folds and $S_3$ was locally present in subhorizontal orientation on the crest of $S_4$ folds. This may explain both the nonrotation of the garnets (seen by us) and the prepondence of subhorizontal and subvertical discontinuities observed by Bell and Johnson (1989) and by Hayward (1992). Subsequent tightening of $F_2$ and $F_3$ folds by $F_4$ and $F_5$ did distort relationships in the matrix and the garnets did remain relatively fixed in space because deformation following their growth was confined to subhorizontal shortening by $F_4$, $F_5$ and possibly later events.

In the last analysis it may prove that the Chester-Athens dome area was the wrong locality to use to test the irrotational model of garnets because in fact they grew too late in the deformational history here to have been rotated by early events.

Regarding the "unconformity" garnets described by Rosenfeld (1968) and by Karabinos (1984a, b), we have noted that garnets throughout this area do contain a concentric zone of oriented ilmenite and rutile inside the otherwise clear rims. This textural change marks the unconformity as defined by Karabinos and by Rosenfeld. Armstrong and Tracy (1991) and Armstrong (1992) explain the rutile-ilmenite concentration and the clear garnet rims by decrease in garnet growth rate near the rims and relatively complete resorption of the inclusions in the

rim. This model is supported by the observation that many of the "unconformity" garnets have no irregularities in chemical zoning and contain continuous core to rim inclusions trails of matrix phases that cross the unconformity. Based on these observations Armstrong and Tracy concluded that unconformities they see are apparent unconformities only.

## REGIONAL GEOLOGY

Middle Proterozoic rocks of the Mount Holly Complex dated at 1.3 to 1 Ga (Ratcliffe and others, 1991), form the core of the Green Mountain massif, as well as of the Rayponda, Sadawga and Athens domes to the east. In each area post-Grenville (Ratcliffe, 1991a) granites of the Cardinal Brook Intrusive Suite (Stamford Granite, Somerset Reservoir Granite, Harriman Reservoir Granite and Bull Hill Gneiss as used by Ratcliffe, 1991) dated at approximately 960 Ma years old (Karabinos and Aleinikoff, 1990), intrude the Mount Holly Complex. Rocks of the Cardinal Brook Intrusive Suite do not contain the coarse-grained hornblende granulite-facies gneissosity present in the Mount Holly. The first foliation developed in these rocks is Ordovician or Taconic. Rocks of the Hoosac Formation of Late Proterozoic and Early Cambrian age unconformably rest on the Mount Holly Complex and Cardinal Brook Intrusive Suite. The Hoosac contains metabasalts having alkalic to tholeiitic chemistry that are thought to be within-plate rift basalts associated with pre-Iapetan rift event. Upsection in the Hoosac the metasediments become finer grained and more aluminous as the basalts evolve into oceanic Morb-like flows (Ratcliffe, 1991b).

### The Rowe Schist Problem

Rocks above the Hoosac Formation consist of a heterogeneous assemblage of rock in the structural position of the Pinney Hollow, Ottauquechee, Stowe Formations of central Vermont, all of which had been mapped in this area by Skehan (1961) and by Doll and others (1961). Our detailed remapping of this area, however, shows that lithologic belts do not correlate well with the formations as previously drawn, in part due to abundant thrust faulting (Ratcliffe, 1990, in press). In figure 1 we have not shown the detailed geology and the lithic units we have mapped within the Rowe Schist belt because of limitations imposed by scale. The detailed distribution of units are presented in Ratcliffe (1992, in press) and in the geologic map of the West Dover and Jacksonville area (Ratcliffe and Armstrong, in prep.). We have chosen to use the nomenclature of Stanley and Hatch (1988) that was employed by them in Massachusetts and refer to all rocks between the top of the Hoosac Formation and the base of the Moretown Formation as Rowe Schist or Rowe Schist(?) pending further work to the north. Indeed, when we map rocks of the upper parts of the Rowe southward, they pass into rocks coextensive with the Rowe of Massachusetts. The entire package of lower rocks (Ꞓr? on figure 1) in the position of the Pinney Hollow, Ottauquechee and Stowe as mapped by Doll and others (1961) wedges out by fault truncation beneath the true Rowe. At the Massachusetts-Vermont state line the Rowe Schist consists almost entirely of amphibolite intercalated with biotite-muscovite-large garnet schist, biotite feldspathic granofels and pale- green magnetite muscovite-chlorite quartz schist similar to the Stowe Formation of central Vermont. We believe that this Rowe Schist (Ꞓr on figure 1) reemerges as the Rowe Schist mantling the Athens dome.

### Major Lithotectonic Units and Levels of Basement Rocks

The general lithologic package discussed above, extending eastward from the Green Mountain massif, is highly imbricated internally by a series of thrust faults (Fig. 2) which carry slivers of basement gneiss and associated Hoosac Formation cover. Ratcliffe (1990, in press, a) has recognized three levels of basement rock in this area separated by major thrust faults. From west (bottom) to east (top) they are (see explanation to figs. 1 and 2):

1) Green Mountain lithotectonic unit west of the Clarksburg, Wardsboro, and Ball Mountain thrusts;
2) Mount Snow lithotectonic unit between the Clarksburg and Wilmington thrusts;
3) Wilmington lithotectonic unit above the Wilmington-Cobb Brook thrust but beneath the family of faults that carry the Rowe Schist rocks over the Hoosac to the west (these include the Whitcomb Summit, Ellis Brook, Dover, Stacey Mountain, and South Wardsboro thrusts).

Basement rocks that re-appear in the core of the Athens dome and in the tectonic cover of the Athens dome (Ratcliffe, 1990, 1991a) may belong to the Wilmington and Mount Snow lithotectonic units. East of Jamaica three fault bounded slivers of basement exist, each carries its attached sedimentary cover. These are termed from the west to the east the Adams Pond, Jamaica and Little Jamaica antiformal sheath folds. As suggested in figure 3 these levels may reappear as part of the tectonic cover of the Athens dome to the east. Figure 3 shows schematically the three levels of Middle Proterozoic rocks extending east from the Jamaica area. In this diagram younger rocks have been graphically removed so as to expose the surfaces of the basement rocks in the core of three antiformal sheath folds. Hingelines of these sheath folds originally plunged about South 65°E parallel to the prominent extension lineation (rodding) present in the fault zone rocks flooring each faulted antiform.

**Tectonic Cover of Athens Dome**

Tightly folded rocks of the Moretown Formation occupy the Townshend-Brownington "syncline." Where the Rowe Schist and Hoosac Formation rocks reappear around the periphery of the Athens dome, the Rowe Schist has been greatly thinned largely because of the absence of the Ottauquechee and Pinney Hollow lithologies of the Єr? belt present along the western limb of the Townshend-Brownington syncline. This thinned section contains abundant amphibolites and large-garnet schists like the type Rowe to the south. We believe the cover rocks above the Hoosac and below the Moretown, which mantle the Athens dome, correlates with the type Rowe. We speculate that this cover is largely equivalent to the Stowe Formation and that the older parts of the section in the Rowe(?) belt have been cut by internal faults. These internal faults are not shown on figures 1 or 2.

Structural relations in the Hoosac belt around the Athens dome are also complicated by faulting. In a zone about one km thick basement, Hoosac, and locally rocks of the Rowe are repeated by thrust faults and by folds to form a tectonic cover similar to that overlying the Green Mountain massif. In figure 1, three levels of basement rock are shown in the Athens dome. At the south margin of the dome a fault sliver of Bull Hill Gneiss is present in what is termed the Kenny Pond slice. Several faults repeat basement and cover along the eastern margin of the dome for 3 km north of Crane Mountain.

**Townshend Thrust and the Moretown Formation**

The contact between the Moretown Formation and the Rowe Schist is interpreted as a major thrust, here named the Townshend thrust. When the fault is traced south and then to the north along the east margin of the Athens dome, more and more section in the footwall is cut out, so that at the northern border of the map (Fig. 1), on the east flank of the Athens dome Moretown rests locally on the Hoosac Formation. These fault relations are also clearly shown along the western border of the Moretown from Jamaica south to Massachusetts where units within the Moretown (units a through f) truncate against the thrust at the base of the Moretown.

**South Newfane Lithotectonic Unit**

The South Newfane lithotectonic unit consists of a series of mafic and felsic meta-igneous rocks (SЄv on Fig. 1) that display complex cross cutting contact relations with one another as well as with interspersed metasedimentary rocks such as the Cram Hill Formation. Available U-Pb zircon ages range from Cambrian through Silurian (Aleinikoff and Karabinos, 1990). When combined with the high degree of internal complexity among rock types, these ages at present do little to resolve the age of the diverse igneous rocks or of the metasedimentary rocks of this unit. The rocks of this belt (SЄv) have previously been referred to as the Barnard Volcanic Member of the Missisquoi Formation (Doll and others, 1961; Hepburn and others, 1984). Pending result of ongoing geologic mapping and U-Pb zircon studies by Armstrong we will avoid the use of the term Barnard for these rocks. Hopefully the new studies will help resolve the uncertainties in age and origin that now exist for these rocks.

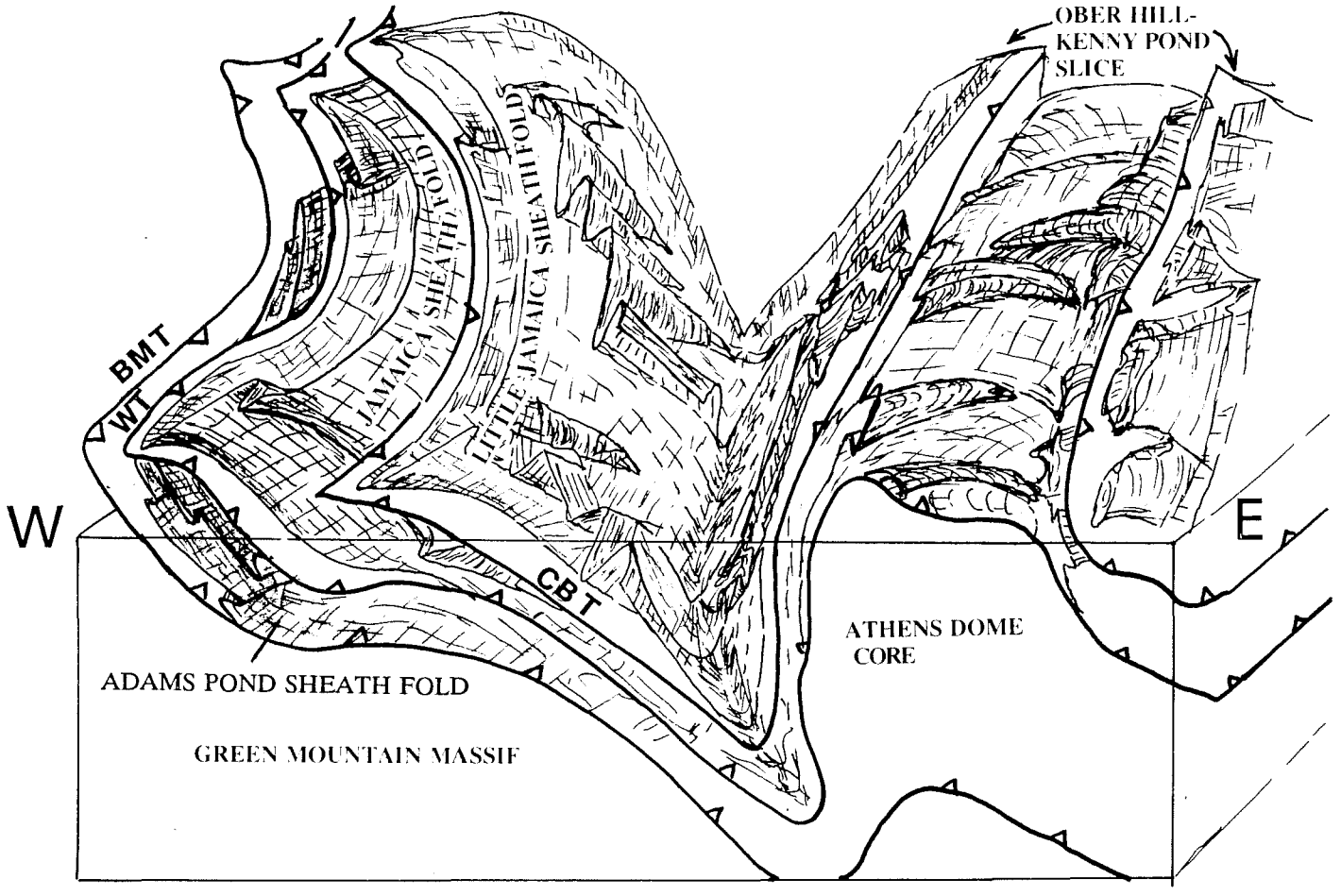



Figure 3. Generalized three dimensional diagram showing three fault-repeated slices of basement-gneiss antiforms in the Jamaica area and their connection with similar slices in the tectonic cover of the Athens dome to the east. Cover rocks have been graphically removed so as to expose the surfaces of the basement gneiss and to accent the sheath fold geometry of Taconian $F_2$ folds. BMT-Ball Mountain thrust, WT-Wardsboro thrust, CBT-Cobb Brook thrust.

The Northfield Formation (as used by Doll and others, 1961) overlies the S€v unit with apparent conformity. We find little evidence for a structural break (surface of structural disharmony, a true fault, or an unconformity) at the base of the Northfield. Complete concordance of structural elements exists between the S€v unit and Northfield within the areas mapped by us. The age of the Northfield is uncertain but probably is Silurian or Early Devonian, thus the structures and metamorphic assemblages in the unit must be Acadian or younger.

### Black Mountain Granite and Related Dikes

The foliated Black Mountain Granite (Hepburn and others, 1984) intrudes the core rocks of the Guilford dome approximately 15 km east of the Athens and Rayponda domes where it is regarded as an Acadian syntectonic intrusive. The age of the Black Mountain Granite has recently been determined by Aleinikoff to be $373 \pm 4$ Ma based on U-Pb zircon studies. This age compares favorably with the $372 \pm 5$ Ma Rb-Sr mineral-whole rock isochron of Naylor (1971), when recalculated using modern constants (personal communication from Aleinikoff, 1992). On the other hand these ages do not support the $326 \pm 17$ Ma Rb-Sr whole rock age of Hayward and others (1988) which would have required post-Acadian deformation in the area of the Guilford dome. Throughout the Athens dome and at a limited number of localities outside the dome to the south and west we have noted dikes of biotite-muscovite granite or aplite. These dikes are as much as 15 meters thick but commonly are 0.5 to 1 m thick. They are uniformly nonfoliated and cross cut structures and foliations in the domes. The locations of some of these dikes are shown on figure 2. If the granite dikes in the Athens dome are coeval with the Black Mountain Granite at $373 \pm 4$ Ma, then the late dome-stage folding of the Athens dome ceased before 373 Ma, whereas deformation continued later or began later to the east in the Guilford dome than in the Athens dome.

## STRUCTURAL GEOLOGY

The Middle Proterozoic through Lower Devonian(?) rocks of this area have been strongly affected by multiple fold events and intense metamorphism in the Acadian orogeny. All workers agree that the dominant fabrics and mineral textures were produced in the Acadian. However, there has been much debate regarding the extent or even presence of Taconian structures in or around the Rayponda-Sadawga and Athens domes. A main thesis of this trip is that relict thrust faults, folds, lineations, and schistosities inherited from the Taconic are still preserved in the domes and in the cover rocks as young as the Moretown Formation. Indeed the major tectonic boundaries are thrust faults of Taconian age (Fig. 2). Rocks younger(?) than the Moretown, or east of a major thrust fault, the South Newfane thrust, contain only Acadian structures, and importantly these structures and their sequence of development differ from Acadian folds found to the west. The mapping of Armstrong has shown that a major deformational front, associated with juxtaposition of unit S€v (Barnard Volcanic Member of the Missisquoi Formation of Doll and others, 1961) against rocks to the west, has overprinted the eastern flank of the domes. In addition, this late deformation increases in intensity eastward and weakens to the west. The overprint relationships, as well as the absence of some western Acadian structures in rocks east of the South Newfane thrust, suggest eastward younging of Acadian structures. Two major deformational regimes are recognized corresponding to these two areas.

### Deformation Regime 1 (West of the South Newfane Thrust)

West of the South Newfane thrust the following structural features are recognized to characterize deformation regime 1. Structural features may be classed as predome and dome structures. Predome structures include in sequential development the following:

(1) Grenvillian gneissosity, folds and tightly folded compositional layering (judging from structures within the Green Mountain massif these structures are subvertical and trend generally NE-SW or NW-SE in broad arcuate belts). These structures are crosscut by igneous rocks of the Cardinal Brook Intrusive Suite, such as the Bull Hill Gneiss.

(2) Taconian schistosity ($S_1$) and thrust faults (fabrics $S_2$) west of the present area. Throughout all of western New England intense Taconian structures are present. As expressed in the Green Mountain massif, this is a northeast-trending east-dipping but refolded mylonitic fabric associated with northwest

directed thrust faults such as the Rake Brook shear zone (Fig. 2). Distinctive and strong downdip lineation (S65°E) is developed. Syntectonic muscovite associated with these shear zones at biotite and garnet grade retain $^{40}Ar/^{39}Ar$ cooling ages of about 400 Ma and are believed to be cooling ages from Taconic metamorphism (Burton and others, 1991). Within the present area $F_2$ structures are associated with major thrust faults that imbricate basement and the Hoosac Formation and Rowe Schist sequence (Fig. 2). These faults truncate folds having $S_1$ axial surface foliation. $F_2$ folds have curvilinear hingelines that become parallel with extension lineations and produce sheath folds. Rodded rocks in $F_2$ shear zones have SE to E apparent axes of rotation for minor folds of the schistosity. Garnet commonly statically overgrows these hinges in all major $F_2$ faults. This may be the explanation for some garnets showing apparent syntectonic growth about E-W axes. We believe this is Rosenfeld's (1968) early (Taconian to pre-Taconian) E to W rotation axis, but do not believe the garnet cores are Taconian or that syngrowth rotation is required to form these inclusions. Rocks of the Cardinal Brook Intrusive Suite (Harriman Reservoir Granite and Bull Hill Gneiss) in or near thrust faults exhibit intense mylonitization and intense rodding in the $F_2$ structures in the Rayponda-Sadawga dome, in the antiformal sheath folds at Jamaica, and in the Athens dome. Formline maps developed by plotting the distribution of $F_2$ schistosity and lineations have allowed us to map out the attitude of a Taconic reference surface in the domes (Figs. 4A and 5A). Cross sections of these maps allow us to accurately determine the amplitude and extent of post $S_2$ warping that produced the domal (and complimentary basinal) structural of Acadian age (Fig. 6).

(3) $F_3$ (Acadian) predome structure. $F_3$ folds are expressed by tightly appressed, overturned folds having a well-defined, penetrative schistosity ($S_3$). Outside the immediate dome areas $S_3$ trends north-northwest and varies from upright to strongly overturned to the southwest. It is highly folded in dome areas where it is broadly subhorizontal. Where $F_3$ folds cross $S_2$ foliation, a marked long-limb, short-limb asymmetry is produced giving rise to "Z" or "S" folds that tend to be consistent over large areas. On the west limb of the Athens dome, $S_3$ is folded along with $S_2$ into subvertical attitude. The amplitude of $F_3$ folds is low in comparison to $F_2$ folds and major $F_3$ closures are not recognized. $F_3$ surfaces form an excellent marker to define the internal structure in the Athens dome.

(4) Dome-stage folds. $F_4$ folds are intensely developed, northeast-trending, upright, easterly overturned to upright folds expressed by a strong biotite foliation or by a spaced crenulation cleavage. Fold forms are tight to open, rarely isoclinal and are relatively high amplitude, short wavelength features. Various orders of $F_4$ exist, from first order features such as the Sadawga antiform and the south plunging core of the Athens dome, to numerous smaller parasitic folds. $F_4$ folds appear to change orientation from N40-45°E-trending structures in the west to N20-15°E trends in the east. Apparent folding of $S_2$ reference surface by $F_4$ is intense, whereas $F_3$ surfaces vary from broadly folded to tight folds. $F_4$ is expressed over most of the area by a strong crenulation cleavage or weak schistosity. $F_4$ plunges are highly variable because of highly folded and refolded attitude of older reference surfaces. $F_5$ structures trend north-south and are upright open folds with no foliation but with a weak crenulation cleavage. $S_2$, $S_3$, and $S_4$ surfaces are broadly warped by $F_5$. Near the eastern border of the Athens dome $F_4$ and $F_5$ structures may merge. A set of N20°E-trending $F_5$ folds, having steep axial surfaces and northeast plunges, offset of the eastern margin of the dome in a series of normal, up-from-the-east sense of drag folds. These $F_5$ folds appear to be the same as the dominant fold in rocks east of the South Newfane thrust in deformation regime 2.

## Deformation Regime 2

In a narrow 1 to 2 km wide zone west of the South Newfane fault (SNF) overprinting of older folds ($F_2$, $F_3$, and $F_4$) by a north-south to N20°E trending set of folds is intense. These folds are tight and plunge uniformly to the northeast. At the South Newfane thrust, intense mylonitization occurs, together with abundant 5- to 10-meter-thick zones of highly disarticulated rocks. Locally the deformed zones resemble tectonic breccias. Rocks east of the South Newfane thrust contain principally one northeast-trending foliation. The uniformity of the structure is striking and contrasts entirely with the complex deformation scheme to the west. We ascribe this simplicity to absence of the $F_2$ (Taconian), $F_3$, and $F_4$ structures in Regime 2. As mentioned before, $F_5$ folds in the western unit may correlate with the dominant folds of the belt east of the South Newfane thrust. No structures identifiable as Acadian early nappe-stage folds of Rosenfeld (1968) were identified and

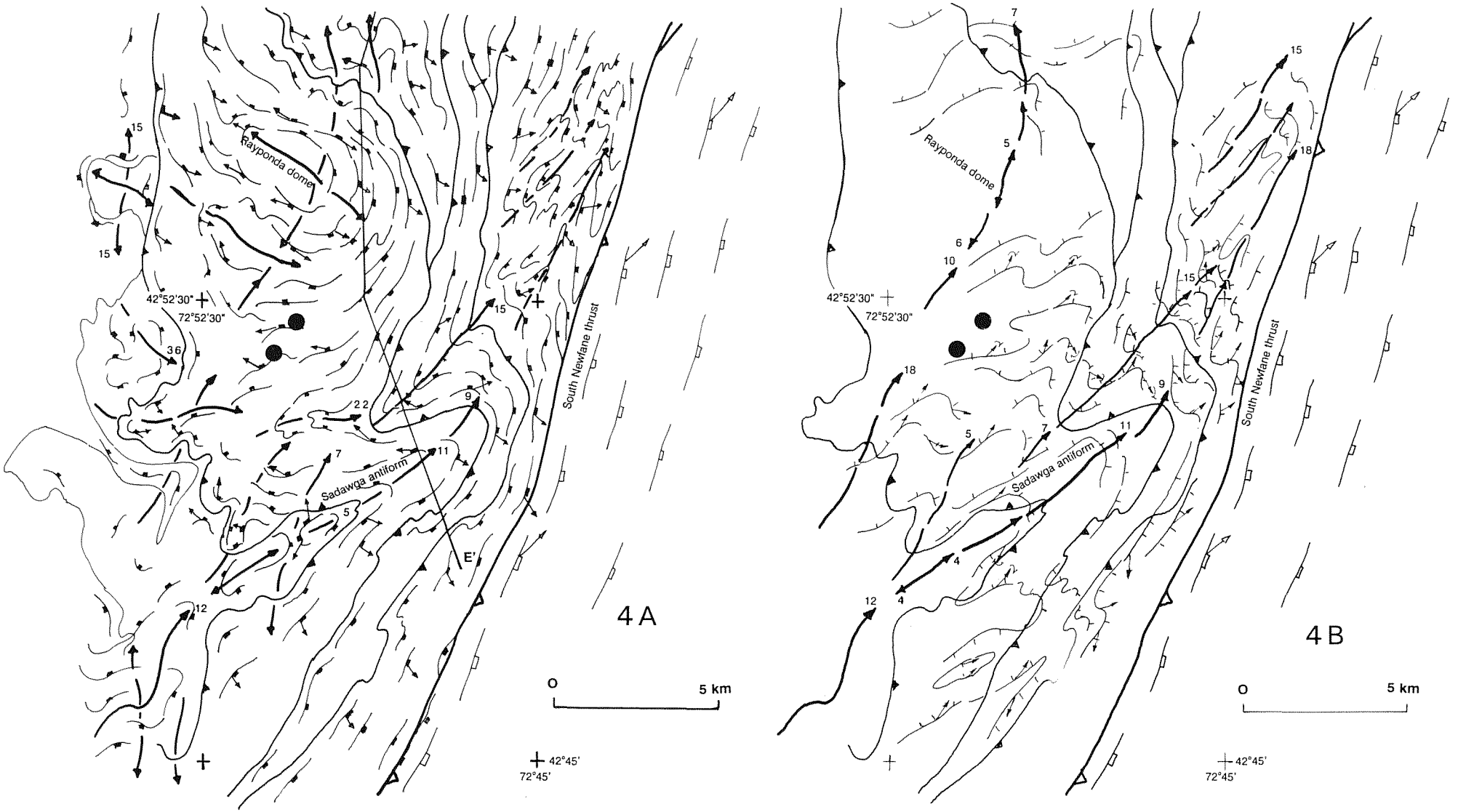



Figure 4. Formline maps of the Rayponda Sadawga dome showing folded pre-dome reference surfaces and axial traces $F_4$ and $F_5$ folds. Thrust faults identified as in figure 2. Dots show stop locations.
4A. Taconian $F_2$ axial surface map; small rectangle shows dip of folded ($S_2$) surface, solid arrow shows bearing of and plunge direction of $F_2$ fold hingelines. Strike and dip direction of $F_5$ axial surfaces shown by strikeline with open rectangle, $F_5$ formlines shown by dash-dot pattern.
4B. $F_3$ (Acadian) axial surface map, barb shows dip direction of axial surface. $F_5$ as identified above.

most importantly pre-dome $F_3$ folds in the Athens, Rayponda, and Sadawga domes are not the same generation folds as the dominant folds in the South Newfane lithotectonic unit, commonly ascribed to the nappe-stage folding.

## THE NAPPE- AND DOME-STAGE MODEL: WHAT CAN BE TESTED?

The nappe- and dome-stage structural model for the Athens dome predicts the existence of symmetrical repetition of Cambrian and Ordovician units in structures with Acadian axial surfaces (Fig. 7B). These structures should be present in the cover rocks on the west limb of the dome. Fold axes associated with the nappe-stage folds should be approximately NE-SW to N-S (Rosenfeld, 1968). dome-stage folds generated by diapiric rise of the core rocks should also have N-S axis with subhorizontal but variable plunges reflecting the doubly plunging form of the dome.

In the model, reverse drag or spruce-tree folds are caused by upward relative movement of the core rocks. Syntectonic garnets on the limbs of the rising dome may record that asymmetry in their inclusion trails, and this evidence from reverse-rotated garnets has been used to identify the dome-stage folds (Rosenfeld, 1968).

In a doubly plunging diapiric dome the upward motion of the core imparts rotational strain to passive markers on the limbs and nose of the fold. Actual axes of rotation vary in azimuth around the structure but are everywhere parallel to the general or sheet dip of the mantle. As long as the cover rock contains the rising mass, radial elongation occurs on the limbs as well as on the crest. The final or composite domal feature has plunging noses that have rotation axes normal to the geometric "axis" or crestline of the dome. It is important that the crestline (defined as the line on map joining high points of successive folded surfaces) is not the axis of physical rotation. This has an important effect on the apparent rotation and bearing of the folded nappe stage hingelines. In the model above, diapiric rise will rotate original horizontal N-S hingelines northward in an amount equal to the dip of the nappe-stage axial surface, that is the azimuth of the nappe-stage hingeline will be approximately parallel to the original azimuth. If the south plunge is equal to the amount northerly plunge hingelines of nappe-stage folds will plunge to the south off the dome. Rotation about an inclined axis may impart greater variation in azimuth of folded lineations.

A diagram (Fig. 7C) illustrates the effect of dome-stage folding expected around the southern closure of the Athens dome. The fold axis chosen, S25°W at 20°, is approximately parallel to the observed late plunges. The open-headed lineation arrow shows the expected orientations of originally horizontal nappe-stage hingelines following this folding. The solid-headed lineations show the expected orientations of a folded lineation inherited from the Taconian $F_2$ event, which was approximately N65°W. Rotation about the late axis does not satisfy the lineations observed. The hingelines lying within the folded $F_2$ and $F_3$ axial surfaces on the southern closure of the Athens dome (Fig. 5B) do not match in azimuth or plunge that predicted for rotated nappe-stage hingelines or rotated Taconian $F_2$ hingelines. A transection model could account for the preponderance of downdip lineations in $F_3$ surfaces if the transecting surface oriented approximately N50°W45°SW cut already folded or arched $S_2$ reference surfaces in the core of the dome. This may be the explanation for the $F_3$ lineations shown in figure 5B.

In addition to the spruce-tree (reverse drag) folds and the garnet rotation senses, the higher metamorphic grade (kyanite-staurolite) in the cores of the domes has been used to support the diapir model. Cross sections of the Chester and Athens domes (Doll and others, 1961; Rosenfeld, 1968) show an amplitude of 10 to 15 km between dome structures. Their sections were constructed utilizing the takeoff projections from the exposed cover contact up or down plunge and did not take into account variation in plunge along the axes of the domes. Because of this we feel that their amplitudes may be as much as three times too great. If the large amplitudes they portray are used, significant increases in pressure and temperature could be preserved in the cores of the domes as opposed to the basins.

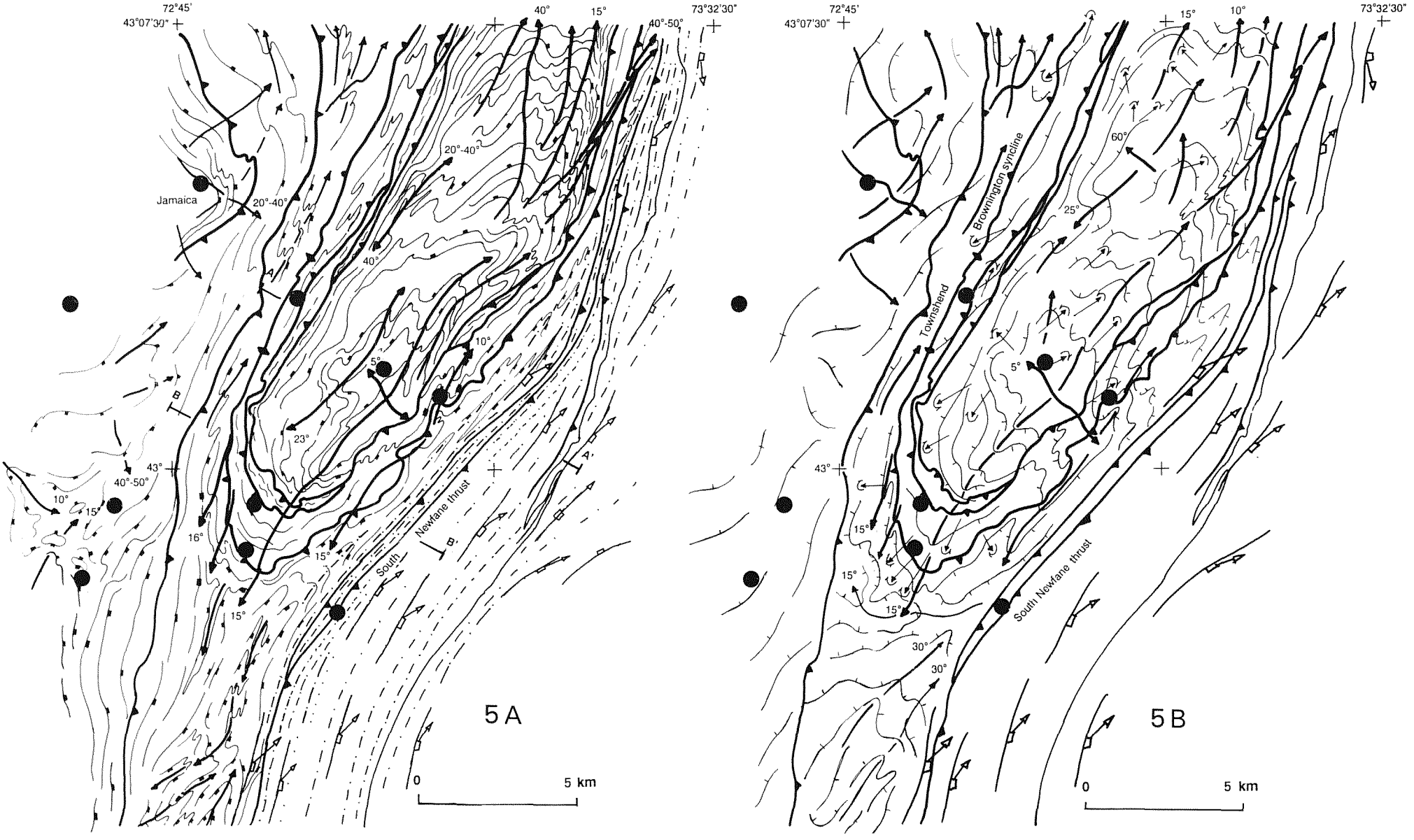



Figure 5. Axial surface formline maps of the Jamaica-Athens dome area to the north of figure 4. Thrust faults identified as in figure 2. Dots show stop locations.

5A. $F_2$ axial surface map, solid-headed arrows show plunge direction of Taconian ($F_2$) hingelines.

5B. $F_3$ axial surface map, open arrows and arcs show plunge direction and rotation sense of $F_3$ folds. Axial traces of post-$F_3$ folds identified by heavy lines; $F_5$ folds shown by strikeline with open rectangle.

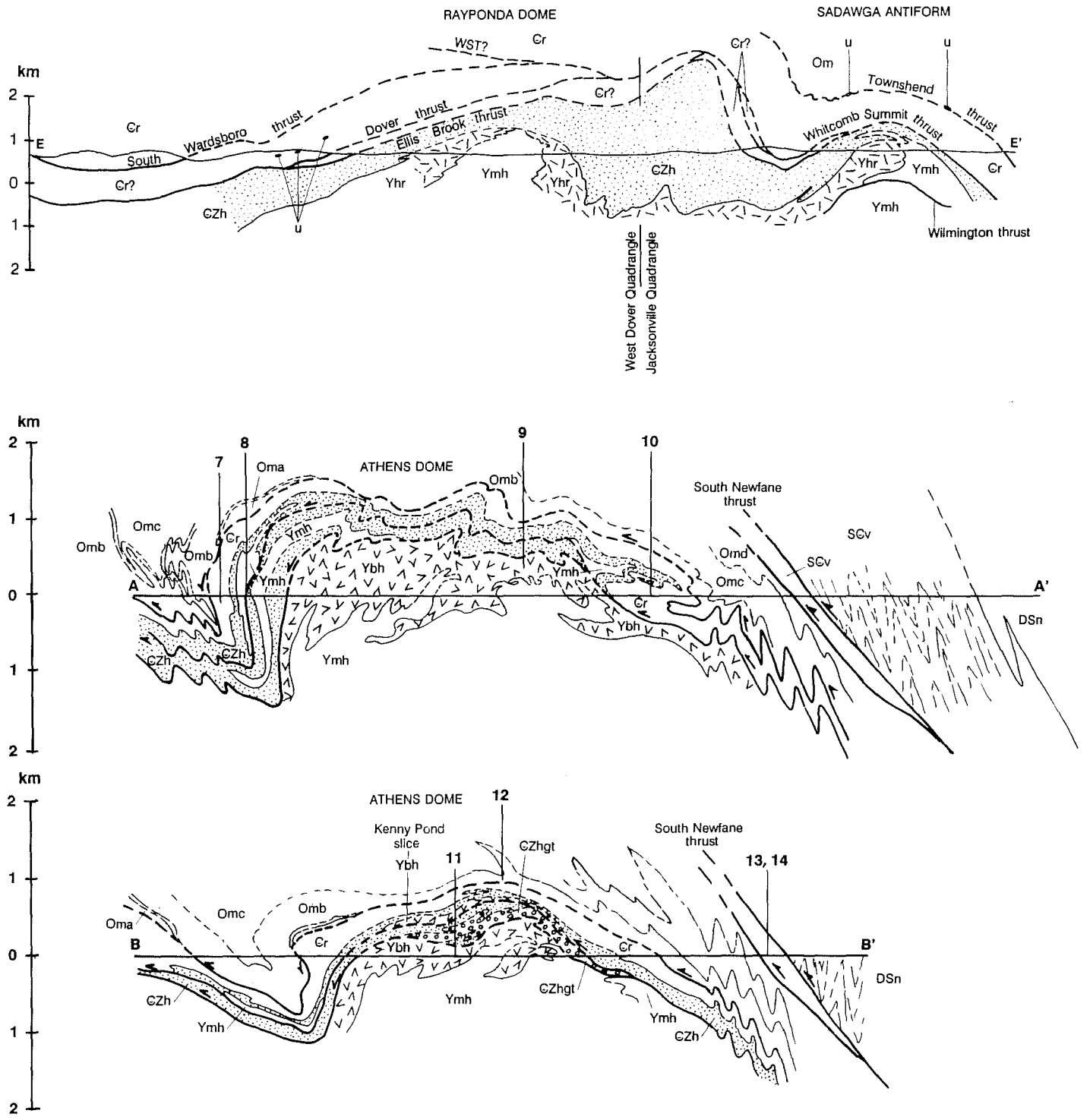



Figure 6. Simplified cross sections E-E' Rayponda and Sadawga dome, section line identified in figures 2 and 4; A-A', B-B' sections of Athens dome section lines identified in figures 2 and 5.

### Pressures and Temperatures Expected in Diapir Models

Several different models have been suggested for the tectonothermal development of Acadian domes in southern Vermont (Figs. 7A, B):

1. Diapiric uprise. Eskola (1949) envisioned dome development through diapiric uprise of hot, buoyant, polydeformed basement through denser, less deformed, cooler mantling strata in which remobilized basement may actually decouple from its root zone and actively rise through the overlying strata in a manner not unlike intrusion of a pluton through overlying country rocks (Fig. 7A). Diapirism in this manner would lead to heating and loading of the mantling strata and development of a thermal gradient with increasing T and decreasing P toward the core of the dome. While the distribution of actual temperature data, at least on the west flank of the Athens dome, are in accord with the distribution predicted by this model, the distribution of the actual pressure data is precisely opposite to that expected from dome development through diapirism.
2. Composite nappe and diapiric dome development (Fig. 7B). First presented by Thompson (1950) and later by Rosenfeld (1968), Hepburn and others (1984), and Rosenfeld and others (1988), this model suggests that west-directed Acadian nappes in this area, recognized by Thompson (1950) and Rosenfeld (1954), were gravitationally driven by diapiric uprise of hot basement rocks which gave rise to domes in a manner similar to that presented by Eskola (1949). This model differs from the diapir model presented above in that nappe-stage and dome-stage deformations are a continuous process, and the distribution of isograds, isobars and isotherms is the result of metamorphic processes operating continuously through both tectonics processes; the nappe-stage and early dome-stage deformation was termed "Diastrophism I" and post-nappe doming as "Diastrophism II" by Rosenfeld (Rosenfeld, 1968; Hepburn and others, 1984). Rosenfeld presented a tectonothermal model for this style of nappe and dome evolution in which isograds for specific lithologic units (and thus isotherms and isobars) dip westwards away from the structural high being established during diapiric uprise in Diastrophism I. A distribution of isotherms and isobars such as those presented in this model would lead to P-T estimates that would define a thermal gradient, for specific lithologic reference surfaces, increasing in value towards the east (towards the diapir), and a pressure gradient showing increasing P to the west (away from the diapir).

## INTEGRATION OF FIELD MAPPING AND METAMORPHIC STUDIES

### Method of Analysis

We have utilized a classical field-structural approach by mapping in detail the distribution of rocks and of the folded S surfaces and lineations contained in them as described above. We have identified and mapped multiple schistosities ranging in age from Taconic through Acadian and have developed from these data formline maps which describe the deformation of the sequentially developed reference surfaces. It is important to understand that the geometric form of a single structure such as a dome varies depending upon orientation of the cross folds, as well as the generation and orientation of reference surface being folded. Likewise, the crossing angles in strike and dip of older surfaces by younger surfaces control the azimuth, plunge, and apparent "drag sense" of each generation of structure. Garnets that grow within a matrix having a relict preferred asymmetry will include those structure regardless of whether or not they also rotated during growth. We have studied the textures in oriented thin sections collected from the entire area and noted the relationship of garnet growth to matrix textures in each foliation surface and fold set. We stress again, however, that the garnets studied by us either statically overgrow preexisting (Taconian and Acadian) microfolds or show enlargement to include contemporaneous kinking and crenulation folding in the matrix.

### Relative Age Relationships of Metamorphism and Structure

All Acadian deformation took place under garnet-grade or staurolite-kyanite grade regional metamorphism but the peak metamorphic conditions as determined by study of crystalloblastic textures and mineral chemistry varies across the area from west to east. In general, later-formed and coarser-grained porphyroblasts are found to the east than to the west and the inclusion fabrics suggest that lastest formed $F_5$ folds near or east of the

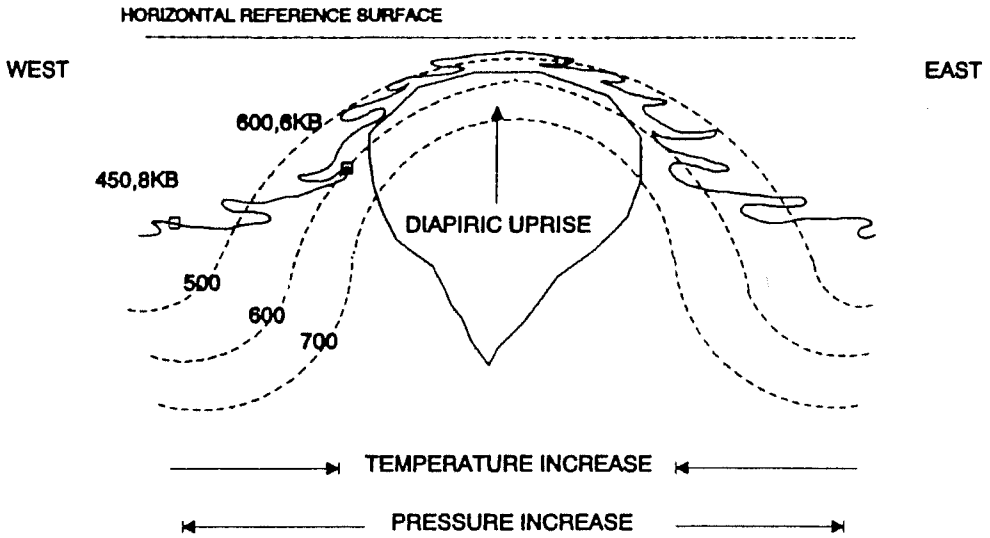
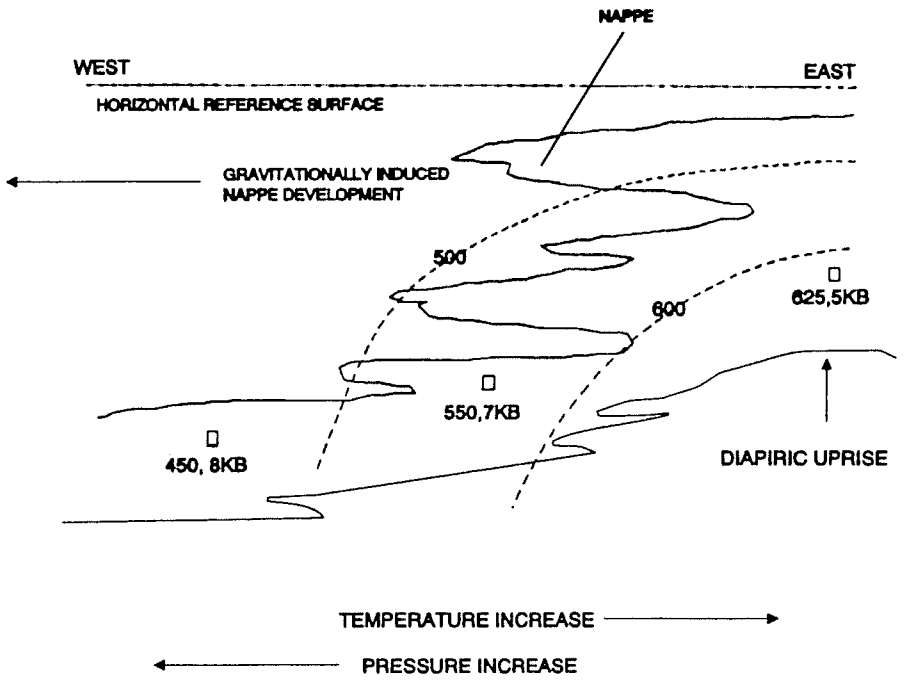



Figure 7A. Schematic figure showing expected synmetamorphic evolution in diapiric development of mantled gneiss domes (Eskola, 1949). Solid lines represent contacts, within the mantling, cover strata with asymmetric "spruce tree" folds and of diapir. Dashed lines are isothermal surfaces resulting from emplacement of hot basement rocks into cold cover; during advection isotherms are folded.




7B. Schematic diagram showing Acadian nappe and dome tectonothermal evolution for the Chester and Athens domes of southern Vermont (modified from Rosenfeld, 1968 and Hepburn and others, 1984). The portrayed isotherms are based upon Rosenfeld's garnet and kyanite-staurolite isograds. The P-T conditions shown are those expected using an arbitrary pressure of 8 kb for the western P-T point. Note the predicted decrease in pressure and an increase in temperature from west to east toward the core of the dome.

South Newfane thrust grew during peak metamorphic conditions there. To the west near the Rayponda and Sadawga domes, coarse garnets and biotites appear to be synchronous with $F_3$ and $F_4$ folds and no porphyroblast growth accompanied the $F_5$ fold event.

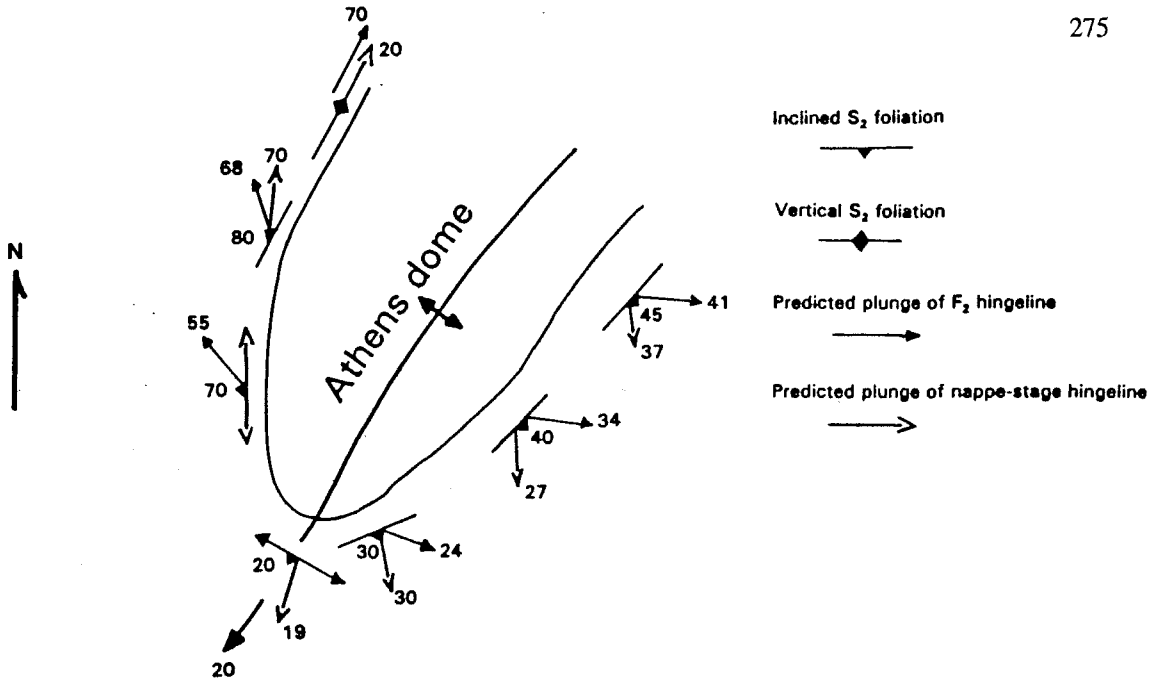
However, in rocks immediately west of the South Newfane thrust, where $F_5$ fold intensity increases eastward, garnet and hornblende include $F_5$ microfolds of older schistosities. These observations suggest that the metamorphism and structural deformation were not coeval across the area, but that peak metamorphic conditions developed relatively late in the structural evolution across all the area.

In the Rayponda-Sadawga and Athens domes, Acadian garnet-grade mineral assemblages developed synchronous with the Acadian $S_3$ and $S_4$ crenulation cleavages. False color x-ray images of oriented samples typically show Taconian $S_2$ foliation defined by aligned matrix minerals ilmenite, muscovite, biotite, and chlorite that is continuous with minerals defining inclusion trails within garnet porphyroblasts (Fig. 8). $S_2$ inclusion trails within these porphyroblasts show a continuous progression from undeformed within garnet cores to moderately crenulated at garnet rims. These folds are coplanar with tight, moderate to pervasive, either $F_3$ or $F_4$ fold generations defined by folded matrix micas and ilmenite. Subsequent cleavage development ($S_5$) postdates garnet growth and is defined by late chlorite and white mica that wrap around garnet porphyroblasts. $F_5$ deformation also appears responsible for local dissolution of garnet along $S_5$ cleavage surfaces normal to the maximum compressive stress direction. Such dissolution is recognizable by the weakly elongate shape of garnets in the $S_5$ orientation, and by "selvages" of ilmenite and rutile grains concentrated on the edge of garnets within the $S_5$ folia.

Many thin sections from this area contain two habits of biotite: 3 to 10 mm porphyroblasts that overgrow folded $S_2$ and a fine-grained biotite that defines $S_2$. Quantitative microprobe analyses indicate biotites are isochemical suggesting that Acadian metamorphism generally caused a chemical "readjustment" of Taconian minerals to compositions associated with the new intensive (P-T) conditions. Only limited occurrences of kyanite and staurolite have been found, largely in the rocks of the Rowe and Moretown south of the Athens dome. The rarity of these index minerals seems to be more related to the excessively calcic composition of the units rather than the regional P-T conditions. These calcic rock-types typically contain coarse-grained (10 to 50 mm) hornblende or cummingtonite porphyroblasts (or both) that form radial sprays or fascicles. Individual grains may have a weak alignment within the $S_2$ foliation in the $F_2$ elongation direction (~S65E) or in the intersection of $S_2$ with $S_3$ or $S_4$ ($F_3$ or $F_4$ hingelines). Amphiboles of either lineation-type are isochemical. Immediately west of the South Newfane thrust (SNT), hornblende lineations become progressively oriented into the $F_5$ elongation direction (~N45E). Hornblende-plagioclase-garnet gneiss east of the South Newfane thrust contains fine-grained hornblende (1 to 10 mm in length) that ranges in orientation from random to strongly lineated in a N45E direction parallel to the extension direction associated with the fault.

In rocks of the Moretown Formation near the South Newfane thrust, inclusion trails in garnets define crenulate folds that are coplanar with $F_5$ folds (northerly strike, nearly vertical dip) in the matrix. This overprint relationship can be seen in outcrops where $F_3/F_4$ axial surfaces are progressively transposed into the $F_5$ axial orientation towards the South Newfane thrust. These above observations suggest that garnet and hornblende growth at garnet, staurolite, and kyanite grades in broad areas west of, but near, the South Newfane thrust grew during or after $F_5$.

Rocks east of the SNT are generally deformed only by a weak $F_5$ deformation but strain along spaced mylonitic zones may be intense. Igneous rocks within the low-strain zones contain pristine igneous textures including hypidiomorphic, ophitic and subophitic texture, euhedral zoned plagioclase phenocrysts, weakly flattened to undeformed xenoliths, and fine-grained chill margins on the periphery of either hypabyssal sills or dikes. Many rocks contain only $F_5$ deformation; although a pre-$S_5$ layering is present in rusty schist horizons within the volcanics and in the garnetiferous phyllite of the overlying Northfield Formation as used by Doll and others (1961). This early layering ($S_1$ for rocks above the SNT) does not have the intense mylonitic fabric, reclined folds, nor SE trending elongation lineation typical of the Taconian $S_2$ found west of the SNTZ. Since no $F_1$ folds have been recognized within this region, we believe that this fabric is an early phase of bed parallel foliation related to either static mineral growth of primary compositional layering or a weak phase of early




7C. Simplified map of southern closure of Athens dome showing predicted attitude of folded $F_2$ lineation (N65°W) and nappe stage folds (NS horizontal) following folding.

Counting resolution: 6nm
Accelerating potential 20kv
Sample current: 20na
Data collected on wavelength dispersive spectrometers

**Low** **High**

Figure 8. False-color x-ray image of 1.13 cm garnet porphyroblast showing continuous Mn-zoning (core to rim decrease) and inclusion trails of $S_2$ foliation with nondeformed (planar) $S_2$ in core but increasing amplitude of crenulate folds toward the rim.

Acadian deformation. Inclusion trail geometries in garnet and plagioclase porphyroblasts, and alignment of fine-grained amphibole needles within the $F_5$ intersection on $S_1$ surface (low strain zones) and along the trend of the $F_5$ elongation lineation (high strain zones), suggest that peak thermal conditions were attained during $F_5$ event.

## THERMOBAROMETRY

### Thermobarometric Techniques

Estimates of peak Acadian metamorphic temperatures and pressures (Fig. 2, table 1) were determined using several exchange and net transfer geothermobarometers. Mineral compositions were obtained at Virginia Polytechnic Institute by quantitative microprobe analyses of porphyroblasts and matrix grains from several different lithologic types, including: quartz-muscovite-garnet-biotite-plagioclase-chlorite-ilmenite$\pm$rutile$\pm$staurolite pelite, quartz-garnet-muscovite-(Ca) paragonite-chlorite-staurolite-rutile-ilmenite$\pm$kyanite pelite, and garnet-chloritoid-chlorite-muscovite-(Ca) paragonite bearing pelites. Points analyzed were chosen after examinations of false-color-analogue x-ray images of porphyroblasts and matrix grains to eliminate potential disequilibrium assemblages. Validity of P-T estimates were checked by comparison with those predicted for the specific assemblages using R. Berman's GeOcalc phase-equilibria program (Berman, 1988). Only compositional data that produce a well constrained intersection of several different equilibria in P-T space and which agree with P-T estimates accrued from the thermobarometers were retained in this study. Mineral compositions were also subjected to a singular-value decomposition-matrix analysis (Fisher, 1989), which showed that reaction histories, especially for staurolite and kyanite bearing samples, were markedly influenced by rutile saturation, as originally proposed by Fletcher and Greenwood (1979). Such reaction histories include resorption of garnet, chlorite, and rutile during staurolite, biotite, ilmenite production as well as simultaneous consumption of staurolite and garnet during kyanite and biotite production. Mass-balance equations for the garnet consuming reactions involve Ca and Mn enrichment of the garnet edge along with concomitant consumption of muscovite and the margarite component of paragonitic white mica. False-color x-ray images showing concentrations of Fe, Mg, Mn, and Ca in garnets and of reacting phases were used to verify the choice of mass balance equations and general reaction history as proposed by Thompson and others (1977) and Armstrong and Tracy (1991). This exercise is necessary, especially when garnet resorption is involved, in order to select mineral compositions appropriate for equilibrium during peak thermal conditions. P-T results, locations and mineral assemblage of the various samples, and the exchange and net transfer thermobarometers used are summarized in Table 1.

### Thermobarometric Results

The calculated P-T data indicate that Acadian peak temperatures and corresponding pressures increase from west to east across the study area (Table 1, Fig. 2). These P-T data were acquired from widely different lithotectonic units and from many different structural positions in Acadian domal areas and basins. Regardless of the structural position the data indicate that rocks now exposed at the surface were deformed under greater lithostatic loads in the east than in the west. The pressure values suggest a thickening tectonic wedge eastward with minimum depths increasing from about 22 km in the west to about 32 km in the east. These pressures also correspond to increasing peak temperatures as well. Because porphyroblast inclusion textures indicate that peak metamorphic conditions occurred during the dome forming events in the Rayponda-Sadawga and Athens dome we can conclude that the domes themselves were formed within this thickened wedge at different depths in the crust. Because pressures and temperatures above were "frozen in" during decompression, the maximum depth of burial could be significantly greater than indicated by the thermobarometry.

A tectonic model which satisfies our data is given in figure 9. The schematic cross-section represents an instant in geologic time subsequent to $F_3/F_4$ development of the Rayponda and Sadawga domes and during $F_4/F_5$ development of the Athens dome. Solid lines represent reference lithologic contacts; heavy solid lines with arrows represent $S_5$ Acadian thrust faults and their relative motion senses, including the South Newfane thrust (SNT). Dashed lines portray the distribution of 500°C, 600°C, and 700°C isotherms. They are cut by faults in the western part of the section, but are roughly synchronous with them in the east, thus showing the west to east diachroniety of peak thermal conditions. The P-T localities refer to projected positions and values for the

**TABLE 1. SUMMARY OF THERMOBAROMETRIC RESULTS LOCALITIES IDENTIFIED ON FIGURE 2.**

| LOCALITY | METHOD | P(KB) | T(°C) |
|---|---|---|---|
| **1** | gar-bio[1] | | 476°±35 |
| (qtz-mus-gar-bio-plag-ilm-rut) | gar-plag-mus-bio[2] | 6.3-6.7 | |
| | gar-plag-rut-ilm-qtz[3] | | |
| **2** | gar-chl[4] | | 462°±60 |
| (qtz-mus-gar-chl-rut-ilm-gra) | | | |
| **3** | gar-bio | | 500°-535° |
| (qtz-plag-mus-bio-gar-chl-ilm-rut) | gar-plag-mus-bio | 7.1-7.5 | |
| | gar-plag-rut-ilm-qtz | 7.7 | |
| **4** | gar-bio | | 520°-540° |
| (qtz-plag-mus-bio-gar-chl-rut-ilm) | gar-plag-mus-bio | 7.3-7.9 | |
| | gar-plag-rut-ilm-qtz | 8.2 | |
| **5** | gar-bio | | 540°±50 |
| (qtz-plag-chl-bio-mus-gar-ilm-rut) | gar-plag-mus-bio | 7.6-7.9 | |
| | gar-plag-rut-ilm-qtz | 7.9 | |
| **6** | gar-bio | | 550°-575° |
| (qtz-plag-mus-gar-bio-sta-chl-ilm-rut) | gar-plag-mus-bio | 8.5 | |
| | gar-plag-rut-ilm-qtz | 8.3-8.7 | |
| **7** | gar-bio | | 580°-595° |
| (qtz-plag-mus-gar-hb-bio-chl-ilm) | gar-plag-mus-bt | 8.1 | |
| | gar-hb-plag[5] | 8.3 | |
| **8** | gar-bio | | 587°-647° |
| (qtz-musc-plag-bio-chl-gar-sta-ilm-rut) | gar-plag-mus-bio | 8.7 | |
| | gar-plag-rut-ilm-qtz | 8.9 | |
| **9** | gar-bio | | 625° |
| (qtz-mus-bio-chl-gar-plag-sta-ilm-rut) | gar-plag-mus-bio | 9.1 | |
| | gar-plag-rut-ilm-qtz | 8.9-9.3 | |
| **10** | gar-bio | | 610°-633° |
| (qtz-mus-chl-gar-sta-bio-kya-ilm-rut) | gar-plag-mus-bt | 8.7 | |
| | gar-plag-rut-ilm-qtz | 9.2-9.5 | |
| | gar-rut-ilm-kya-qtz[6] | 9.3-9.6 | |

**REFERENCES**

[1]Ferry and Spear (1978) with Berman (1990) garnet activity models
[2]Ghent and Stout (1981)
[3]Bohlen and Liotta (1986)
[4]Dickenson and Hewitt (1986)
[5]Kohn and Spear (1990)
[6]Koziol and Newton (1989)

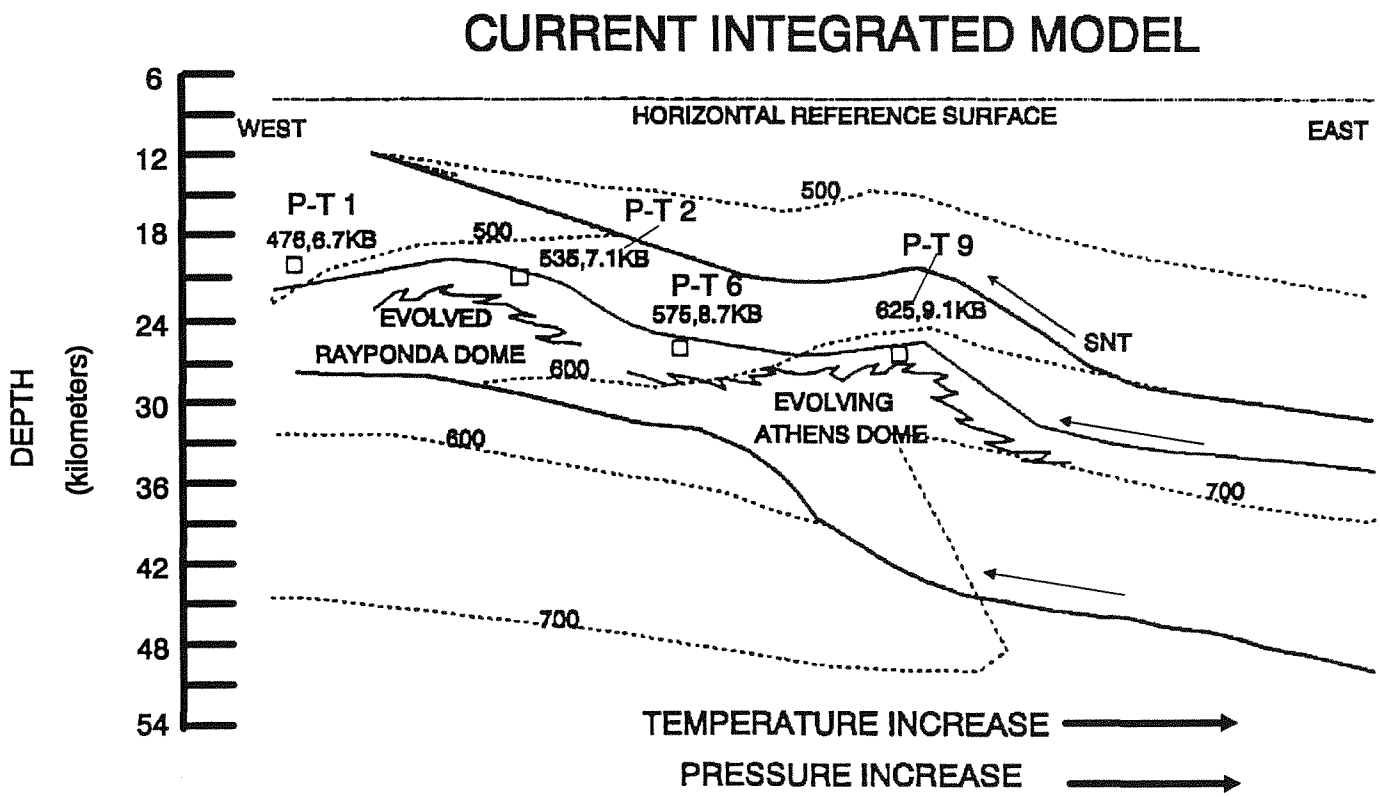



Figure 9. Simplified geologic cross section and thermobaric model for Acadian tectonic evolution in this area of southern Vermont, based on the field mapping and petrology conducted during this study. Solid lines represent generalized lithologic contacts and Acadian thrust faults (with arrows). Dashed lines represent reference isothermal surfaces. Note that truncation of isotherms by the faults in the west and slight folding in the east illustrates eastward younging of peak T conditions across this belt. Diagram portrays an instant between already evolved Rayponda and Sadawga domes at garnet-grade peak T conditions in the west, and evolving domes (Chester, Athens and Guilford) and near-peak T kyanite-staurolite grade conditions to the east. Labeled P-T conditions are the actual ones acquired during this study and are keyed to data from Table 1 and Figure 2. Notice that the data indicate an eastward increase in both P and T, which is in agreement with the spatial distribution of the thermobarometric data and in disagreement with P-T changes expected from the diapir models (Figs. 7A, 7B).

observed thermobarometric data given in Table 1. Notice that regardless of structural level, both peak temperatures and related pressures increase from west to east. Notice also that advection (upward physical transport of heated rocks) is required to satisfy the high temperatures observed, especially in the Athens dome. Because peak metamorphic conditions correspond to the formation of the domes, we conclude that the various domes formed at different crustal levels in a west to east succession in this region.

## CONCLUSIONS

Based on our structural and stratigraphic studies we conclude that the Sadawga, Rayponda, and Athens domes formed by interference of multiple noncoaxial and noncoplanar Acadian fold events. The tectonic cover of the domes consists of highly thrust-faulted mylonitic basement gneiss and cover rock in multiple thrust sheets that formed originally in the Taconic orogeny. This pre-Silurian tectonic cover contains reclined sheath folds having axes subparallel to the prominent Taconic extension lineation of the N65°W. We find no evidence in these pre-Silurian rocks for the NE-SW to NS recumbent nappe stage Acadian folds proposed by Rosenfeld (1954, 1968) and by Thompson (1950) to pre-date Acadian diapiric uprise of the domes.

Acadian dome-stage (and basin stage) folds, rather than having dynamically controlled rotation senses derived from upward rise of core rocks in the domes, appear to be caused by transection of older inclined reference surfaces by $F_3$, $F_4$, and $F_5$ Acadian axial surface foliation and crenulation cleavage. Throughout all of the area garnet inclusion structures do not require syntectonic rotation in either pre-dome or dome stage folds but inclusion geometries are explained by overgrowth of microfolds of older schistosities produced by $F_3$, and $F_4$ folds. Modulated wave trains of folded schistosity that varies from undeformed in the core to plicated at margins of garnets indicates to us that garnet over most of this area grew during the $F_4$ event. This late growth of garnet may explain the nonrotation of garnet inclusion trails seen by Hayward (1990).

Determination of the peak T and P conditions during Acadian metamorphism from the pre-Silurian rocks indicates an eastward increase in pressure corresponding to changes in depths at the peak of metamorphism from 22 km on the west flank of the Rayponda-Sadawga dome to 32 km on the crest of the Athens dome. Amplitudes of the Athens and Sadawga domes are moderately low (less than 3 km) and are not reflected in the pressure data. From our maps of the Taconian and folded Acadian foliations in the cores of the domes, we suspect that amplitudes shown by others (Doll and others, 1961) for the domes are far in excess of the actual amplitudes required. This probably explains in part the poor correspondence between the structure of the domes as seen in seismic reflection data (Ando and others, 1984) and that shown by Doll and others (1961).

Finally we suggest, following Ratcliffe (1990), that the Acadian domes formed as result of reactivation of a Taconian thrust stack in which inherited northeast trending ramp- and tipline-fault geometry controlled the northeast trends and amplitudes of the $F_4$ folds, whereas $F_3$ and other northwest trending folds were produced by Acadian compressional folds initially formed on Taconian flats and gently dipping ramps. The increase in pressure eastward during Acadian $F_4$ and $F_5$ deformation is most likely caused by westward encroachment of the Bronson Hill terrane of Massachusetts and New Hampshire (Armstrong and others, 1992). Importantly, we believe that deformation and peak metamorphic conditions young eastward towards that advancing terrane and that structures also young eastward from the Athens dome. Integration of structural and quantitative P-T information requires a tectonothermal model that portrays development of an eastwardly increasing thermobaric gradient, prior to the end stages of Acadian deformation. Younger peak thermal conditions from west to east, and rapid uplift of the terrane during dome evolution helped preserve the pristine prograde assemblages and create the thermochronologic age distribution presented by Sutter and others (1985), Sutter and Hatch (1985), and Laird (1991). The preserved younging of peak thermal conditions from west to east, as recognized by petrographic and mesoscopic structural analysis, is the result of rapid uplift and cooling of rocks during dome development. Because the domes are of insufficent amplitude to cause such an effect, it is necessary to implement an alternative mechanism for uplift, such as thrusting. Movement of rocks westward over a pre-existing east dipping ramp would not only create disharmonic folding and dome development, it would also lead to *en bloc* uplift of the entire package moving over the ramp structure and related eastward younging of peak T conditions for rocks at progressively deeper crustal levels. Such faults may include the South Newfane thrust and the Searsburg fault system which surfaces within the Mt. Snow and Readsboro quads on the immediate east flank of the Green Mountain massif (Ratcliffe, 1992).

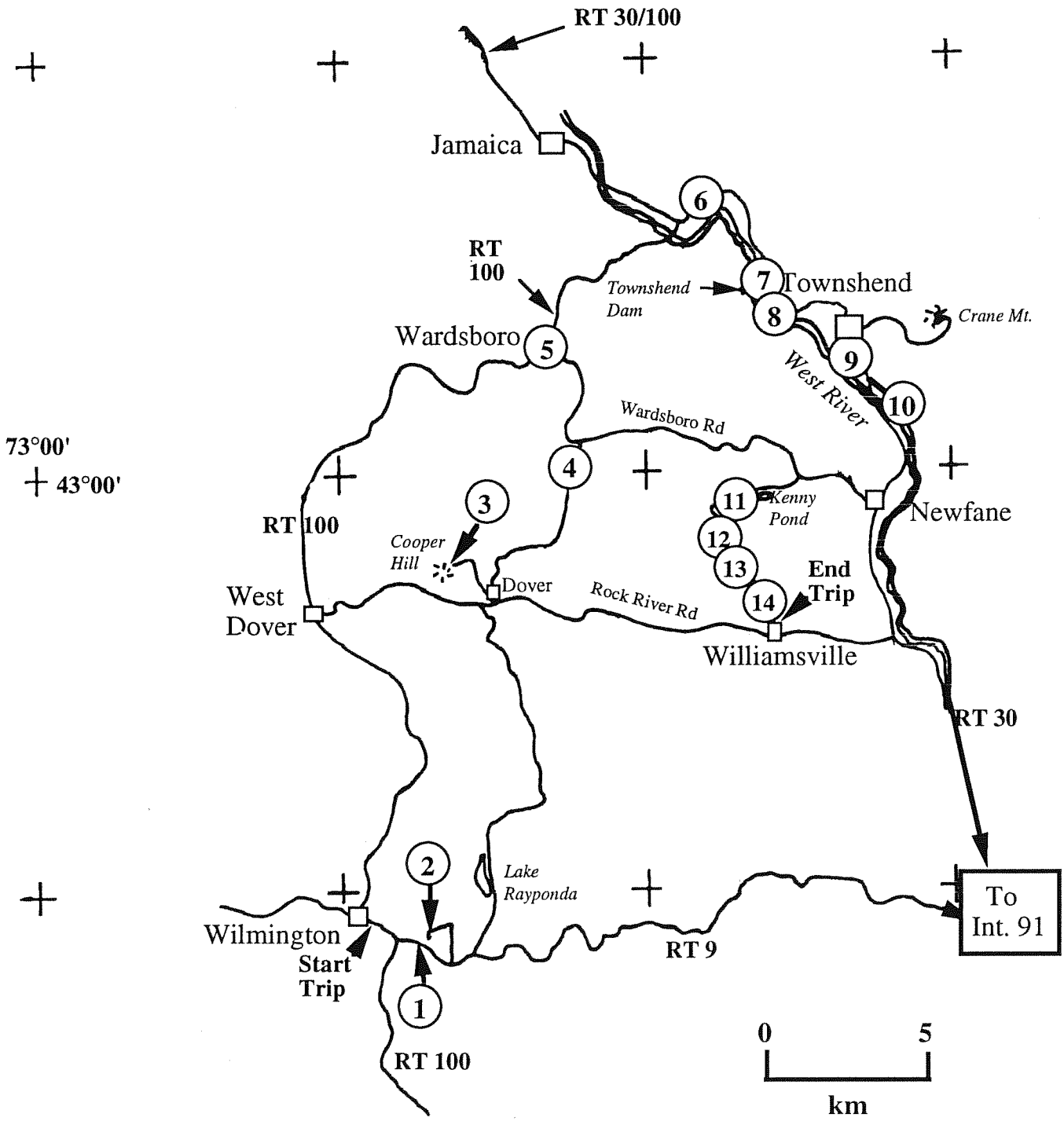



Figure 10. Field trip route showing stops.

## ROAD LOG

Starting point Grand Union Parking lot, Route 9, 0.5 miles east of Wilmington, VT.
Location of field trip stops identified in figures 2 and 10.

**Mileage**

0.0 Turn east onto Route 9.
0.4 Intersection Routes 9 and 100, bear left.
0.6 Cemetery on left, post on right shoulder.

**STOP 1. PROTEROZOIC CORE ROCKS AND HOOSAC FORMATION, NORTH FLANK OF SADAWGA DOME** (15 MINUTES) (Jacksonville Quadrangle). Stop shows basement cover contacts typical of the domes. Cross road walk 100 feet N20°E to first crop in woods of tectonically laminated gneiss (Harriman Reservoir Granite) showing large reclined $F_2$ folds and younger $F_3$ crossfolds. Ledges immediately above are biotite-albite-quartz granofels and pebbly quartzite of the overlying Hoosac Formation. Note structural conformity of foliation in both rocks at contact and the similarity in $F_2$, $F_3$, and finally $F_5$ late upright folds to those in the basement. Basement and cover share the same fabric elements. Amphibolites of the Turkey Mountain Member of the Hoosac start in cliffs a short distance to the north.

Continue east on Route 9 to Ballou Road on left.
0.9 Turn left on Butler Road, which becomes dirt
1.3 Turn left on River Road and wind way slowly up hill
1.8 Park in drive to left just opposite low cliffs on right

**STOP 2. ALBITIC GRANOFELS OF HOOSAC FORMATION HAVING $F_2$, $F_3$, and $F_4$ GENERATION FOLDS** (15 MINUTES) (Jacksonville Quadrangle). This stop shows typical deformational elements of cover rocks here on the east-west north-plunging border of the Sadawga dome. Small south-facing cliffs show nicely the typical Hoosac Formation and well-developed "Z" style folds of $S_2$ ($F_3$ folds) and gentle warping of $F_3$ axial surfaces by $F_4$ folds. The patterns seen here are developed throughout the Jacksonville and West Dover Quadrangles, where domal and basinal structures result from noncoaxial refolding. Importantly, the foliations, cleavages and crenulation cleavages can be mapped, and from these data axial trace or form-line maps constructed which elucidate the geometry and evolution of the dome structures (see Fig. 4). Section E-E' (Fig. 6) shows a cross section through the Rayponda dome and Sadawga antiform illustrating our interpretation of the structures.

Turn around and return to Ballou Road.
2.3 Turn right on Ballou Road heading back to Route 9.
2.2 Turn left on Route 9.
4.1 Turn left on Lake Rayponda Road, crops east or west of road are schists, granofels, and amphibolites of the Hoosac Formation in a complex syncline and basin between the Sadawga and Rayponda domes. Continue past Lake Rayponda.
6.4 At north end of lake pavement begins and pass over contact between core rocks of Rayponda dome to north and south-dipping Hoosac. $S_2$ foliation frames the dome and depicts the internal structure of the domes (see Figure 2).
7.0 T intersection with Higley Hill Road. Turn right. We are now heading east toward the eastern border of the Rayponda dome.
8.8 Turn left onto dirt road at triangle, just after Miller's Salvage on right.
11.6 T intersection with dirt road (Captain Copeland Road). Turn left.
11.9 Intersection with paved road (E. Dover Road). Turn right, but be prepared to turn left a short distance down the hill.
12.1 Turn left on Dover Common Road.
12.2 Y intersection. Bear left opposite library.

12.3 Hill on left exposes greenstone, amphibolites having abundant inclusions of talc-carbonate pods rocks of the Rowe Schist all structurally above the Hoosac Formation on the east flank of the Rayponda dome. Abundant thrust faults of $F_2$ age (Taconian) shred the cover rocks along the eastern flank of the Rayponda and Sadawga domes.

13.1 New roadcuts of dark gray rusty-weathered chlorite-muscovite-plagioclase quartz$\pm$biotite$\pm$garnet schists at Cooper Hill.

13.3

**STOP 3. COOPER HILL EXPOSURE AND REGIONAL VIEW OF TOPOGRAPHIC FEATURES BETWEEN THE RAYPONDA AND SADAWGA DOMES, ON THE SOUTH AND THE ATHENS DOME ON THE NORTH, MAJOR TECTONIC UNITS, DISTRIBUTION AND THEIR STRUCTURAL ELEMENTS** (30 MINUTES) (West Dover Quadrangle). Roadcuts of phyllite of the Rowe Schist(?) as exposed on and near Cooper Hill. Main fabric in schist is a strong anastomosing $S_2$ foliation spatially associated with thrust faults that imbricate and wedge out many units of the Rowe Schist between here and the exposures of the type Rowe 20 km to the south in Massachusetts. This area is on a major zone of $F_2$ thrust faults (Taconian) related to the Whitcomb Summit thrust of Massachusetts that places rocks of the Rowe Schist on rocks of the Hoosac Formation. The fault zone is thought to be the best approximation of the root zone for the Taconic allochthons.

13.3 Turn around and head down hill back to Dover Commons.

14.3 Turn left onto Holland Road.

14.9 Small crops by barn of coarse-grained amphibolite metagabbro(?) in Rowe Schist.

15.8 Intersection at bridge over Rock River, turn left into downtown Goose City (1 or 2 houses) immediately turn at red house on left taking right "Y".

15.9 Immediately (a short distance uphill) turn right and follow road uphill through pastures containing crops of biotite-quartz-feldspar schist and granofels characteristic of the upper Rowe Schist in other areas (large equivalent to rocks present in the Stowe Formation of central Vermont).

16.8 Intersection turn left on Wardsboro Road

17.0 Continue north past Jockey Hollow Road

19.4 **STOP 4. ЄRGS-ЄRA CONTACT AND DISCUSSION OF THE CHESTER AMPHIBOLITE PROBLEM** (15 MINUTES) (West Dover Quadrangle). Crops on west side of road include a coarse-grained ilmenite-biotite-garnet-muscovite quartz$\pm$hornblende schist (Єrgs) having coarse garnets with static inclusion fabrics of deformed $S_2$, and showing syntectonic growth on $F_4$ crenulations. Coarse garnets contain undeformed $S_2$ inclusion fabric in cores with progressive modulation (crenulation) by Acadian $F_4$ within rims (see Fig. 8 for similar structure). This large-garnet schist is in fault contact with the underlying Rowe Schist(?) that contains discontinuous amphibolites (Єra), and other lithic types. When mapped in detail these units truncate against upper and lower plates in South Wardsboro thrust zone and the amphibolite shown by Doll and others (1961) as the Chester Amphibolite does not form a continuous unit. Thermobarometric analysis of Єrgs-Єra rocks near this locality (Table 1, P-T locality 4) using garnet-biotite thermometry (Ferry and Spear, 1978 with Berman, 1990, garnet activity models) and garnet-plagioclase-muscovite-biotite barometry (Ghent and Stout, 1981) has yielded P-T (at peak T) conditions of 520-540°C, 7.3-7.9 kb (8.2 kb for garnet-plagioclase-rutile-ilmenite-quartz; Bohlen and Liotta, 1986).

19.8 Bear left at fork in road.

20.3 Stop sign turn left onto South Wardsboro Road, past west over thinned and faulted Єra (Chester Amphibolite of Doll and others, 1961)

20.8 Downtown South Wardsboro, turn right black arrow and downhill

23.6 Bridge over Wardsboro Creek park on right.

**STOP 5. MYLONITIC PROTEROZOIC GRANITE (HARRIMAN RESERVOIR GRANITE) NEAR WILMINGTON THRUST FAULT - TECTONIC COVER OF GREEN MOUNTAIN MASSIF** (15 MINUTES) (Jamaica quadrangle). Mylonite gneiss formed from coarse-grained Middle Proterozoic granites of the Cardinal Brook Intrusive Suite (960 Ma) record only Taconic and Acadian structure because these intrusive post-date Grenville deformation of the Mount Holly Complex. The well-layered $F_2$ tectonic fabric (Taconian)

here is the result of fault strain common in slivers of basement and cover that mantle the east side of the Green Mountains. Reclined folds having sheath fold geometry characterize the zones of faulting. The tectonic layering in this rock is the probable explanation why previous workers regarded these augen gneisses as metamorphosed rhyolites in the cover sequence (see Rosenfeld in Hepburn and others, 1984). In the Athens dome, this rock is called the Bull Hill Gneiss.

23.6 Turn right on Route 100
25.2 Low crops to the left in Wardsboro Creek of Harriman Reservoir Granite (augen gneiss dated by U-Pb zircon as 960 Ma (Karabinos and Aleinikoff, 1988)).
28.1 Intersection Route 30, turn right.
28.3 Outcrops start on left, park right shoulder.

**STOP 6. BASEMENT GNEISS IN CORE OF MIDDLE LEVEL OF TECTONIC COVER OF GREEN MOUNTAINS NEAR JAMAICA-SHEATH FOLDS AND $F_2$ FABRIC** (20 MINUTES) (Jamaica Quadrangle). This granite gneiss forms the core of the Jamaica sheath fold, the east plunges seen here are the hingelines of the $F_2$ generation folds in basement and cover rocks. This granite intrudes the Mount Holly Complex and is a fine-grained equivalent of the Harriman Reservoir Granite seen at STOP 5. Cross sections through this area depict the structure as the sheath folds of the tectonic cover that will reappear as the mantle the Athens dome to the east (see figure 3).

29.6 West Townshend
30.0 Exposures of the Moretown Formation in Townshend Brownington "syncline." Mapping here and elsewhere in southern Vermont shows that the Moretown units are discordant to the base of the Moretown and that sub-Moretown rocks are truncated by the base as well. We interpret the Moretown to be in fault contact with the underlying rocks on a Taconian thrust fault. The repeated Middle Proterozoic rocks, Hoosac Formation, Rowe Schist, and Moretown together form the highly thrust-faulted tectonic cover of the Athens dome.
32.0 West Townshend Dam and large roadcuts of Moretown and Rowe Schist.
32.4 Take hard right onto dirt road just past Army Corp of Engineers Headquarters.
32.2 Park by utility building in parking lot.

**STOP 7. AMPHIBOLITES AND LARGE GARNET SCHISTS OF THE ROWE SCHIST AND DISCUSSION OF GARNET INCLUSION FABRICS--SYN GROWTH ROTATION OR LATE OVERGROWTH OF COMPLEX MATRIX** (30 MINUTES) (Townshend Quadrangle). Walk to dam outlet and crops of amphibolite and large-garnet, staurolite-bearing schist behind salmon transfer station. An exposure of highly folded amphibolite having steep subvertical fold axes as well as shallow north plunging later $F_4$ folds. Interference patterns describe unusual basins and domes found throughout the rocks of the Townshend Brownington syncline. From these crops walk east to water-polished ledges of large-garnet schist at fish tanks. These huge garnets have clear inclusion fabrics showing apparent clockwise, counter clockwise rotations as well as inclusions of the folded geometry present in the matrix. We think these garnets overgrew microfolds of the matrix produced by $F_2$, $F_3$, and $F_4$ hinges but that the same garnets may contain modulated wave trains of the $F_4$ or younger crenulate cleavages. From these data, we conclude that the garnets completed most of their growth during the $F_4$ or the $F_5$ events. $F_5$ post-garnet kinking of matrix at the rim of the garnet is quite common and can be seen to continue into the matrix.

Our interpretation of the garnet, therefore, differs from that of Rosenfeld (1968), who believed the garnets record synrotational growth, and that they have rotation senses that can be used to predict structural positions on major nappe and dome stage structures. We also do not entirely agree with Bell and Johnson (1989) or with Hayward (1992) that garnets grew early enough to preserve structures no longer preserved in the matrix, indeed as can be seen in this outcrop the garnets include fabric elements, fold hinges, deflections of schistosity still present in the matrix. We believe that garnet growth post-dated $F_2$ (Taconian) and $F_3$ deformation, but was synchronous with $F_4$ and possibly $F_5$ deformation. According to our scheme, the garnet can not preferentially exhibit structures older than $F_3$ because they were not present prior to $F_3$.

False color x-ray images of garnets from staurolite-absent pelites at this locality show monotonic chemical zoning similar to Stop 4. Bell and Johnson (1989) proposed in a radically new tectonic model that garnet growth in this area occurred during several discontinuous metamorphic events, each punctuated by a period of deformation induced garnet dissolution. Resultant changes in P-T would cause drastic changes in the reaction history which should be observable in garnet x-ray images as reversals or irregularities in cation concentrations; no such features are observed in the staurolite-absent pelites. Garnet x-ray images from pelites containing staurolite do show a single period of garnet resorption, evidenced by increases in Mn and Ca concentrations in an irregularly distributed concentric pattern rimward of which staurolite inclusions are first found and chlorite inclusions are absent. However, in contrast to the interpretations of Bell and Johnson (1989), $S_2$ inclusion fabrics are continuous across this boundary which Armstrong and Tracy (1991) interpret as the result of the staurolite-in reaction.

32.6 Scott Covered Bridge park on right

**STOP 8. PROTEROZOIC GNEISS IN TECTONIC COVER OF ATHENS DOME AND EXPLANATION OF ASYMMETRY DOME STAGE FOLDS** (20 MINUTES) (Townshend Quadrangle). Crops in road cut and by west river are microcline biotite granitic gneisses similar to those seen at STOP 6. Basement gneisses and Hoosac Formation form several repeats on the west limb of the Athens dome duplicating the relations in the Jamaica area. These belts are repeated by faulting and by folding giving rise to the 1 km-thick tectonic cover of the Athens dome.

In these outcrops, east-up asymmetric folds of gneissosity are clear and could be used to support the diapiric rise of the core of the dome as suggested by Rosenfeld's model. Here, however, it can be seen that the apparent rotation sense is produced by the angle of intersection of gneissosity (N20-30°E subvertical) with a N40-50°E west-dipping spaced cleavage either in $S_3$ or $S_4$. We find a consistent asymmetry produced by $S_2/S_3$ intersections in the area that suggest incorrectly, up-from-the east sense of motion of the core gneisses.

32.6 Continue east on Route 30 into Townshend.
34.1 Townshend, continue on Route 30.
34.5 Harmonyville, turn right onto Depot Road, signs to Kamporama. Follow road into camp and park at base, cliffs to right.

**STOP 9. CORE GNEISSES OF THE ATHENS DOME AND INTERNAL STRUCTURES ON "AXIS" OF DOME. DISCUSSION OF FOLDED $S_2$, $S_3$ REFERENCE SURFACES** (15 MINUTES) (Townshend Quadrangle). These cliffs of Bull Hill Gneiss contain excellent subhorizontal $F_3$ axial surfaces typical of the core gneisses of the dome as well as later steep-trending $F_4$ folds that produce the north plunges. The axial- surface formline maps of the Athens dome (Figure 5) show the curvilinear pattern of $F_3$ folds as well as the variable plunge of the $F_4$ folds. The $F_3$ reference surfaces have been affected by post $F_3$, but pre-$F_4$, warping along northwest trending folds. This porpoising effect on the axis of the dome is important because it reduces the height of the crest of the dome in cross section. The $S_3$ reference surface and the $F_4$ plunge reversals require a much lower amplitude structure for the dome than shown on Doll and others (1961) (see section A-A', B-B', Figure 6).

34.5 Log resumes at Route 30 and Depot Road. Turn right on Route 30.
35.0 Plumb Road on left and route to alternate STOP 10 at Crane Mountain.
35.3 Turn left on dirt road before bridge over West River.
35.7 Park on left side of road at entrance of small logging road

**STOP 10. EAST LIMB OF ATHENS DOME FAULTED CONTACT AND ORIGIN OF WEST-SIDE-UP FOLDS** (40 MINUTES) (Townshend Quadrangle). This stop consists of four substops A, B, C, and D. From cars, walk 150 ft west to small cliffs of Bull Hill Gneiss (10A). Mylonitic Kspar biotite gneiss with strong downdip ($F_2$) lineation is deformed Bull Hill Gneiss. The high strain in the rocks is the result of thrust faulting that cuts out almost all of the Hoosac Formation beneath the overlying amphibolites of the Rowe Schist exposed at the parking spot. To the north, this contact branches into a series of thrust faults that duplicate the

core gneisses and Hoosac Formation on the east flank of the Athens dome at Crane Mountain.
10B - Walk from cars to exposures in the riverbank of amphibolites and garnet plagioclase schist. Folds of $F_3$ generation show consistent up from the west asymmetry and north plunges.
10C - 200 ft east on the north side of road is exposure of amphibolite and rusty weathered large-garnet schist of the Rowe Schist in a moderately large fold. The axial surface of the $F_3$ fold is east-west trending and north-dipping but shows the same rotation sense (clockwise) as at 10B. At the east side of the garnet schist note the steep axial surface of the $F_4$ fold responsible for arching of the $S_3$ reference surface.

Walk 250 ft east to crops of amphibolite in the river. STOP 10D. In these exposures the $F_3$ axial surfaces are folded by northeast trending $F_4$ folds having a strong foliation. These stops illustrate the overprint relations of $S_2$, $S_3$, and $S_4$ characteristic of the eastern limb of the Athens dome. The asymmetry of the $F_3$ folds (up from the west) are controlled by the crossing angle between $S_2$ and $S_3$ and are not indicative of west-side up motion during doming.

Turn around and head back to Route 30.
36.1 Turn left on Route 30.
39.3 Newfane, General Store, turn right by Newfane Inn.
39.4 Turn right
41.5 Sign for Grout Road, turn left.
41.9 Y intersection, bear left.
42.5 Kenny Pond on left.
43.0 Intersection, continue straight.
Wind around hairpin turn to left and
43.4 Park on right shoulder of road, just before house on right side of road.

**STOP 11. CRESTAL AREA OF SOUTH PLUNGING ATHENS DOME--TECTONIC REPEAT OF BASEMENT (BULL HILL GNEISS) AT SOUTHERN CLOSURE OF ATHENS DOME** (20 MINUTES) (Newfane Quadrangle). Exposures in stream to north and in cliffs west of road are highly deformed Bull Hill Gneiss caught in a fault sliver above one belt of the Hoosac Formation. This fault repetition duplicates relations seen at STOP 8 and that referred to at STOP 9, but here located on the crest or the dome. The basement repeat mantle the entire dome. Crops nearest the cars farthest to the south consist of albite Hoosac, quartzite and minor quartz-pebble conglomerate.

The contact between Bull Hill Gneiss and the Hoosac is exposed in the cliffs north of the road although the actual contact is difficult to locate because of the grain size reduction in the Bull Hill Gneiss.

Continue to south
43.9 Hairpin turn to left; do not take road to right.
45.1 Bridge crossing brook part in drive headed to cabins on west (right side of road)

**STOP 12. ROWE SCHIST AMPHIBOLITE AND STRUCTURES ON DOME CLOSURE AND APLITE DIKES** (20 MINUTES) (Newfane Quadrangle). Exposures under bridge are schists and amphibolites in the Rowe Schist. Axial surfaces of $F_2$ folds dip south and plunge southeast, $F_3$ folds are subparallel to $F_2$, but generally dip south more gently than $S_2$. Resultant $S_3/S_2$ intersections produce counter clockwise rotation for folds plunging to southwest. The two foliations $S_3$ and $S_2$ and their downdip lineations are both folded around the southern closure of the dome by northeast-trending $F_4$ folds. $F_4$ is expressed by broad to tight folds having subvertical to northwest-dipping axial surfaces and by a cleavage that varies from a weak crenulation cleavage to a strong foliation. $F_5$ folds in a north-south subvertical orientation are locally present, but the real closure of the dome is controlled by southwest plunging $F_4$ folds. Downstream from the bridge are exposures of coarse-garnet, rutile-ilmenite-plagioclase-staurolite-biotite-chlorite-white mica-quartz schist in the Rowe Schist, that can be traced around the southern closure of the dome. P-T conditions (at peak T) are estimated at 625°C, 8.9-9.3 kb (Table 1, P-T locality 9). Garnet growth here includes $S_2$ and $S_3$, but appears to predate in part $F_4$ deformation. The relations around the southern closure duplicate those seen at STOPS 8 and 9, namely that the

dome-stage folds fold the $F_3$ folds that have the asymmetric forms commonly ascribed to dome stage diapiric rise!. It is clear to us that the rotation axes of the $F_3$ folds are folded with the $F_3$ axial surfaces and these axes ($F_3$) are not related to dome-stage folding. Instead, dome stage folds are predominantly northeast trending to north trending later upright folds.

The small nonfoliated aplite dikes that cross cut $S_2$, $S_3$, and, $S_4$, are common in this area and are thought to be fine-grained equivalents of the Black Mountain Granite. Throughout the Athens dome, these dikes are nonfoliated and crosscut the $F_4$ folds. If the Black Mountain dikes of the Athens dome are coeval with the dated 373 Ma Black Mountain Granite (Aleinikoff, pers. comm.), deformation of the Athens dome predated 373 Ma. The Black Mountain Granite of the Guilford dome (Hepburn and others, 1984) however is strongly foliated and folded according to them. This suggests that the Guilford dome is younger than the Athens dome.

Dome forming events in the Athens dome ($F_4$ folds) ceased before dome stage events in the Guilford dome to the east. This interpretation fits the observations of Armstrong that areas immediately to the east of the Athens dome, east of the South Newfane thrust contain a new young and strong $F_5$ deformation, which is weakly expressed in ares to the west. The strong strain increase eastward at the South Newfane thrust, the absence of $F_2$, $F_3$, and $F_4$ folds events in the rocks east of the south Newfane thrust as well as the deformation of the Black Mountain Granite suggest eastward younging of the dome stage events in southern Vermont.

45.1 Continue south into Moretown Formation crops in brook.
47.4 Bridge over Baker Brook and road to right pull up on right side of road; walk 100 ft southeast to crops in Baker Brook.

**STOP 13. INTRUSIVE FELSIC GNEISS AND MAFIC XENOLITHS AND THE DEFORMATION PLAN OF THE SOUTH NEWFANE LITHOTECTONIC UNIT (BRATTLEBORO QUADRANGLE)** (15 MINUTES). This stream outcrop typifies the coarse-grained, equigranular quartz-plagioclase gneiss unit present within the structurally lowermost part of the South Newfane lithotectonic unit and belonging to the Barnard Volcanic Member of the Missisquoi Formation of Doll and others (1961). Notice the abundant mafic "pods," some of which contain plagioclase laths (phenocrysts) that commonly preserve concentric igneous zoning. The pods are interpreted as xenoliths within the felsic intrusive host. Compare this relationship to the reverse relations at the next stop (14). Notice that the dominant sole cleavage at this locality is undeformed, without any recognizable overprinting crenulation. Foliation surfaces tend to have a moderate to well-developed rodding or mineral-elongation lineation that plunges moderately N.45E. This lineation and related foliation, $S_5$, increases in intensity to the west up to the contact with the underlying Moretown Formation which is itnerpreted as an Acadian synmetamorphic fault zone. This fault, named the South Newfane thrust, is marked by a zone as much as 0.5 km wide of strong mylonitic foliation. Within this zone asymmetric feldspar augen and quartz grains have preferred grain-shape orientations obliquely plunging in the shear zone foliation. Fault-related meso- and micro-scale folds have axial planes oriented at an angle to the shear zone foliation. All of these kinematic indicators record east-over-west reverse (thrust) motion with a right lateral component of slip. Lithic truncations of both upper (South Newfane lithotectonic unit) and lower plate (Moretown) units support the fault interpretation. In addition, Taconian $S_2$ foliation and associated down dip (South 65 East trend) lineation, and Acadian $S_3$ and $S_4$ cleavages are present within the Moretown, but are transposed into $S_5$ as the fault zone is approached. Rocks of the South Newfane lithotectonic unit above this fault in areas removed from the fault have very low $S_5$ strain and preserve a weak compositional layering ($S_1$) apparently not related to folding. No Taconian $S_2$ nor Acadian $S_3/S_4$ have been recognized within these rocks. These observations, coupled with the abundance of mafic and felsic intrusive (hypabyssal) rocks in the Barnard, and their complete absence from the underlying Moretown have led us to believe that the STZ is a very significant Acadian tectonic boundary separating a unique Acadian deformational regime in the east from that to the west.

Continue southeast
47.6 Park by white house, behind which are crops.

**STOP 14. CROSS-CUTTING MAFIC DIKES WITHIN LAYERED FELSIC GNEISS OF THE SOUTH NEWFANE LITHOTECTONIC UNIT (BRATTLEBORO QUADRANGLE)** (15 MINUTES). BE CAREFUL! ROCKS ARE SLIPPERY. Brook exposure of well-layered felsic gneiss, possibly recrystallized volcanics (water-lain tuffs), with well-developed $S_1$ compositoinal layering (bedding?) that is truncated by a series of younger mafic intrusives. Although compositionally similar to the felsic gneiss at the previous stop (13), this gneiss can be mapped separately on the basis of its strong layering. Contacts with the massive felsic gneiss are typically sharp. The massive gneiss forms 1- to 5-m thick layers (possibly dikes) in the layered gneiss parallel to the $S_5$ foliation. The mafic rocks at this locality contain plagioclase laths (phenocrysts) which are commonly absent in "chilled" margins 10 to 30 cm from the contacts with the felsic gneiss. This relationship is common throughout this belt where mafic rocks judged to be intrusive dikes or sills exhibit well-developed chilled margins. Some of the mafic intrusives contain xenoliths of the layered felsic gneiss but do not contain xenoliths of the more massive felsic gneiss. Several relations need to be resolved: 1) what are the ages of the mafic and felsic rock-types, and can any of the presently available radiometric age data be used to constrain the ages of either the mafic intrusions or the deformations? 2) is the layered felsic gneiss at this stop the volcanic edifice of the massive felsic (intrusive) gneiss at stop 13? 3) are the mafic rocks that appear to intrude massive gneiss at stop 13 part of the same intrusive phase as mafic rocks at this stop?

## END OF TRIP

## INSTRUCTIONS TO ROUTE 30

Continue southeast from STOP 13; at intersection make left at Williamsville; at east side of Williamsville, turn right on bridge over Rock River; follow this road about 2 miles to intersection with Route 30; turn right, follow to Brattlesboro to Route 5; entrance to 91 south from Route 5 is about 2 miles south.

## REFERENCES CITED

Aleinikoff, J.N., and Karabinos, P., 1990, Zircon U-Pb data for the Moretown and Barnard Volcanic Members of the Missisquoi Formation and a dike cutting the Standing Pond Volcanics, southeastern Vermont: in Slack, J.F., ed., Summary Results of the Glens Falls CUSMAP Project, New York, Vermont and New Hampshire: U.S. Geological Surve Bulletin 1887, p. D1-D10.

Ando, C.J., Czuchra, B.L., Klemperer, S.L., Brown, L.D., Cheadle, M.J., Cook, F.A., Oliver, J.E., Kaufman, S., Walsh, T., Thompson, J.B., Jr., Lyons, J.B., and Rosenfeld, J.L., 1984, Crustal profile of mountain belts: COCORP deep seismic-reflection profiling in New England Appalachians and implications for architecture of convergent mountain chains: American Association of Petroleum Geologist Bulletin, v. 68, p. 819-837.

Armstrong, T.R., 1992, Progressive evolution of Acadian dynamothermal events in southern Vermont: Evidence for time transgressive dome development: Geological Society of America, Abstracts with Programs, v. 24, no. 3, p. 4.

Armstrong, T.R. and Tracy, R.J., 1991, Tectonometamorphic evolution of Acadian domes in southern Vermont: Geological Society of America, Abstracts with Programs, v. 23, no. 1, p. 3.

Armstrong, T.R., Tracy, R.J., and Hames, W.E., 1992, Contrasting styles of Taconian, Eastern Acadian and Western Acadian metamorphism, central and western New England: Journal of Metamorphic Geology, 10, p. 415-426.

Bell, T.H., Johnson, S.E., Davis, B., Forde, A., Hayward, N., and Wilkins, C., 1992, Porphyroblast inclusion-trail orientation data: eppure non son girate*!: Journal of Metamorphic Geology, no. 10, p. 295-307.

Bell, T.H., and Johnson, S.E., 1989, Porphyroblast inclusion trails: the key to orogenesis: Journal of Metamorphic Geology, no. 7, p. 279-310.

Berman, R.G., 1988, Internally-consistent thermodynamic data for minerals in the system $K_2O$-$Na_2O$-CaO-MgO-FeO-$Fe_2O_3$-$Al_2O_3$-$SiO_2$-$TiO_2$-$H_2O$-$CO_2$: Journal of Petrology, 29, p. 445-522.

____1990, Mixing properties of Ca-Mg-Fe-Mn garnets: American Mineralogist, v. 75, p. 328-344.

Bohlen, S.R., and Liotta, J.J., 1986, A barometer for garnet amphibolites and garnet granulites: Journal of Petrology, v. 27, p. 1025-1034.

Burton, W.C., 1991, Bedrock geologic map of the Woodford quadrangle: U.S. Geological Survey Geologic Quadrangle Map, GQ-1687, scale 1:24,000.

Burton, W.C., in press, Preliminary bedrock geologic map of the Sunderland quadrangle, Vt: U.S. Geological Survey Miscellaneous Field Investigations Map MF_____, scale 1:24,000.

Burton, W.C., Kunk, M.J., and Ratcliffe, N.M., 1990, Muscovite and microcline $^{40}Ar/^{39}Ar$ ages from the Taconian- and Acadian-overprinted central Green Mountain massif, Vermont: Geological Society of America Abstracts with Programs, v. 22, no. 2, p. 7.

Burton, W.C., Kunk, M.J., and Ratcliffe, N.M., 1991, Microline and biotite $^{40}Ar/^{39}Ar$ ages and their bearing on the Taconian and Acadian thermal history of the central Green Mountain massif, Vermont: Geological Society of America Abstracts with Programs, v. 23, no. 1, p. 13.

Chidester, A.H., Hatch, N.L., Jr., Osberg, P.H., Norton, S.A. and Hartshorn, J.H., 1967, Geologic map of the Rowe quadrangle, Massachusetts and Vermont: U.S. Geological Survey Geologic Quadrangle Map, GQ-642, Scale 1:24,000.

Dickenson, M.P., and Hewitt, D.A., 1986, A garnet-chlorite geothermometer: Geological Society of America, Abstracts with Programs, v. 18, p. 584.

Doll, C.G., Cady, W.M., Thompson, J.B., Jr., and Billings, M.P., 1961, Centennial Geologic Map of Vermont: Montpelier, Vermont Geological Survey, scale 1:250,000.

Eskola, P., 1949, The problem of mantled gneiss domes: Quarterly Journal of the Geological Society of London, v. 104, p. 461-476.

Ferry, J.M., and Spear, F.S., 1978, Experimental calibration of partitioning of Fe and Mg between biotite and garnet: Contributions to Mineralogy and Petrology, v. 66, p. 113-117.

Fisher, G.W., 1989, Matrix analysis of metamorphic mineral assemblages and reactions: Contributions to Mineralogy and Petrology, 102, p. 69-77.

Fletcher, C.J.N., and Greenwood, H.J., 1979, Metamorphism and structure of the Penfold Creek area, near Quesnel Lake, British Columbia: Journal of Petrology, 20, p. 743-794.

Ghent, E.D., and Stout, M.Z., 1981, Geobarometry and geothermometry of plagioclase-biotite-garnet-muscovite assemblages: Contributions to Mineralogy and Petrology, v. 76, p. 92-97.

Hatch, N.L., Jr., Chidester, A.H., Osberg, P.H., and Norton, S.A., 1966, Redefinition of the Rowe Schist in northwestern Massachusetts: in changes in stratigraphic nomenclature by the U.S. Geological Survey 1965, Cohee, G.V. and West, W.S. eds., U.S. Geological Survey Bulletin 1244-A, p. A33-A35.

Hayward, J.A., Gaudette, H.E., and Olszewski, W.J., 1988, Isotopic characteristics of and constraints on parental sources of two-mica granites in northern New England: Geological Society of America Abstracts with Programs, v. 20, no. 1, p. 26.

Hayward, N., 1992, Microstructural analysis of the classical snowball garnets of southeast Vermont: evidence for nonrotation: Journal of Metamorphic Geology, no. 10, p. 567-587.

Hepburn, J.C., Trask, N.J., Rosenfeld, J.L., and Thompson, J.B., Jr., 1984, Bedrock Geology of the Brattleboro quadrangle, Vermont-New Hampshire: Vermont Geological Survey Bulletin, 32.

Karabinos, Paul, 1984a, Deformation and metamorphism on the eastside of the Green Mountain massif in southern Vermont: Geological Society of America Bulletin, v. 95, no. 5, p.584-593.

____1984b, Polymetamorphic garnet zoning from southeastern Vermont: American Journal of Science, v. 284, no. 9, p. 1008-1025.

Karabinos, Paul, and Aleinikoff, J.N., 1990, Evidence for a major Middle Proterozoic, post-Grenvillian igneous event in western New England: American Journal of Science, v. 290, no. 8, p. 959-974.

Karabinos, Paul, and Laird, Jo, 1988, Structure and metamorphism from Jamaica to the Athens dome, Vermont, Trip C-3, in Bothner, W.A., ed., Guidebook for field trips in southwestern New Hampshire, southeastern Vermont and north-central Massachusetts: New England Intercollegiate Geological Conference, 80th Annual Meeting, Keene, New Hampshire, p. 281-292.

Kohn, M.J., and Spear, F.S., 1990, Two new geobarometers for garnet amphibolites with applications to southeastern Vermont: American Mineralogist, v. 75, p. 89-96.

Koziol, A.M., and Newton, R.C., 1989, Redetermination of the anorthite breakdown reaction and improvement of the plagioclase-garnet-$Al_2SiO_5$-quartz barometer: American Mineralogist, v. 73, p. 216-223.

Laird, Jo, 1991, Acadian metamorphism about the Taconian line, southeastern Vermont: Geological Society of America Abstracts with Programs, v. 23, no. 1, p. 55.

Laird, Jo, and Albee, A. L., 1981, Pressure, temperature, and time indicators in mafic schist: their application to reconstructing the polymetamorphic history of Vermont: American Journal of Science, v. 281, no. 2, p. 127-175.

Laird, Jo, Lanphere, M.A., and Albee, A.C., 1984, Distribution of Ordovician and Devonian metamorphism in mafic and pelitic schists from northern Vermont: American Journal of Science, v. 284, no. 4-5, p. 376-413.

Naylor, R.S., 1971, Acadian Orogeny: an abrupt and brief event: Science, v. 172, p. 558-560.

Nisbet, B.W., 1976, Structural studies on the northern Chester dome of east-central Vermont: Unpublished Ph.D. thesis, State University of New York at Albany, p. 265.

Passchier, C.W., Trouw, R.A.J., Zwart, H.J., and Vissers, R.L.M., 1992, Porphyroblast rotation: eppur si muove*!: Journal of Metamorphic Geology, no. 10, p. 283-294.

Ratcliffe, N.M., 1990, Basement cover relationships, Paleozoic structural chronology and tectonic history of the Green Mountain massif and eastern domes of southern Vermont: Comparison with other massifs in the orogen: Geological Society of America Abstracts with Programs, v.22, no. 2, p. 63.

____1991a, Revisions to nomenclature of some Middle Proterozoic granitic rocks in the northern Berkshire massif, Massachusetts, and in the southern Green Mountains, Vermont and Massachusetts, in Stratigraphic Notes, 1989-90: U.S. Geological Survey Bulletin 1935, p. 9-26.

____1991b, Chemistry and tectonic setting of metabasalts of the Hoosac Formation, southern Vermont: Geological Society of America Abstracts with Programs, v. 23, no. 1, p. 122.

____1992, Bedrock geologic maps of the Mount Snow and Readsboro quadrangle, Bennington and Windham Counties, Vermont: U.S. Geological Survey Geologic Miscellaneous Investigations Series, Map I 2307, scale 1:24,000.

Ratcliffe, N.M., in press, Bedrock geologic map of the Jamaica Quadrangle and portions of the adjacent Townshend Quadrangle, Windham County, Vermont: U.S. Geological Survey Miscellaneous Investigations Map, I-_____, scale 1:24,000.

Ratcliffe, N.M., Aleinikoff, J.N., Burton, W.C. and Paul Karabinos, 1991, Trondhjemitic 1.35 - 1.31 Ga gneisses of the Mount Holly Complex of Vermont: evidence for an Elzevirian event in the Grenville basement of U.S. Appalachians: Canadian Journal of Earth Science, v. 28, p. 77-93.

Ratcliffe, N.M. and Burton, W.C., 1989, Preliminary bedrock geologic map of the Stratton Mountain quadrangle: U.S. Geological Survey Open-File Report 89-200, scale 1:24,000.

Ratcliffe, N.M., Burton, W.C., Sutter, J.F., and Mukasa, S.A., 1988, Stratigraphy, structural geology and thermochronology of the northern Berkshire massif and southern Green Mountains, Trips A-1 and B-1, in Bothner, W.A., ed.: Guidebook for field trips in southwestern New Hampshire, southeastern Vermont, and northcentral Massachusetts, New England Intercollegiate Geological Conference, 80th Annual Meeting, Keene, N.H., p. 1-32 and 126-135.

Rosenfeld, J.R., 1954, Geology of the southern part of the Chester dome, Vt: Ph.D thesis, Harvard University, Cambridge, MA, 303 p.

Rosenfeld, J.R., 1968, Garnet rotation due to major Paleozoic deformations in southeastern Vermont: in Studies of Appalachian Geology, Northern and Maritime, E-an Zen, W.S. White, J.B. Hodby, and J.B. Thompson, Jr. (eds.), Interscience, New York, p. 185-202.

Rosenfeld, J.R., Christenson, J.N., and DePaolo, D.J., 1988, Snowball garnets revisited, southeast Vermont, in Guidebook for field trips in southwestern New Hampshire, southeastern Vermont and north central Massachusetts, W.A. Bothner, ed., New England Intercollegiate Geological Conference 80th Annual Meeting, Keene, New Hampshire, p. 223-240.

Skehan, J.W., S.J., 1961, The Green Mountain anticlinorium in the vicinity of Wilmington and Woodford, Vermont: Vermont Geological Survey Bulletin No. 17, 159 p., scale 1:62,500.

Stanley, R.S., and Hatch, N.L., Jr., 1988, The Pre-Silurian geology of the Rowe-Hawley zone, Chapter A in Hatch, N.L., Jr., ed., The bedrock geology of Massachusetts: U.S. Geological Survey Professional Paper 1366, p. A1-A39.

Sutter, J.F., and Hatch, N.L., 1985, Timing of metamorphism in the Rowe-Hawley Zone, western Massachusetts: Geological Society of America, Abstracts with Programs, 17, p. 65.

Sutter, J.F., Ratcliffe, N.M., and Mukasa, S.B., 1985, $^{40}Ar/^{39}Ar$ and K-Ar data bearing on the metamorphic and tectonic history of western New England: Geological Society of America Bulletin, 96, p. 123-136.

Thompson, A.B., Tracy, R.J., Lyttle, P.T., and Thompson, J.B., Jr., 1977, Prograde reaction histories deduced from compositional zonation and mineral inclusions in garnet from the Gassetts Schist, Vermont: American Journal of Science, 277, p. 1152-1167.

Thompson, J.B., Jr., 1950, A gneiss dome in southeastern Vermont: Unpublished Ph.D. Thesis, Harvard University.

Thompson, J.B. Jr., McLelland, J.M., and Rankin, D.W., 1990, Simplified geologic map of the Glens Falls 1°x2° quadrangle, New York, Vermont, and New Hampshire: U.S. Geological Survey Miscellaneous Field Studies Map, MF-2073, scale 1:250,000.

# THE CHICKEN YARD LINE / WHATELY FAULT DEBATE: FROM SPRINGFIELD,VERMONT TO WHATELY, MASSACHUSETTS

by
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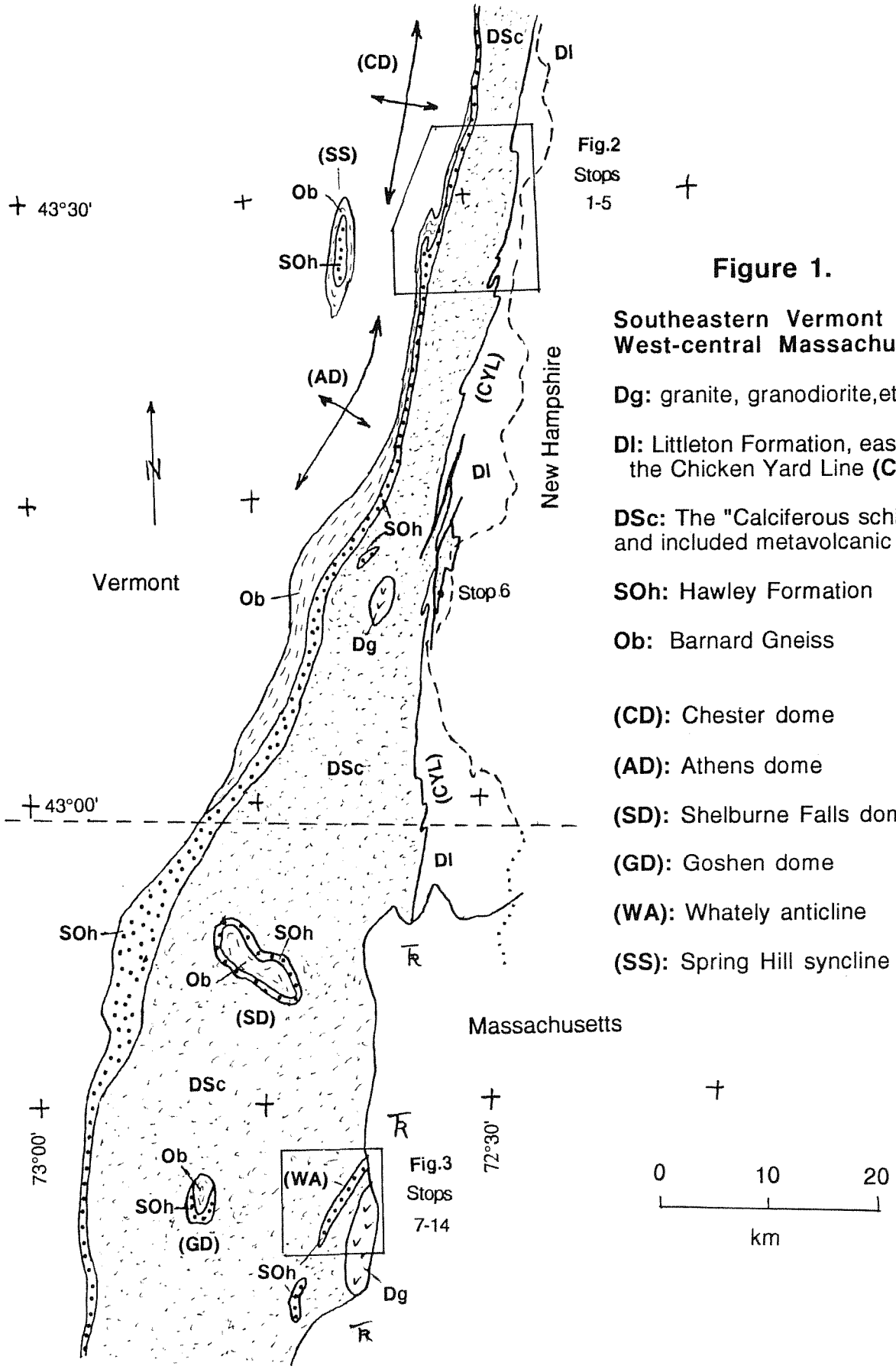
## INTRODUCTION

Although there exist certain similarities in rock types of comparable ages, the "Vermont sequence" has always been a separate entity from the "New Hampshire sequence" (Fig. 1) as coined by Billings (1956). Even before, and then with the coming of plate tectonics the meld of these two sequences along the Connecticut River has received less than unanimous consensus. In comparison to the Cambro-Ordovician rocks from the two sequences, which are far enough separated, the coming together of the Lower Devonian strata from Vermont with those from New Hampshire has seen considerable spilled ink. From central Vermont down through central Massachusetts along the Connecticut River, two interpretations have emerged as to how the "Vermont Sequence" meets the "New Hampshire Sequence". In one scenario the transition from the Gile Mountain Formation of Vermont to the Littleton Formation of New Hampshire is an erosional unconformity marked by a quartz conglomerate in the Littleton Formation as originally defined in a chicken yard 0.5 miles south of Dutton Pines State Forest Park northeast of Brattleboro, Vermont (Hepburn *et al.*, 1984). From this locale the passage from the Vermont to the New Hampshire sequence is taken to occur across the "chicken yard line", an erosional unconformity. In compiling the state geologic map of Massachusetts, Robinson et al. (1988) invoked a fault between rocks assigned by them to the Gile Mountain and Littleton Formations in the Whately area, Massachusetts. This is because they correlated the mafic volcanics and associated rocks found in a small anticline at Whately (Trzcienski, 1966, 1967) to the Erving Formation of Robinson (1963), that in areas to the east overlies rocks, in Robinson's view, the Lower Devonian Littleton Formation. In the Whately area (Fig. 3) these mafic volcanics underlie rocks assigned herein to the Waits River and Gile Mountain Formations, and are believed to be correlative with the rocks of the Hawley Formation further to the west. The stratigraphic interpretation proposed by Robinson *et al.* (1988), in contrast, requires a westward directed thrust in order that the Littleton Formation can overlie presumably younger rocks to the west. If the Erving farther east as defined by Robinson (1963) is indeed the same as our Hawley, then its relationship to the nearby Littleton, or supposed Littleton should perhaps be re-examined.

The purpose of this field trip is to look at the various rock lithologies and associated structures and metamorphism along the Connecticut River Valley from Springfield, Vermont southward to Whately, Massachusetts in order to debate and/or discuss the arguments and evidence for the two scenarios outlined above. A *caveat* to fieldtrippers; the four co-leaders sit, more or less, on the same side of the fence when it comes to interpreting the nature of the Vermont to New Hampshire sequence transition.

## RIGHT-OF-ACCESS

We are embarking, much of the time, onto privately owned land. Furthermore, most of the properties that we will be visiting are posted - **NO TRESPASSING**. The trip leaders have, for this day only, obtained permission to visit outcrops on these private lands. This does not give anyone the right in the future to pass onto these properties without first obtaining permission to do so. Therefore, if you wish to revisit the localities described in this field trip guide at some future date, plan ahead and obtain the necessary permission before you trespass. This is serious business, for one landowner encountered by the first author stated flatly that since he was removed from normal police patrols he had, first his own means to protect his property, and only second would he ask questions.




## Figure 1.

**Southeastern Vermont and West-central Massachusetts**

**Dg:** granite, granodiorite,etc.

**Dl:** Littleton Formation, east of the Chicken Yard Line **(CYL)**

**DSc:** The "Calciferous schists" and included metavolcanic rocks

**SOh:** Hawley Formation

**Ob:** Barnard Gneiss

**(CD):** Chester dome

**(AD):** Athens dome

**(SD):** Shelburne Falls dome

**(GD):** Goshen dome

**(WA):** Whately anticline

**(SS):** Spring Hill syncline

## STRATIGRAPHY

The central area of Fig. 1 is underlain largely by rocks corresponding in large part to the "calciferous schists" of Hitchcock et al. (p. 476, 1861). These include the Northfield of southern Vermont, the Waits River, Gile Mountain Formations, Meetinghouse Slate and associated Standing Pond Volcanics of Vermont, and the Goshen, Conway, and Leyden (at least in part) Formations of Massachusetts. To the west and in anticlines farther to the east these rocks are underlain by mafic volcanics and sulfidic schists, here assigned, to the Hawley Formation. These in turn are underlain by felsic gneisses and amphibolites of the Barnard Gneiss of Richardson (1927; Barnard Volcanic member of the Missisquoi Formation, Doll *et al.*, 1961). To the east the calciferous schists lie in contact with the Littleton Formation. The contact, known informally as the "chicken yard line", is interpreted by us as an eastward facing unconformity, though locally offset by faulting.

### Barnard Gneiss

The Barnard Gneiss will be seen only at two localities. The dominant gneisses are quartz, plagioclase, biotite gneisses with minor muscovite and potassium feldspar. These are interlayered with hornblende gneiss and amphibolites. A volcanic origin is believed to have been the precursor for these rocks, but some intrusive rocks may also be present.

### Hawley Formation

The Hawley Formation was originally named by Emerson (1898). Its extension into Vermont was assigned by Doll *et al.* (1961) to the Cram Hill member of the Missisquoi Formation and in part to the Shaw Mountain Formation. It also includes the Unnamed Unit and Russell Mountain Formation of Hepburn *et al.* (1984). As used here the Hawley includes amphibolites, sulfidic rusty schists, abundant coticules, silvery schists, quartzites and quartz conglomerates, and quartz, feldspar, biotite granulites. The quartzites and quartz conglomerates occur at two positions in rocks here assigned to the Hawley. The outcrops to be seen are in the lower part of the formation beneath a thick amphibolite. Quartzite and quartz conglomerate, however, also occur locally near the top of our Hawley Formation and have been mapped either as Russell Mountain Formation (Hatch *et al.*, 1970 , Hepburn *et al.*, 1984) or as Shaw Mountain Formation (Doll *et al.*, 1961) in the area of Fig. 1.

### Calciferous Schists

In Fig. 2 and 3 the calciferous schists are divided into three major metasedimentary units, but all three contain essentially the same lithologic types but in different proportions. The westernmost metasedimentary unit is a garnetiferous, gray, mica schist with impure micaceous quartzite and thin punky weathering marbles. These constitute the Goshen Formation of Massachusetts and the Northfield in southern Vermont. This unit thins northward so that the Waits River Formation, the central unit, rest directly on the Hawley on the east flank of the Chester Dome northwest of Springfield, Vermont. The Goshen Formation is also missing on both flanks of the Whately anticline where the Waits River lies directly upon the Hawley Formation. We believe that the Waits River Formation and the more westerly Goshen and Northfield are facies equivalents of one another. The Waits River Formation differs from the Goshen and Northfield by having a much higher percentage of punky weathering marble and less aluminous schist.

The eastern unit of the calciferous schists is the Gile Mountain Formation (Doll, 1944) on recent state maps of Vermont and Massachusetts. It also includes the western portion of Emerson's (1917) Leyden Formation. The boundary between the Waits River and Gile Mountain Formations is clearly not a time line. In places the mafic Standing Pond Volcanics have been mapped as within the Waits River (Central Vermont), elsewhere they are the boundary between the Waits River and Gile Mountain Formations (southern Vermont) or

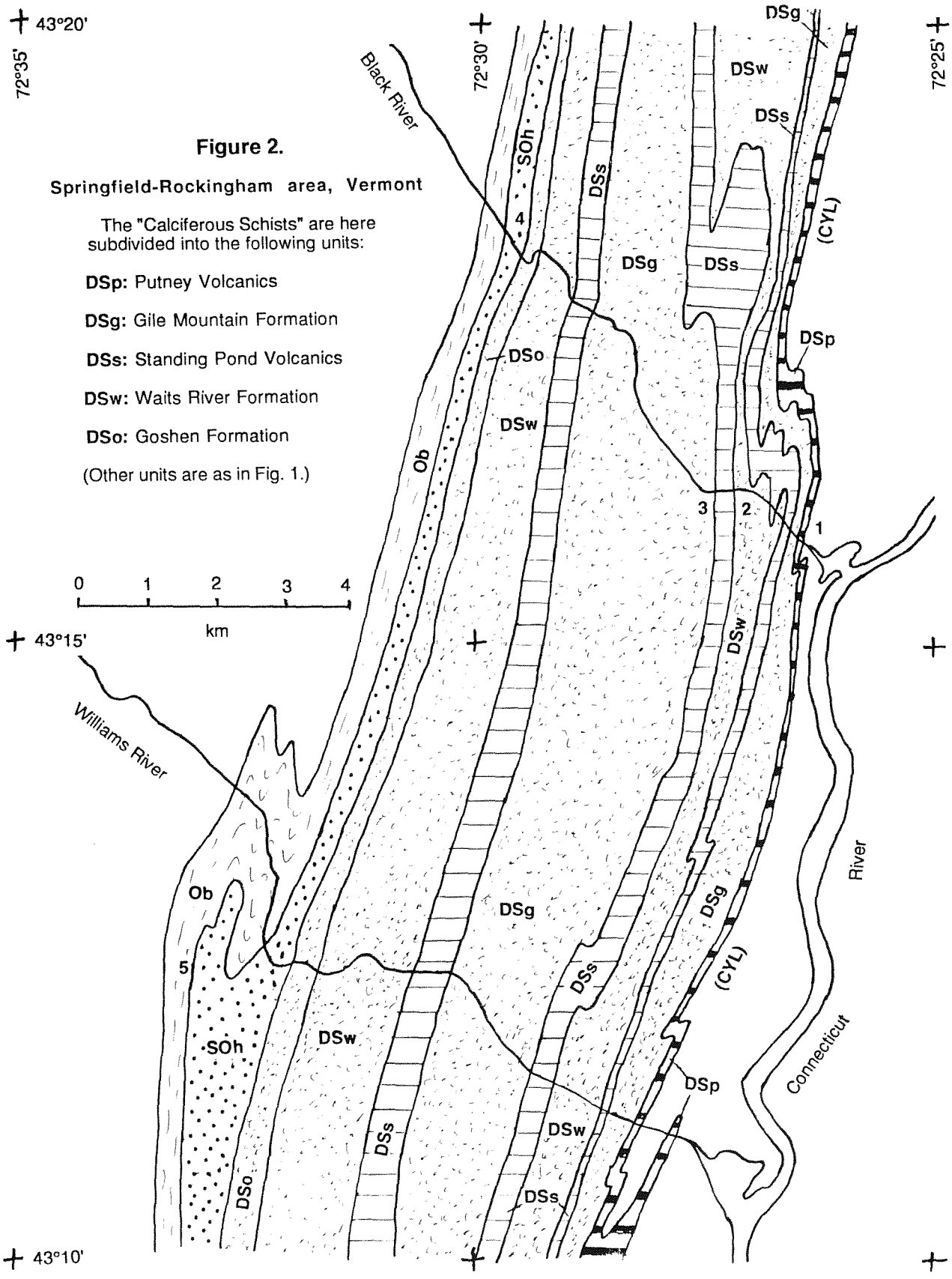



**Figure 2.**

**Springfield-Rockingham area, Vermont**

The "Calciferous Schists" are here subdivided into the following units:

**DSp:** Putney Volcanics

**DSg:** Gile Mountain Formation

**DSs:** Standing Pond Volcanics

**DSw:** Waits River Formation

**DSo:** Goshen Formation

(Other units are as in Fig. 1.)


Figure 3. Geological sketch map of the Whately area.

within the Gile Mountain (northern Vermont). Unfortunately topping evidence is ambiguous in the vicinity of the Standing Pond, although we interpret the Gile Mountain as being generally younger. In Whately, however, the Whately Pound channel deposit clearly indicates tops to the east from Waits River into Gile Mountain types. This channel deposit has been correlated with the "chicken yard line", but we believe that this correlation is invalid and therefore, misleading. Furthermore, recent fossil evidence (Hueber *et al.*, 1990) indicates that the Gile Mountain formation is Emsian in age, and therefore not necessarily much younger than the Littleton Formation as a "Whately thrust" requires.

**Putney Volcanics**

The Putney Volcanics are variable in composition and vary from feldspar-rich felsic rocks to more mafic chlorite-rich rocks. Just west of the "chicken yard line", these volcanics contain thin bedded feldspathic granulites (Hepburn *et al.*, 1984), some of which show cross bedding and possible load casts indicating tops to the east toward the "chicken yard line".

**Littleton Formation**

The Littleton Formation east of the "chicken yard line" is largely a gray slate (locally quarried as a roofing slate) or phyllite **devoid** of punky weathering, impure marbles. Against the "chicken yard line", within the Littleton Formation there are quartzites, conglomerates, and slate matrix conglomerates. Apparent channeling within these conglomerates indicates tops east.

**ACKNOWLEDGEMENTS**

Our sincere thanks to the landowners of Vermont and Massachusetts who have enthusiastically permitted our entering upon their properties to look at their stones, ledges, and rocks.

> "The Goshen and Conway schists and the Leyden argillite form a continuous series, separated from the formations below and above by considerable unconformities. The deposits doubtless were laid down in a sea that was shallow but continuous, as is indicated by the abundant carbon, and whose boundaries and history were different from those of the seas which preceded and followed it. On the south and west the formations were successively narrower, the Leyden, the latest of the three, lying within the Conway, and the Conway within the Goshen - the oldest. Hence it may be inferred that the sea had shrunk northeastward, or the land had expanded in that direction, during the corresponding epochs."
>
> B.K. Emerson (p. 46, 1917)

## ITINERARY

Assemble at Howard Johnson's parking lot off Exit 7 of I91 (west side), Springfield, Vermont.

Mileage

0.0 **STOP 1. HOWARD JOHNSON'S AND CHICKEN YARD LINE**

A sheared version of the "chicken yard line" is exposed on the north side of Rte 11, just west of the Howard Johnson's parking lot. On the west side of the nearly vertical contact lie greenstones and associated rocks of the Putney Volcanics, and on the east side are grey, slaty phyllites of the Littleton Formation. Several boudins of sheared, pebbly quartzite lie along the contact. After inspection of these features walk south across the bridge on Rte 5, about 0.3 mi. to a side road leading to a YMCA camp. Cross the side road to a clump of trees containing an outcrop of pebbly quartz conglomerate forming a steeply-plunging, northward-opening syncline containing grey phyllite of the Littleton Formation. Return to Rte 5 and walk south a few meters to an outcrop on the right containing greenstone of the Putney Volcanics. This greenstone is in the core of a southward-opening anticline that, with the syncline just seen, defines one of the many, steeply plunging, east-side-north folds that offset the "chicken yard line" from here south into Massachusetts. Accompanying these folds in many places, not clearly visible here, is a pervasive northeast-oriented, near-vertical, east-side north crenulation cleavage. The outcrops between here and the I-91 overpass are grey phyllites and phyllite-matrix conglomerates of the Littleton Formation. The "chicken yard line" is exposed again in cuts on the east side of I-91, a short distance south, as we shall see later this morning as we drive south on the highway after **STOP 4**.

From here return to Howard Johnson's and walk west along the north side of Rte 11 to see more of the Putney Volcanics, part of the Gile Mountain Formation, and finally, just east of a side road heading north, an exposure of the contact between the Gile Mountain Formation (to the east) and the Standing Pond Volcanics (to the west). **RETURN** to vehicles.

0.9 **STOP 2. WAITS RIVER FORMATION**

This is a good example of the Waits River Formation and the punky weathering, impure marble that is typical of this formation. On the east side of the outcrop the Waits River Formation is in contact with the Standing Pond Volcanics.

1.0 **STOP 3. STANDING POND VOLCANICS**

This outcrop is in the Standing Pond Volcanics of the eastern gunwale of the canoe-shaped pattern of the Standing Pond Volcanics as shown on the Vermont State map (Doll *et al.*, 1961). Near to the contact with the Gile Mountain Formation on the west side, a dike cutting the volcanics has yielded a zircon date of 423±2 Ma (Aleinikoff and Karabinos, 1990). **RETURN** to vehicles.

3.9 Turn right off of Rte 11 onto Rte 106

4.2 This is a small outcrop of the Barnard Gneiss on the northeast side of Rte 106 that occurs to the west of the Hawley Formation. We will see the Barnard Gneiss at **STOP 5**. From here turn around in the parking area of Sherwin Williams Paints and return eastward on Rte 106.

4.4 Park on the southwest side of Rte 106 in McDonald's parking area (**BE QUICK OR BE LEFT BEHIND**).

**STOP 4. HAWLEY FORMATION**

On the northeast side of the highway, displayed along the sidewalk is part of the Hawley Formation. The outcrop is composed of amphibolites, some of which may be fragmental. Recently Spear and Harrison (1989) have obtained an $^{40}Ar/^{39}Ar$ hornblende release spectrum date of 433±3 Ma from this outcrop. From here walk eastward to the Mobil station and turn left into the parking lot. We

have come across the contact of the Hawley Formation with the calciferous schists of the Northfield Formation. Note the punky marble in the small outcrop of the Northfield exposed here. Across the road on the south side of the river below the dam are thick punky weathering marbles of the Waits River Formation. **RETURN** to the vehicles.

Head eastward on Rte 11, back to Howard Johnson's where the remaining vehicles should be occupied. From Howard Johnson's we will travel, in caravan, and we will not stop again until getting off I-91 at EXIT 6 southbound to go west on Rte 103. As we head south on I-91, however, the trip leaders will slow down at the observed mileages (±) so fieldtrippers can observe at ≈40 MPH some of the important geologic features along the interstate.

9.6 Enter I-91 from Rte 11.
10.45 Littleton Formation in median strip close to the "chicken yard line".
10.55 The contact between the Putney Volcanics and the Littleton Formation is on the right. By close inspection one can again see the small, stretched boudins of quartzites that were seen in the outcrop just west of Howard Johnson's. The fieldtrip will be crossing several times, back and forth, the "chicken yard line" as we head south on I-91.
10.65 Low in the ditch-line is a good punky weathering marble of the Gile Mountain Formation.
11.3 Standing Pond Volcanics on right.
12.0 Gile Mountain Formation.
12.9 Gile Mountain Formation with a large limy zone.
13.7 Gile Mountain Formation.
13.9 Granites that intrude the Littleton Formation.
14.6 From here to EXIT 6 we will be in the Littleton Formation.
15.8 Cross over the Williams River.
16.1 **EXIT 6** Leave I-91 onto westbound Rte 103.
16.45 Sunoco Station on right side.
16.55 We will assemble in the yard of the inactive gas station here and maximize occupiable vehicle space - take as few vehicles as necessary from here to **STOP 5**. Continue westward on Rte 103. We are presently east of the "chicken yard line" in the Littleton Formation, and we will traverse a section, but with little road exposure, that is similar to that seen along Rte 11 between **STOPS 1** and **3**.
22.0 Turn left onto road opposite Vermont State Police Barracks.
22.1 Turn left.
22.15 Turn right
23.15 Barnard Gneiss.

23.5 **STOP 5. BARNARD GNEISS AND HAWLEY FORMATION.**

The outcrops here are essentially on strike with those seen at **STOP 3** and the first part of **STOP 4**. We will walk east from the road over outcrops of Barnard Gneiss into rusty-weathering schists containing coticules and amphibolites of the Hawley Formation. Zircons separated from the Barnard gneiss at this locality have given an approximate age of 524±4 Ma (Aleinikoff and Karabinos, 1990). More recent work by Karabinos (personal communication, August, 1992) has refined this date; now giving 484±2 Ma. These are succeeded eastward by quartz conglomerate and quartzite that are overlain in turn by a thick series of amphibolites also of the Hawley Formation. We shall walk only to the crest of the first ridge then return to the road by a more northerly route. Between the last amphibolites seen and the calciferous schists to the east there are coticules, iron formations, and more amphibolites interbedded with rusty-weathering schist and granofels. These can be seen in good outcrop on the hills just north of the Williams River, but we do not have enough time to visit there. A well-exposed and strikingly similar sequence can also be seen approximately six to seven kilometers farther west (and west of the axis of the Chester and Athens domes) in the Spring Hill syncline (Fig. 1). The rocks in the core of the Spring Hill syncline are largely amphibolites in the upper part of the Hawley Formation, and *not* a part of the calciferous schists as erroneously shown by Doll *et al.*, 1961,

the result of a communications snafu between two of your trip leaders! **RETURN** to vehicles and Rte 103.

25.0 Turn right onto Rte 103.
30.5 Reassemble caravan for trip south.
30.6 Entrance onto I-91 south. The large road cuts here are in the Littleton Formation.
31.9 Littleton Formation
32.1 Littleton Formation
32.3 We have again crossed the "chicken yard line".
32.5 Putney Volcanics in both the median strip and the right-hand side of the highway.
33.0 From here to EXIT 4 we are again in the Littleton Formation.
48.1 EXIT 4 Leave I-91 to the right.
48.2 Turn left onto southbound Rte 5. We will be passing by outcrops of the Gile Mountain Formation.
51.0 Putney Volcanics along road.
51.5 Dutton Park State Forest.
52.0 Turn into driveway on right and park. **BE CAREFUL**, for we are on a blind corner to traffic from both the north and the south.

**STOP 6. THE CHICKEN YARD**

This is the famous "chicken yard line" locality which we will see by walking several tens of meters to the north into the woods in an old, abandoned chicken yard (Watch your step!). Note the exposed conglomerate. At one time channeling was visible indicating tops to the east. We may or may not have cleaned the outcrop sufficiently to make these features visible. (See Hepburn *et al.*, 1984, Fig. 2-7, for topping evidence in slate and granofels of the Putney Volcanics.) After visiting this outcrop we will walk south along Rte 5. **TRAFFIC!** Phyllite, quartzite, and phyllite matrix conglomerate are exposed along both sides of the highway. A small roofing slate quarry can be seen east of the highway just beyond the road cuts in the Littleton Formation. **RETURN** to vehicles and continue south on Rte 5.

52.8 Putney Volcanics are exposed all along on our left.
54.5 On the hillside to the left, the quartzite of the "chicken yard line" is again exposed. To the south is the Littleton Formation, and to the north are the Putney Volcanics. The contact strikes southeasterly to easterly and is truncated by a fault along or just west of the highway.
53.7 Cross over I-91.
54.3 Turn right at the lights and continue west to enter back onto I-91 south.
57.2 Standing Pond Volcanics are on the left where the northbound entrance ramp of interchange #2 merges with I-91. Beyond the volcanics to the east is the marble member of the Gile Mountain Formation (Hepburn *et al.*, 1984).
58.7 Interchange #1 is within the Gile Mountain Formation. Several punky marbles can be seen in exposures on the west side of I-91.
65.8 In the northbound lane to the east is a granite dike cutting the Gile Mountain Formation, where again there are several beds of punky marble.
77.9 The red rocks on the left are the Triassic Sugarloaf Formation.
78.4 Interchange with Rte 2.

85.3 **EXIT 25 Deerfield-Conway. Leave I-91 by turning right. On the exit ramp turn right again to head west on Rte 116. Almost immediately on entering Rte 116 turn left onto Whately Road. (Mileage in parentheses is from 0.0 at this point.)**

85.4 (0.0) Continue south on Whately Road.
85.9 (0.5) Turn right and continue west on Whately Road.
86.7 (1.3) Cross Bloody Brook, continue westward. Historical note: Name of this brook is derived from the time of the French and Indian Wars when an attack on the settlers of Deerfield supposedly turned this stream blood red.
86.9 (1.5) Intersection with Town Farm Road - continue toward the left on Whately Road.
87.1 (1.7) Whately-Deerfield town line.
87.5 (2.1) For the next 1.5 miles we will be driving along the interface between the Sugarloaf and Waits River Formations.
88.0 (2.6) Cross Roaring Brook.

**To those for whom poison ivy or oak is a potential danger look for it at every outcrop.**

88.65 (3.25) Park along roadway.

**STOP 7. MASSIVE VOLCANICS OF THE HAWLEY FORMATION**

Please pay attention to and DO NOT DISPLACE the fence we will be crossing in order to look at the outcrop in the pasture. This is the northeasternmost exposure of the massive amphibolites of the Hawley Formation. The rock is a good epidote amphibolite that contains a number of small folds, the majority of which plunge to the northeast. Several of these folds can be seen right under the electrified fence. **RETURN** to vehicles and continue south.

89.3 (3.9) QuonQuont Farm sign on right, continue straight into Whately Center.
89.55 (4.15) Whately Inn (Good food and drink but not a stop).
Turn right here onto Haydenville Road and continue westward.
89.8 (4.4) Whately Pound on right. Historical note (last one): This stone enclosure was used back in the "good old days" to keep stray cattle, which were common then, until the owner found it missing and came looking for it here at this central gathering place. The land owner across the street is the current "Pound Keeper".
89.9 (4.5) Park in pullout on the south side of the road. Pay attention to traffic whipping around the corner from the west.

**STOP 8. GILE MOUNTAIN FORMATION**

**NO HAMMERS ON CHANNEL DEPOSIT**. This is the Whately Pound channel deposit in the Gile Mountain Formation. Walk into the woods along the culvert to the outcrop just south of the ditch. This is a rather nice outcrop of a stream channel deposit composed primarily of quartz. It is graded and indicates that tops are to the east. Note especially the two cycles of graded bedding and the load-casts in the middle of the channel. We will traverse a short distance toward the east to look at the typical, alternating sequence of phyllite and sandy beds of the Gile Mountain Formation. The contrast between the phyllitic beds and the quartzose ones accentuates in many places the style of deformation found in this unit. In addition several stops will be made to look at thin impure marble horizons that define this unit as the Gile Mountain Formation rather than the Littleton Formation. Across the road to the north other impure marbles can be found, and northwestward the rocks begin the transition into the Waits River Formation. **RETURN** to vehicles.

89.975 (4.575) Turn right onto Dickinson Hill Road and climb hill.

90.6 (5.2) Enter onto dirt road.
90.65 (5.25) Turn into driveway on the right and park in either the hayfield or along the road.

**STOP 9. HAWLEY FORMATION**

The first small outcrops encountered on the knoll in the hayfield are of the massive amphibolite. There is then a white sugary rock in a small outcrop that is the westernmost outcropping of the quartzitic horizon. We will then continue into the woods to look at the black rusty schists, massive quartzite, a garnet-rich bed, and minor folds. Continuing to the east flank of the Whately anticline one finds the best exposures of the massive quartzite. These outcrops of massive, vitreous quartzite weather a light brown, and they contain a significant percentage of muscovite. There is also exposed here a nicely formed, northeasterly plunging anticline within the quartzite. Following this unit northeastward one does see some small pebbles in places and blue quartz grains. If one were to continue on beyond the stone fence one would encounter massive amphibolite, and a walk up onto Bull Hill would cross rusty schist within the massive amphibolite. **RETURN** to vehicles, turn around and return to Haydenville Road.

91.3 (5.9) Turn right onto Haydenville Road.
92.275 (6.875) Outcrop of massive volcanics in ditch on right. Over the last mile we have been driving, more or less, parallel to the axis of the Whately Anticline.
92.5 (7.1) Masterson Road on the right; West Brook Road on the left; continue west on Haydenville Road.
92.6 (7.2) Park on the right side of the road in the pullout. Walk approximately 100m westward along the road and enter the pasture on the south side of the road through the gate.

**STOP 10. CORE OF THE WHATELY ANTICLINE**

This stop is a walk across the core of the Whately Anticline looking at the Hawley Formation where there is no quartzite. Several meters to the southwest of the shed are lenses of amphibolite caught up in the rusty schist. Under the northeast corner of the shed are excellent examples of ptygmatically folded coticule within the schist. Nearby are other amphibolites with shapes suggestive of pillow structures. Also to be seen are porphyritic amphibolites, and a metamorphosed biotite granite dike. Scattered throughout the pasture are dikes of the Williamsburg Granodiorite crosscutting the older rocks. Walk down to the brook at the southeast corner of the pasture and continue eastward across the fence. As one walks back up the knoll away from the brook one encounters massive amphibolite, then schist with coticule giving way to a silvery-gray, garnet, staurolite schist. If one continues further eastward toward West Brook Road one encounters again the massive amphibolite. **RETURN** to vehicles. Continue slowly westward.

92.8 (7.4) Turn right onto Conway Road.
92.85 (7.45) Outcrops on both sides of the road are in the Waits River Formation.
93.05 (7.65) Park along roadway, and watch out for the infrequent traffic.

**STOP 11. WAITS RIVER FORMATION**

The outcrop is along the streambed. **CAREFUL: THE ROCKS CAN BE SLIPPERY**. This is an outcrop of the Waits River formation on the west flank of the Whately anticline. There is massive impure marble here along with black schist, finely bedded and graded (?) sandstones, and a thin amphibolite; all that typify the Waits River. Note the percentage of carbonate and the knobby surface in some of the beds. The detailed structural relations are well displayed in that part of the outcrop bordering the south side of the brook as it turns westward. **RETURN** to vehicles. Turn around and return to Haydenville Road.

93.35 (7.95) Intersection of Conway and Haydenville Roads. Turn left and return towards the east.
93.6 (8.2) Turn right onto Westbrook Road.
93.7 (8.3) Cross West Brook and turn left to head east.
93.9 (8.5) Turn right onto gravel road and continue up the hill to the corner of roadway.
94.1 (8.7) Park in wide part of corner in the road.

**STOP 12. WAITS RIVER FORMATION**

This is a new outcrop uncovered during road construction and is in the Waits River Formation on the east flank of the Whately anticline. There is lots of punky, impure marble here along with some interbedded phyllite and thin quartzites. Graded bedding on the west side of the outcrop indicates that tops are to the west (opposite to that seen in the channel deposit). Also here is a thin bed containing quartz, plagioclase, biotite, garnet, a nearly colorless clinoamphibole, and zircon. It is interpreted as a felsic igneous rock, and it has been collected for zircon analysis. **RETURN** to vehicles and West Brook Road.

94.25 (8.85) Turn right onto West Brook Road.
94.7 (9.3) Cross West Brook and park on the left side of the road.

**STOP 13. GILE MOUNTAIN FORMATION**

We are well within the Gile Mountain Formation and will be looking at contact effects of the Hatfield Pluton on the Gile Mountain. The first part of this short traverse is to look at the contact metamorphism that has produced andalusite and sillimanite in the phyllitic beds. Then a little further to the northeast and away from the contact are sandy beds up to 14 cm in thickness intercalated with the phyllitic beds. The structures here are much more open then seen in most of our previous stops. An outcrop along the brook does show, however, good isoclinal folds, some of which have a southwesterly plunge. **RETURN** to vehicles.

94.9 (9.5) Approximate contact between Hatfield Pluton and the Gile Mountain Formation.
95.2 (9.8) Turn left into driveway. Walk in along wood road.

**STOP 14. HATFIELD PLUTON**

The first outcrops are those of the Hatfield Pluton. The tonalite is characterized by hornblende, plagioclase, quartz and minor biotite. Stoeck (1971) described the petrography of these plutonic rocks and the surrounding contact aureole and divided the pluton into an outer granodioritic zone and an inner tonalitic zone where biotite was "slightly more abundant than hornblende in the tonalite (Stoeck, p.1, 1971). This suite of rocks is correlated to the Belchertown Tonalite that occurs to the east of the Connecticut Valley, Mesozoic basin, and which has given a Devonian age of 380±2 Ma (Ashwal *et al.*, 1979). Along the road occurs one of the large xenoliths of the Gile Mountain Formation. It is a sillimanite, muscovite schist with minor biotite, and where garnet and staurolite are generally absent. We will also look at a hybrid rock that is rich in biotite and may either represent foundered schist or a biotite-rich facies of the Hatfield Pluton. **RETURN** to vehicles.

95.8 (10.4) Park in entrance to gravel pit.

**STOP 15. (OPTIONAL) HATFIELD PLUTON**

The interesting features at this stop are mafic inclusions that are easily seen in outcrop and the numerous shear bands that have developed within the pluton.

**END OF FIELDTRIP**

To return to Amherst:

Go to the end of West Brook Road [mile 95.9 (10.5)] and turn right onto North Street. Cross West Brook and the Hatfield-Whately town line. At mile 96.0 (10.6) continue past Mountain Road and at mile 96.1 (10.7) turn left onto Mountain Drive and continue slowly towards the east and pass over I-91 [mile 96.3 (10.9)] to the intersection with routes 5 & 10. If you turn left here you will return to Rte 116 which will take you to Sunderland and then Amherst. If you turn right you will go to Northampton where at the lights (Dana Chevrolet on hill to the right) you turn left onto Damon Road. Continue on Damon Road and turn left onto Rte 9 at the lights. This is the southern route into Amherst via Hadley.

**BONNE ROUTE**

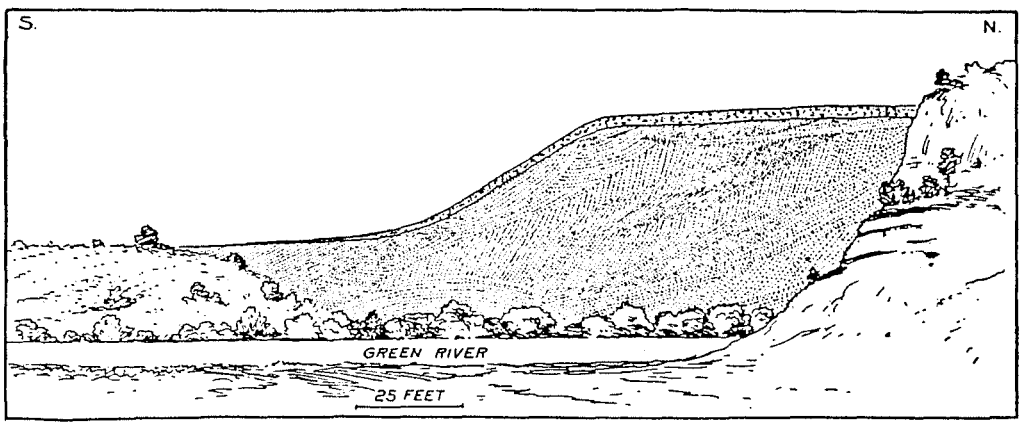



FIG. 36.—Section of the Green River delta at the north end of the Green River basin, where the stream comes out of its rocky canyon, showing that the delta was sent but little into the lake, and its front not eroded.

## REFERENCES CITED

Aleinikoff, J.N. and Karabinos, P., 1990, Zircon U-Pb data for the Moretown and Barnard volcanic members of the Missisquoi formation and a dike cutting the Standing Pond Volcanics, southeastern Vermont: U.S. Geological Survey bulletin 1887-D, 10p.

Ashwal, L.D., Leo, G.W., Robinson, Peter, Zartman, R.E., and Hall, D.J., 1979, the Belchertown quartz monzodiorite pluton, west-central, Massachusetts, a syntectonic Acadian intrusion: American Journal of Science, v. 290, p. 936-969.

Billings, M.P., 1956, The Geology of New Hampshire, Part II, Bedrock Geology: New Hampshire Planning and Development Commission, 203p.

Doll, C.G., 1944, A preliminary report on the geology of the Strafford quadrangle, Vermont: Vermont State Geologist, 24th Annual Report, 1943-1944, p.14-28.

Doll, C.G., Cady, W.M., Thompson, J.B., Jr., and Billings, M.P., compilers and editors, 1961, Centennial geologic map of Vermont: Vermont Geological Survey, Montpelier, Vt., Scale 1:250,000.

Emerson, B.K., 1898, Geology of old Hampshire County, Massachusetts, comprising Franklin, Hampshire, and Hampden counties: U.S. Geological Survey Monograph 29, 790 p.

Emerson, B.K., 1917, Geology of Massachusetts and Rhode Island: U.S. Geological Survey Bulletin 597, 289 p.

Hatch, N.L., Jr., Stanley, R.S., and Clark, S.F., Jr., 1970, the Russell Mountain Formation, a new stratigraphic unit in western Massachusetts: U.S. Geological survey Bulletin 1324-B, p.B1-B10.

Hepburn, C.J., Trask, N.J., Rosenfeld, J.L., and Thompson, J.B., Jr., 1984, Bedrock geology of the Brattleboro quadrangle, Vermont-New Hampshire: Vermont Geological Survey Bulletin No. 32, 162p.

Hitchcock, E., Hitchcock, E., Jr., Hager, A.P., and Hitchcock, C., 1861, Report on the geology of Vermont, descriptive, theoretical, economical, and scenological: Volumes 1 and 2, Claremont, N.H., Claremont Manufacturing Co., 982p.

Hueber, F.M., Bothner, W.A., Hatch, N.L., Jr., Finney, S.C., and Aleinikoff, J.N., 1990, Devonian plants from southern Quebec and northern New Hampshire and the age of the Connecticut Valley trough: American Journal of Science, v.290, p.360-395.

Richardson, C.H., 1927, The geology and petrography of Barnard, Pomfret, and Woodstock, Vermont: Vermont State Geologist, 15th Annual Report, 1925-1926. p.127-159.

Robinson, Peter, 1963, Gneiss domes of the Orange area, Massachusetts and New Hampshire: Cambridge, Mass., Harvard University, PhD thesis, 253p.

Robinson, Peter, Hatch, N.L., Jr., and Stanley, R.S., 1988, the Whately thrust: A structural solution to the stratigraphic dilemma of the Erving Formation: U.S. Geological Survey Professional Paper 1366-d, 22p.

Spear, F.S. and Harrison, T.M., 1989, Geochronologic studies in central New England I: Evidence for pre-Cambrian metamorphism in eastern Vermont: GEOLOGY, v. 17, p.181-184.

Stoeck, P.L., 1971, Petrology and contact effects of the Hatfield Pluton of Belchertown tonalite in the Whately-Northampton area, western Massachusetts: Amherst Mass., University of Massachusetts, M.S. thesis, 83p.

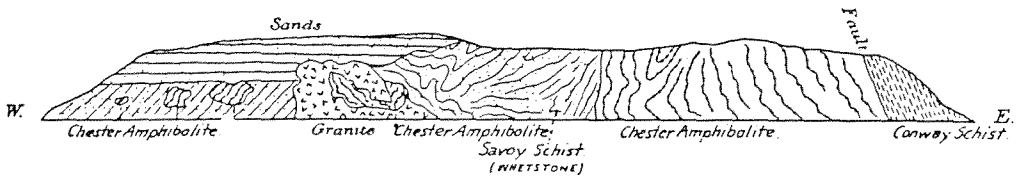



FIG. 13.—Section on railroad east of Erving station.

# ICE-WEDGE CASTS, PINGO SCARS, AND THE DRAINAGE OF GLACIAL LAKE HITCHCOCK

by

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and
J.P. Schafer, formerly of the U.S. Geological Survey, now deceased
(to whom we dedicate this fieldtrip)

## INTRODUCTION

A harsh postglacial climate, in which permafrost was present, existed for several thousand years following deglaciation until after the drainage of glacial Lake Hitchcock. On this fieldtrip (fig. 1) we will examine paleoclimatic evidence in the form of ice-wedge casts, pingo scars, eolian deposits, cryoturbation structures, and the paleobotanical record, that provides evidence for this harsh climate as well as new evidence concerning the timing of the drainage of Lake Hitchcock.

Wedge-shaped structures developed in glacial meltwater deposits capped by eolian mantle have been exposed for many years at the Hain Brothers pit in Windham CT. We will examine evidence (including their demonstrable polygonal ground pattern) which argues strongly for the interpretation of these structures as ice-wedge casts indicative of permafrost conditions. We will examine the stratigraphy of the glacial meltwater deposits at depth in the section, the deformational characteristics of the linear wedge structures that cut the surface of these deposits, and the "pedogenic soil" (Agawam series) developed on the cryoturbated eolian material that mantles the surface and penetrates the wedge structures. Evidence will be presented that argues against former interpretations of the structures as 1) "pseudo-ice-wedge casts" resulting from tensional faulting unrelated to permafrost (Black, 1983) and 2) "earthquake-induced fissures" that served as conduits for liquified sediment (Thorson and others, 1986).

On lake-bottom surfaces of glacial Lake Hitchcock we will visit several localities where clusters of rimmed depressions are interpreted as pingo scars that formed as a result of the growth and decay of permafrost lens ice on the lake-bottom surface following drainage of the lake (Stone and Ashley, 1989; Stone and others, 1991). These features occur extensively on surfaces below paleo-lake level that were not fluvially terraced by the postlake stream incision. We will visit the Kelsey-Ferguson clay pit, excavated in a terraced lake-bottom surface that has no pingo scars but does exhibit well-developed eolian morphology in the form of dunes and deflation hollows.

We will present and discuss new $^{14}C$ dates and plant macrofossil and pollen assemblages from the upper section of Lake Hitchcock lake-bottom sediments and from the clastic and organic fill of the pingo-scar depressions. Dates from the pingo-scar peat clearly demonstrate drainage of glacial Lake Hitchcock at least 2000 years before the 10,700-yr BP date of Flint (1956), and dates on detrital plant debris in the upper lake beds indicate that the lake probably drained nearly 2000 years before the suggested 12,400-yr BP date of Ridge and Larsen (1990).

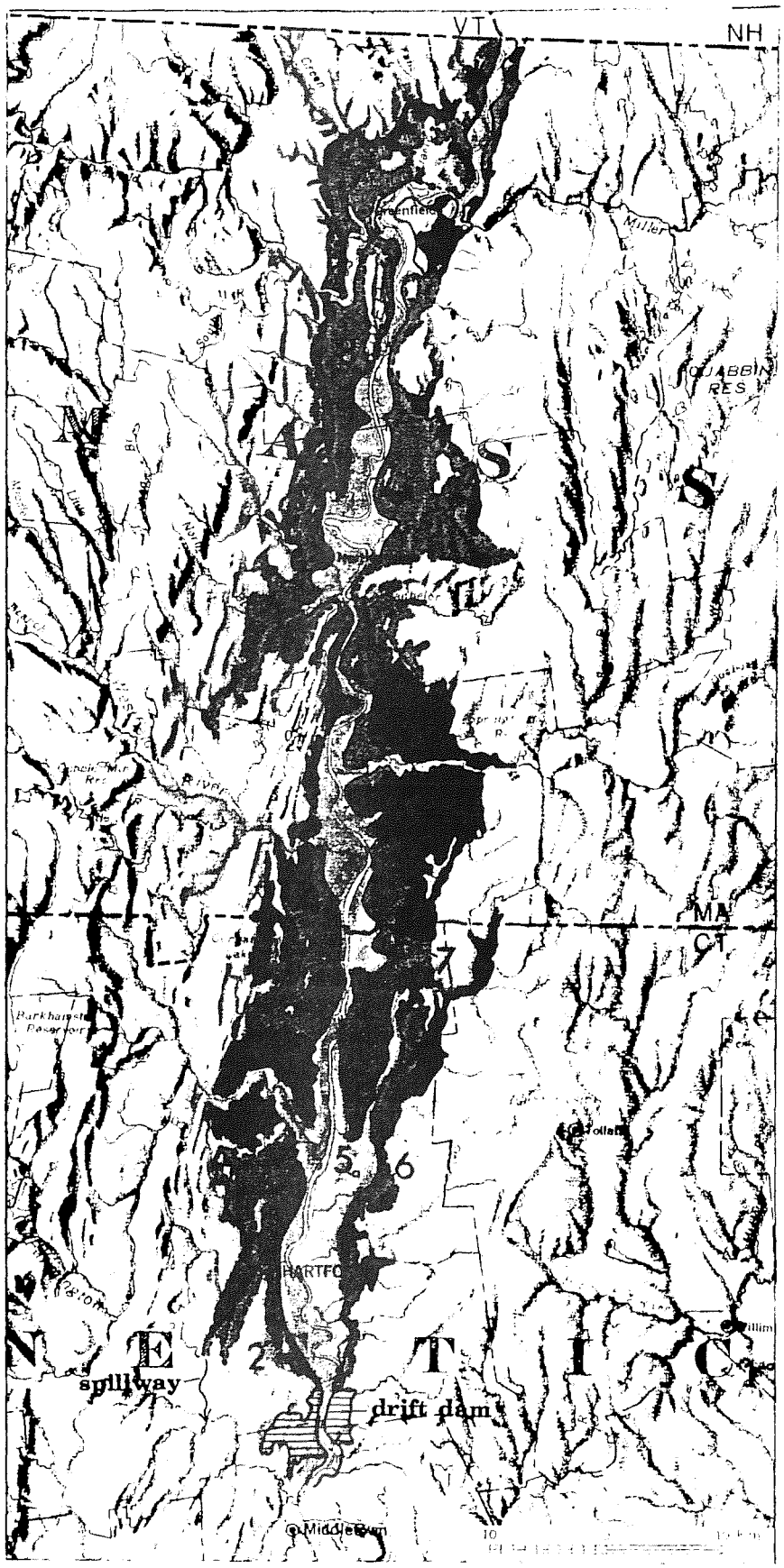


**Figure 1.** Glacial Lake Hitchcock surfaces in Connecticut and Massachusetts. Pingo scars occur extensively on the lake-bottom surfaces. Numbered localities are FIELDTRIP STOPS.

delta surfaces lake-bottom surfaces post-lake incised terrace surfaces

## ICE-WEDGE CASTS AND POLYGONAL PATTERNED GROUND

Wedge-shaped structures that deform late-Wisconsinan glacial sediments have been found, mostly in solitary occurrence and without demonstrable polygonal ground pattern, throughout southern New England and have been interpreted by a number of workers (fig. 2) as ice-wedge casts indicative of former permafrost conditions. The wedge structures were first suggested to be ice-wedge casts by J.P. Schafer (Schafer and Hartshorn, 1965; Schafer, 1968). For many years the best exposure of multiple structures has been in the Hain Pit in Windham CT, where a number of workers have both described and interpreted the structures in different ways. They are clearly linear, vertical, filled extensional fractures that deform the upper few meters of an eolian-mantled glacial meltwater deposit.

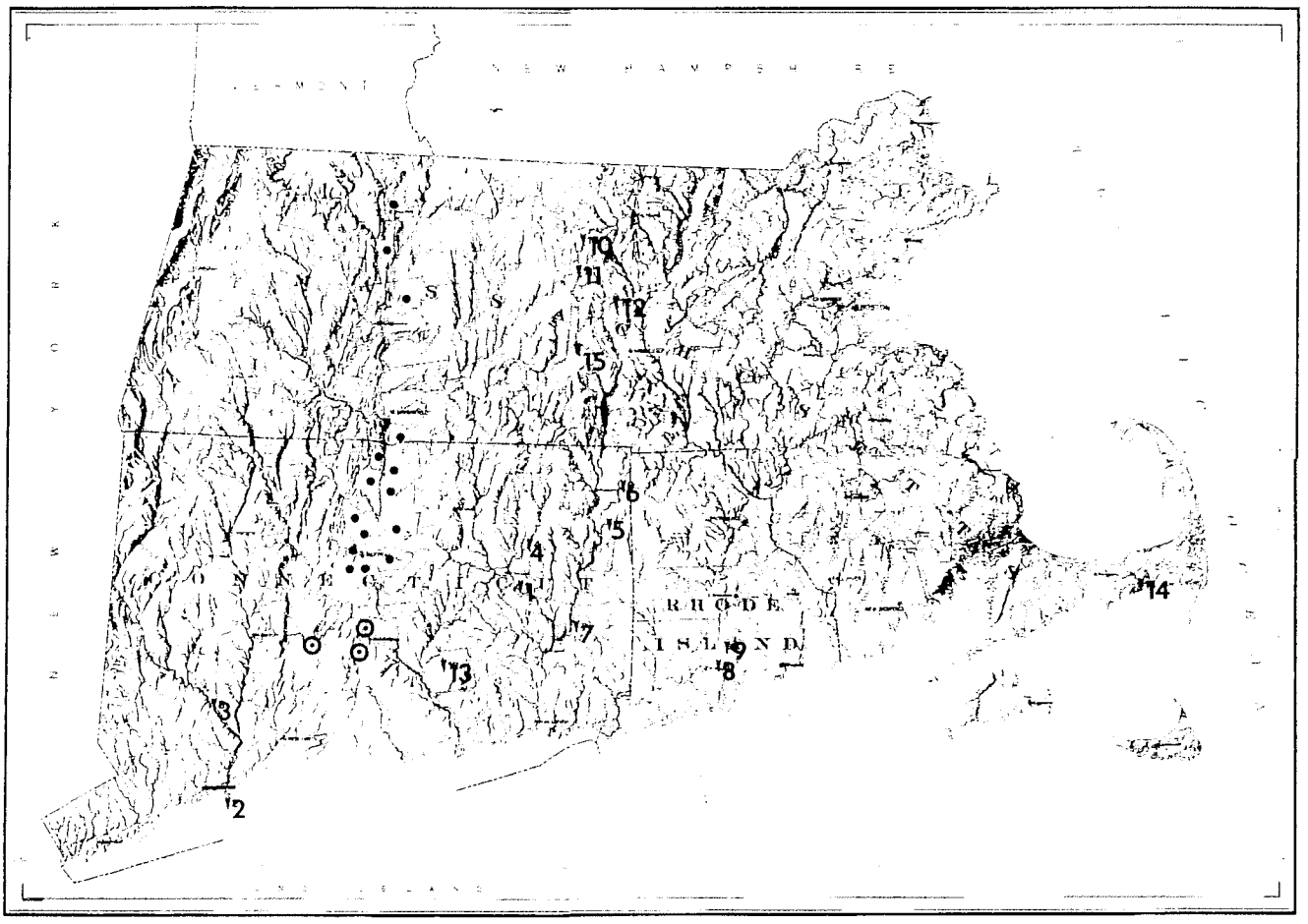


**Figure 2.** Locations of features indicative of former permafrost in southern New England.

▼ **Ice wedge casts-** Number keyed to references below
1 Hain Pit, South Windham, CT (this paper)
2 Lordship, Stratford, CT (Denny, 1936)
3 Seymour, CT (Schafer, 1968)
4 Bedlam Corner, Chaplin, CT (Schafer, unpublished)
5 Dayville, CT (E.H. London, unpublished)
6 Munyon Road, Thompson, CT (Schafer and Hartshorn, 1965)
7 Jewett City, CT (B.D.Stone, 1978)
8 University Pit, Kingston, RI (Hartshorn and Schafer, 1965)
9 W. Allentown Rd., N. Kingston, RI (Boothroyd and Lawson, 1989)
10 Hubbardston, MA (J.R. Stone, unpublished)
11 Barre, MA (Larsen, 1979)
12 Holden, MA (B.D.Stone, unpublished)
13 Moodus, CT (O'Leary, 1975)
14 Dennis, MA (Oldale, 1974)
15 E. Brookfield, MA (Stone and others, this volume)

● **Pingo-scar clusters-** on drained Lake Hitchcock surfaces
⊙ **Pingo-scar clusters-** on other glacial lake surfaces

In cross-section (fig. 3), the structures are vertical and wedge-shaped, up to 1.0 m wide at the top and tapering downward to termination commonly at depths of 3-4 m. Deformed wall strata attenuate downward in steeply nested V's and become (by gradual loss of stratification) the nonbedded central cores of the filling. The cores are characterized by vertically oriented clasts. The upper part of filling consists of tongues or plugs of eolian sandy silt containing ventifacted pebbles. At its base this material is often gray (nonoxidized) and displays crude concave-upward layering. The eolian material commonly penetrates the upper 1-2 m of the wedge; it penetrates as deep as 3 m in places along an east-west striking feature currently being excavated at the north end of the Hain pit. Where the deformation cuts across wall beds of medium to coarse sand, rather than gravel, normal faults bound the fracture and the sand may be deformed downward into the filling by a series of normal faults. In some cross-sectional exposures, particularly in the sandy wall-bed material, thickened and/or upturned bedding and reverse faults are present on both sides of the normally downdropped fill. This indicates initial pressure outward from the center before inward collapse. The above cross-section description and (fig. 3) sketch apply to views where the pit face is vertical and perpendicular to the strike of the wedge. It is important to note that in situations where this is not the case, the wedge structure will appear to be asymetric and non-vertical. In situations where the pit face is oblique or parallel with the wedge strike, and in places where the pit face cuts wedge intersections, complex configurations occur which are easy to misinterpret.

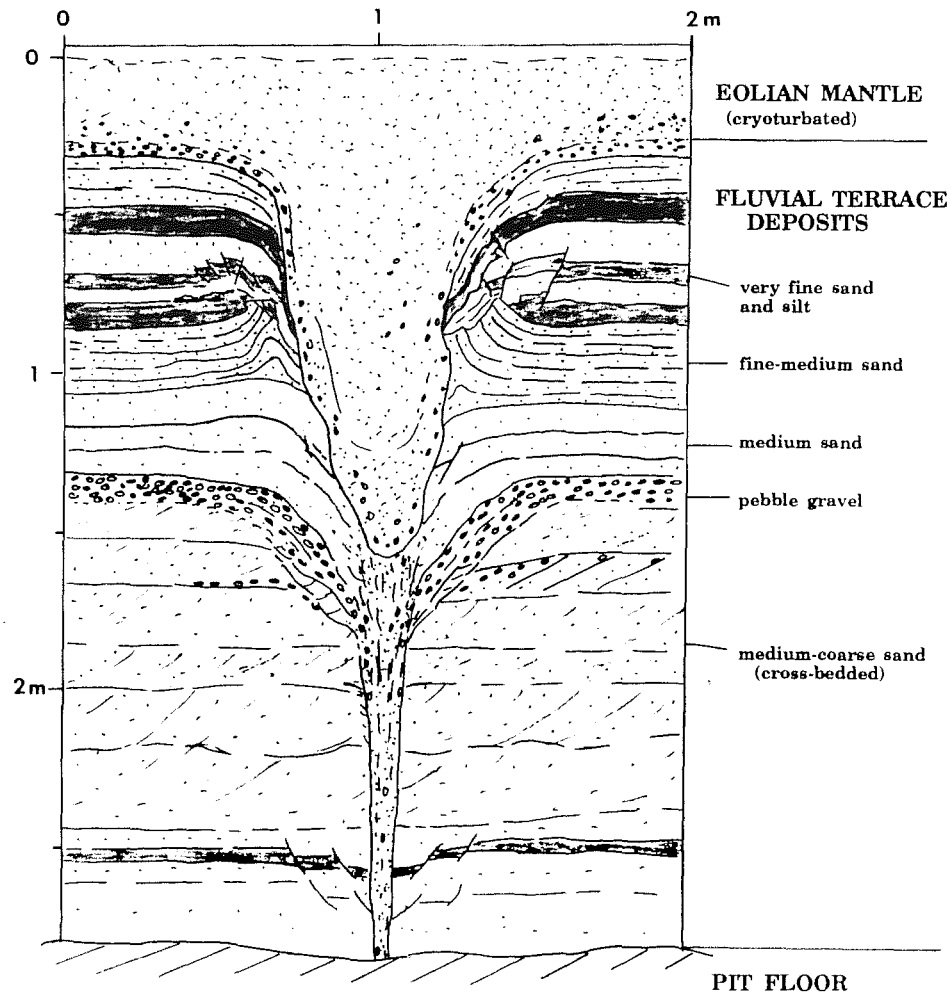


**Figure 3.** Sketch of ice-wedge cast, Hain Pit, Windham, CT.

In plan view (fig. 4), the wedge structures are linear, filled fractures that locally intersect. Figure 4E (from Oakes, 1992) shows all recorded strike directions for wedges in the Hain Pit including those measured by Thorson and others (1986). The nature of their horizontal pattern has been seen in a limited way by recording wedge traces through continuing excavation of pit faces (Thorson and others, 1986) and at depth by traces on small areas of the current pit floor.

On some aerial photographs of the pit area, traces of the wedges can be seen on surfaces stripped of topsoil prior to excavation. Only small surface areas in this condition have been observed, thus resulting in only an incomplete and speculative assessment of the size and character of the polygons bounded by the wedge fractures. Recently, close examination of 1980 airphotos (FL-69, 4198-99) in the Hain pit area has revealed a polygonal ground pattern on a 250-m by 180-m field located 200 m north of the Hain Pit (fig. 4A and B). At the time these photos were taken (April 19, 1980) this particular field had optimum conditions to allow higher

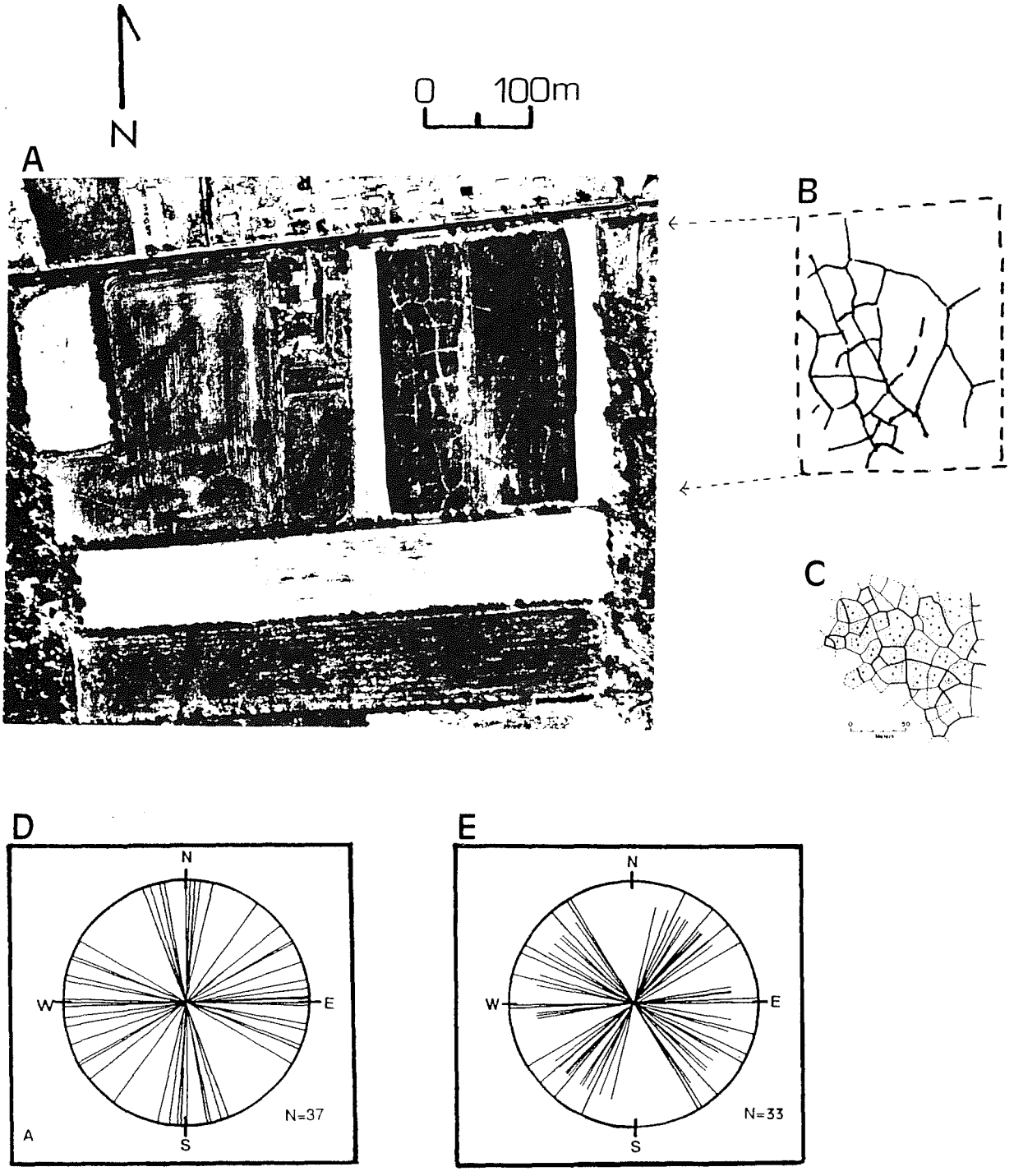


**Figure 4.** **A)** Section of airphoto (April 1980, FL-69, 4198-99) showing polygonal ground pattern on field north of Hain Pit, Windham, CT. **B)** Ground pattern traced directly from original airphoto. **C)** Polygonal ground pattern on sand and gravel surface in Barrow, Alaska (from Black, 1974, fig. 23). **D)** Strike directions measured from patterned ground on field north of Hain Pit (from Oakes, 1992). **E)** Wedge-strike directions recorded from Hain Pit (from Oakes, 1992).

moisture content in the finer-grained eolian material of the upper wedge fill to be visible from the air. Similar conditions allow relict patterned ground in central Illinois to be visible on airphotos (Johnson, 1990). Patterned-ground polygons on the field north of the Hain pit have diameters of 20-50 m; these polygons appear to be secondary to a larger, 150-180-m-diameter polygon. Strike directions on the field pattern plotted from airphoto measurements are shown in figure 4D (from Oakes, 1992). One can compare the configuration and size of polygons with those on a surface underlain by sand and gravel at Barrow, Alaska (fig. 4C) from Black (1974, fig. 23).

### The Pseudo-Ice-Wedge Cast Interpretation

The strongest opponent of permafrost genesis of wedge-shaped structures in New England glacial deposits was Robert F. Black. Black argued that the postulated thermal environment during deglaciation of Connecticut (i.e. wet-based nature of the ice sheet, rapidity of its retreat, nearness to the ocean, and latitude) precluded the possibility that permafrost could have existed for 1,000 yrs or more during deglaciation (Black, 1983). Black described the wedge structures at the Hain Pit and the Bedlam Corner pit 12 km to the north as "pseudo-ice-wedge-casts because they are deceptively similar to true ice-wedge casts". He argued that these features did not meet the criteria as defined by Black (1976) to be interpreted as permafrost features. His reasons included; 1) commonly solitary nature of wedge occurrence, 2) lack of polygonal ground pattern with appropriate polygon diameters for the host material, 3) absence of pressure effects and upturned wall beds in at least some of the structures within a group, and 4) absence of other permafrost indicators within the region. Black also described what he believed to be evidence for relatively recent deformation associated with these structures because at least one of them appeared to have B-horizon soil material deep within the wedge filling (he cites Harvey Luce, oral comm., 1979), and because many of the wedge structures cut (and therefore postdate) horizontal iron-stained layers. Black concluded that the wedge structures in Connecticut are most likely tensional fractures produced by the loading of coarse clastics on fine clastics, and are associated with a lowering water table produced by postglacial stream incision; he concluded that deformation continued into modern times.

### The Earthquake Interpretation

Interestingly, sediment deformation produced by permafrost may resemble that produced by earthquakes. Thorson and others (1986) interpreted the wedge structures at the Hain pit as earthquake-induced ground fissures. They describe the sediment filling, in particular the upper plug material of massive, gray silty sand, as subjacent lacustrine material that liquified during seismic excitation, was vented to the surface, and then flowed into the upper part of the ground fissures. Thorson and others (1986) also believed that the structures were significantly younger than the glacial sediments in which they occur; they point out, as did Black (1983), that the wedge structures cut strongly iron-stained layers that result from prolonged contact with aerated ground water to produce the oxidation. They suggest that the seismic event which produced the fissures and liquified sediment fill was clearly prehistoric but may have occurred within the past few thousand years. A 1050$\pm$100-yr BP date on slightly organic silty colluvium associated with the surface of one structure is used to substantiate the recent age of these features; however, the context and interpretation of this date, as presented, are extremely nebulous.

### The Ice-Wedge Cast Interpretation

The wedge-shaped structures at the Hain pit and patterned ground on the field to the north are here interpreted as ice-wedge casts indicative of permafrost conditions in eastern Connecticut following retreat of the ice sheet from that area about 16.5 ka. These features can now be shown to possess all of the requisite characteristics identified by Black (1976) and other authorities on past and present permafrost to be classified as ice-wedge casts. These characteristics include:

1) similarity to ice-wedge casts found in northern latitudes today in areas of degrading permafrost
2) large-diameter polygonal ground pattern (20-50 m), which is consistent with the thermal expansion properties of sand and gravel
3) upturned wall beds and pressure effects preserved within wall-bed sediment in some of the structures which provide evidence for ice expansion

4) gently downwarped wall beds in nested V-shaped layers that are evidence for slow grain-by-grain replacement of ice with sediment in the fracture
5) upper part of wedge filling consists of downdropped eolian material that commonly contains ventifacted pebbles and displays crude concave upward layering that suggests filling of an open contraction crack within the seasonally thawed active zone, and
6) presence of complex soil involutions clearly formed by cryogenic processes under severe cold-climate conditions although not necessarily in permafrost; these commonly "pot-shaped" cryoturbation structures occur separately from the wedges, but also in places deform the wedge structures.

None of the many wedge structures currently exposed in eastern Connecticut involve deformation of the modern soil horizon. Dr. Harvey Luce (professor of soil science, University of Connecticut) has examined many wedge fillings at the Hain pit as well as the normal soil profile (see STOP 1, section f). He describes the pedogenic soil as Agawam series developed in relatively thick (up to 40 cm) eolian mantle overlying well-drained sand and gravel. The normal soil profile crosses the wedge features therefore, soil formation postdates the wedge deformation. Strongly oxidized layers in the host sands and gravels appear to be cut by the wedge structures and associated deformation. Prolonged exposure to aerated ground water is required to produce the oxidation, as stipulated by Black, (1983) and Thorson and others, (1986). However, the oxidation takes place in the coarser-grained areas where permeability is high regardless of the bedding configuration of those areas. This point is demonstrated by the oxidation within the coarsest material in complex sedimentary structures in these same beds. The oxidation postdates the sedimentary structures and the wedge deformation as well.

Another requisite factor defined by Black (1976) for the interpretation of wedge structures as ice-wedge casts is that there must be other cold-climate indicators in the region. Such evidence is present, in the form of pingo scars, extensive eolian deposits, and tundra vegetation, and is discussed below.

## PINGO SCARS ON DRAINED GLACIAL LAKE-BOTTOM SURFACES

Clusters of circular to subcircular shallow depressions, many with subtly raised rims, occur on drained glacial lake-bottom surfaces in Connecticut and Massachusetts (figs. 1 and 2). They are interpreted as pingo scars (Stone and Ashley, 1989; Stone and others, 1991) based on; 1) their striking resemblence in morphology and distribution to ramparted depressions identified as pingo scars and other ground-ice depressions in northwestern Europe (Svensson, 1969; Mitchell, 1973; De Gans, 1988), 2) the deformational character of their internal structure as seen in a cross-sectional exposure (fig. 5, Stop 3) and on ground penetrating radar (GPR) lines across the surface features, and 3) the age and character of the clastic and organic fill within the deformational structures.

The surface depressions generally have less than 10 ft (3 m) of relief and therefore do not show up well on 1:24,000-scale topographic maps with 10-ft contour interval. They were not reported or described on any of the detailed surficial geologic maps of the area, but were first noticed during compilation and synthesis of the detailed maps for the State Quaternary Geologic Maps of Connecticut (Stone and others, 1992; Stone and Schafer, in prep.). Comprehensive and detailed air-photo analysis has revealed these clustered depressions to be characteristic of former lake-bottom surfaces, the most extensive of which are those of glacial Lake Hitchcock.

### Morphology and Distribution

These landforms occur predominantly on surfaces underlain by fine-grained lake-bottom sediments (fine sands, silts and clays). Locally they are found developed on till and sandy spit-deposit surfaces on the sides of drumlins that were islands in Lake Hitchcock. They occur only below paleo-lake level. In places, they are closely associated with eolian deposits and dunes. They are not found on delta surfaces which were above paleo-lake level; they are absent on fluvially-modified post-lake surfaces such as stream terraces and floodplains. They occur in clusters of isolated and mutually interfering forms. Maximum densities of pingo scars are approximately 150/km$^2$. Diameters are generally from 20 to 40 m, but range from 5 to 100 m; one depression has a diameter of 250 m. The surface depressions are generally 1-3 m deep. Rims range from 0.3 to 1.5 m high. They are clearly visible on large scale maps with contour

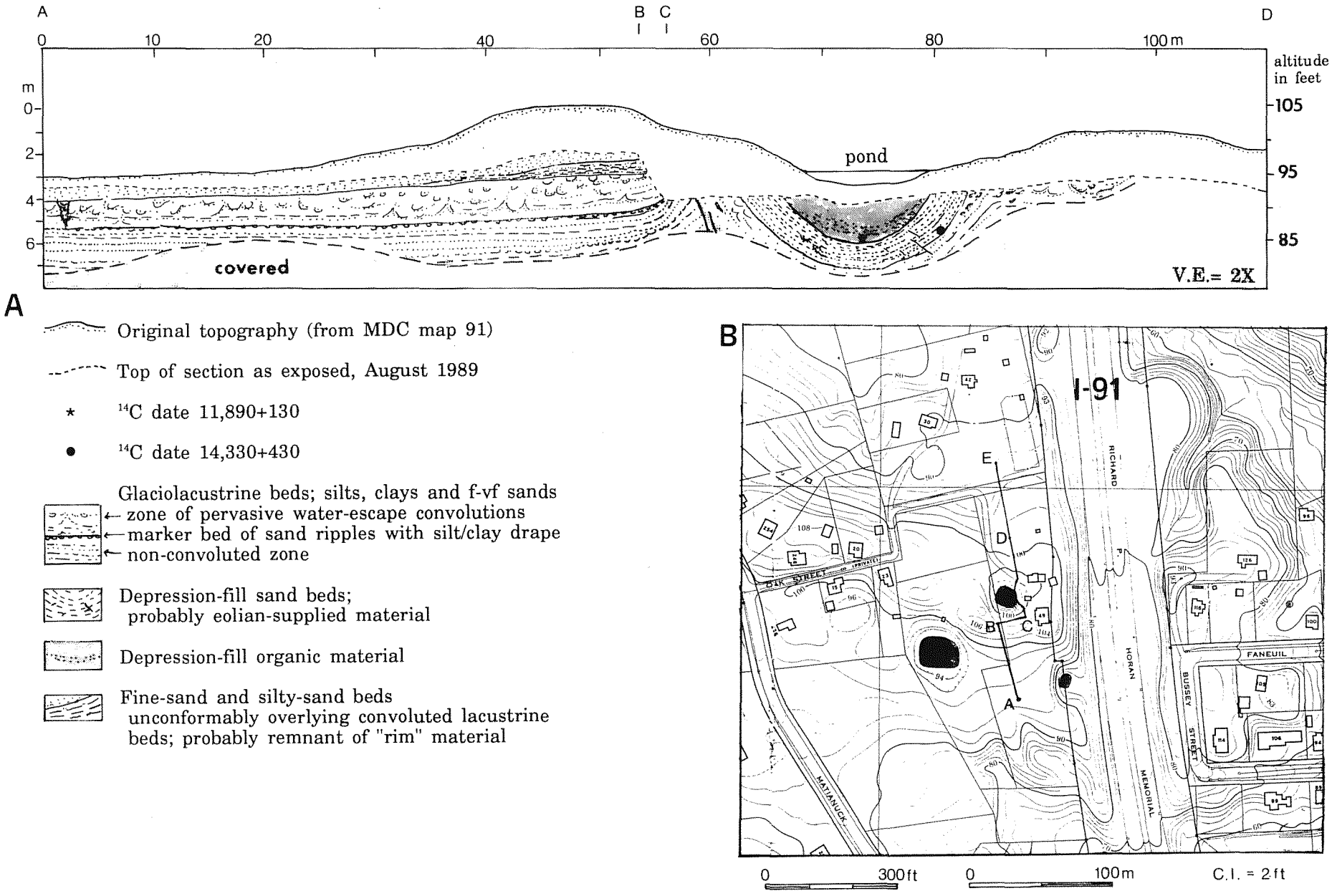


**Figure 5. A)** Measured Section at I-91 roadcut north of Hartford CT. **B)** Section of MDC Map 91 (1977 ed.) showing topography of area and location of section line (A - E).

intervals of 1 and 2 ft, such as the Metropolitan District Commission (MDC) 1:2,400-scale maps (figs. 5B, 9B, and 12) and (fig. 12, Stop 6) and on 1:12,000-scale air photos. Closed depressions that contain small ponds, or wetlands/vernal pools are common, as well as many features in which the rims are breached and C-shaped contours mark the pingo scars.

### Internal Structure

Stratigraphy and structure beneath the surface depressions are known from a roadcut exposure on Route I-91 between exits 34 and 35 (fig. 5 and Stop 3). Ground penetrating radar (GPR) data supplemented by vibracores across several other depressions add to our knowledge of the internal structure. The following internal characteristics (fig. 5) have been observed:

1) Beneath the outer edges and rim of the surface depression, upper lacustrine fine-sand and silt beds are gently deformed upward as much as 1.5 m and are broken by fractures. Inward from the fracture zone, lacustrine beds are intensely collapsed as much as 4 m downward. Beneath the central part of the depression, lacustrine materials are completely homogenized as a massive gray silt.
2) The structural depression created by collapse of the lacustrine beds is filled with a bowl-shaped body of medium sand in concentrically dipping beds. These beds display small-scale reverse faults displacing them downward toward the center of the structural depression and indicate that filling was synchronous with later stages of collapse of the underlying lacustrine materials. The sand is coarser grained than most of the lacustrine sediment. This fact and the nature of the bedding indicates that the sand fill is probably eolian-supplied material that settled out in ponded water in the collapsing depression.
3) Further filling of the structural depression consists of a 2-m-thick pod of gyttja and peat above the sand and below the center of the surface depression. Sandy clastic lenses extend into the peat body from the sides.

### Age and Paleobotanical record

Three $^{14}C$ dates were obtained from fill material of the I-91 pingo scar. The lower part of the eolian sand body contained detrital wood fragments which yielded a date of **14,330±430 yr BP** (Beta-35211); the wood was identified as *Salix* (willow). The upper lacustrine material just below the sand fill also contained plant debris. As yet we have not obtained a date on this material, but plant macrofossils identified by Lucinda McWeeney, Yale University include: Twigs- *Vaccinium oxycoccos* (small cranberry), *Vaccinium uliginosum* (bilberry); Seeds- *Empetrum nigrum* L. (crowberry), *Potamogeton sp* (flat stemmed pondweed), *Ranunculus sp* (buttercup), *Vaccinium sp* (mountain cranberry or alpine bilberry), *Cyperus sp* (flat sedge), *Carix sp* (sedge). It is possible that the plant material in the eolian pingo-scar fill may have been redeposited from the upper lake beds during the uplift and collapse of the pingo deformation.

A sample from the lowermost section of the peat body yielded a date of **11,890±130 yr BP** (Beta-34820). A second (AMS) date of **12,050±110 yr BP** was obtained from the basal peat sample on several spruce needles and a cone bract. Macrofossil plants and pollen types (identified by L. McWeeney and D. Peteet) include: Needles- *Abies balsamea* (balsam fir), *Picea sp* (spruce), cf *Juniperus* (juniper), *Larix* (larch); Seeds- *Picea sp* (spruce); Bracts- *Abies balsamea* (balsam fir), *Betula sp* (birch); Leaves- *Vaccinium sp* (bilberry or cranberry); Twigs- *Larix* (larch), *Betula sp* (birch), Ericaceae- heath family, *Myrica gale* (sweet gale), *Shepherdia canadensis* (soapberry); Other small plants- *Gaultheria procumbens* (mountain tea), *Vaccinium oxycoccos* (cranberry), *Cyperaceae* (sedge); Other plant parts- Nymphaceae (rhizome fragments-water lilly), *Equisetum sp* (stems-horsetail), Gramineae- stems grass sp, *Chara fragilis* (green algae), *Mycorrhizal sclerotia* (fungi); Animal- beetle exoskeleton fragments, *Daphnia* (water flea); Pollen- poplar, birch conifer, willow, alder, and spruce.

Four $^{14}C$ dates were obtained by consultant geologists with EBASCO who were investigating the possibility that the rimmed depressions might be of seismogenic origin. These dates were on samples from vibracores through the fill of pingo scar F, a 50-m-diameter depression located approximately 1 km west of the I-91 cut (STOP 3). Five cores were obtained; GPR lines across the depression aided in interpretation of the vibracores. Core F-7 located in the center of the depression penetrated to a depth of 4 m, and revealed 2.7 m of brown\black peat with minor clastic sediment over 1 m of thinly bedded yellowish brown very fine sand\silt with minor peaty interbeds over 0.3 m of nonbedded gray silt. The lowest thin peat layer in the core yielded a date of **12,630±240 yr BP** (Beta-46514); the base of the continuous peat body yielded

a date of **12,350±110 yr BP** (Beta-46513). Two other shallower cores F-3 and F-5 penetrated about 2 m into the peat body; a date of **8,325±100 yr BP** (GX-17050) was obtained from core F-5; a date of **7,850±220 yr BP** (GX-17053) from core F-3.

A date of **12,200±250 yr BP** (W-828) was reported by Colton (1960) from "a small peat bog" excavated during construction of the NE runway at Bradley International Airport; this feature is now known to have been one of the many pingo scars in that area.

**Genesis of drained Lake Hitchcock Pingo Scars**

We suggest the following scenario at the time of pingo formation on drained Lake Hitchcock bottom surfaces probably between 14 and 13 ka. The ice had left the area several thousand years earlier, but the climate remained cold; cold enough to support at least discontinuous permafrost in areas favorable to its development, such as drained lake beds (Mackay, 1979, 1988a, 1988b). Permafrost began to develop on the supersaturated lake-bottom surfaces as soon as the lake drained (fig. 6). As permafrost aggraded downward in the lacustrine sediments pore water was put under hydrostatic pressure. Hydraulic pipes piercing the permafrost released the pressure by localized upward migration of water which turned to ice at the base or within fractures in the permafrost. The repeated addition of ice mass by freezing ground water and the 9% expansion due to the freezing process gradually built a dome of ice causing the sediment cover to be stretched thin over the top. Mass movement in the active layer during summer months moved some of the sediment downslope.

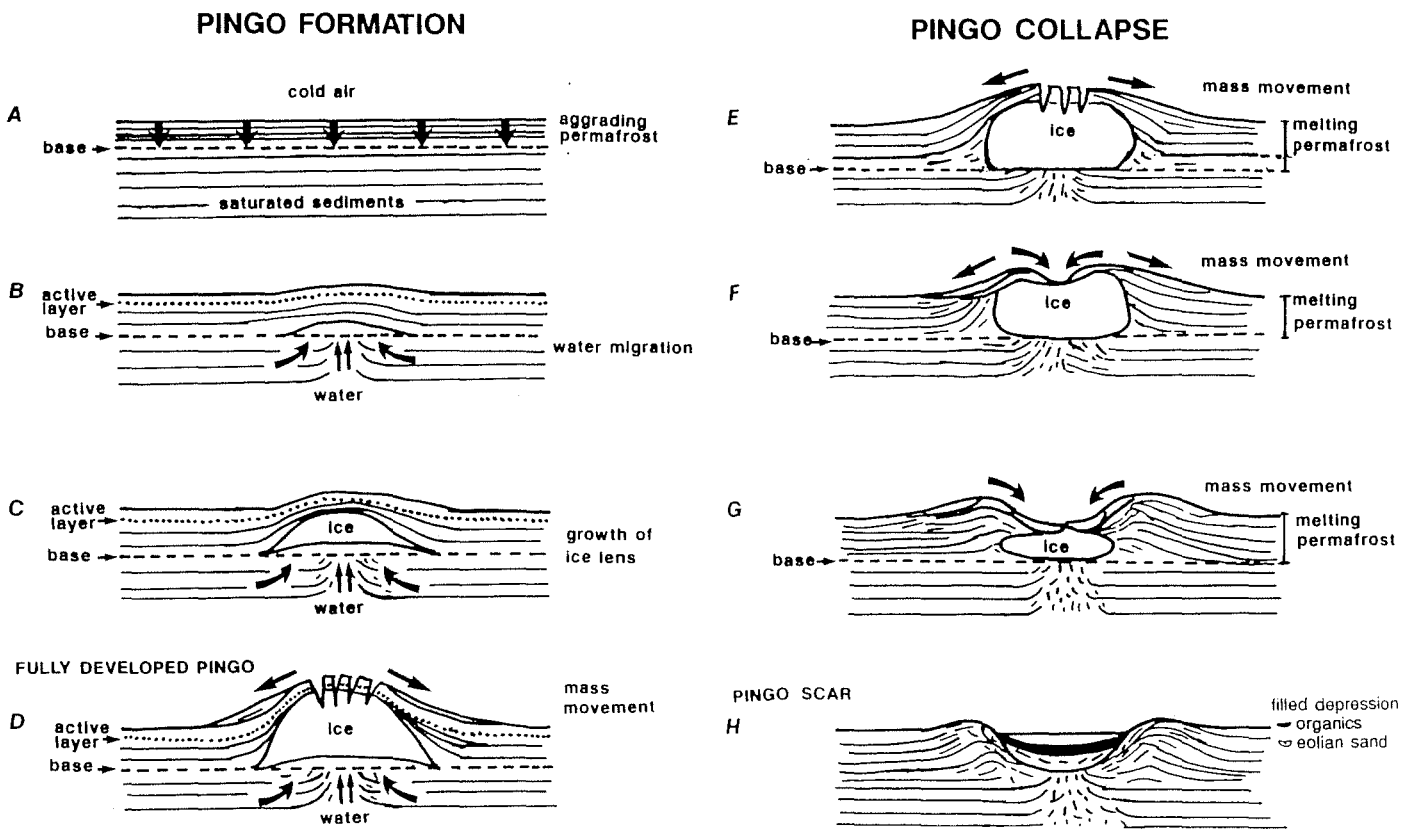


**Figure 6. Pingo formation and collapse.**
Aggrading permafrost (A) places ground water under pressure (B) which is released through pipes (C) and accumulates (freezes) at the base of permafrost and in cracks within the frozen sediment (D). Sediments immediately below the pingo are deformed and sediments on the surface (in the active layer) are subjected to mass movement. Mass movement continues both into and away from the pingo dome as the permafrost and pingo ice core melt from the surface downward (E, F, and G). The pingo scar (H) is a bowl-shaped depression enclosed by a raised rim which later may be partially filled with eolian sediment or organic material (peat).

Based on size and shape of modern analogs, the 20 to 40-m-diameter pingos might have grown to be 6 to 15 m high. Pingo growth rates on catastrophically drained lakes on Tuktoyaktuk Penninsula, Northwest Territories (Mackay, 1990) yielded a minimum time period of 100 years to generate a pingo of 15 m height.

Before about 12.5 ka the climatic regime had changed such that the pingo ice cores melted creating an inversion of topography (fig. 6E-H). The rampart that built up at the base of the ice-covered dome by colluvial processes became the rim enclosing the depression. Soon after the collapse, wind-transported sand partially filled some of the ponded depressions, followed by predominantly organic deposition after 12.5 to 12.0 ka.

## THE DRAINAGE OF GLACIAL LAKE HITCHCOCK

Lake Hitchcock was the longest-lived of the southern New England glacial lakes and the inference of permafrost continuing until after lake drainage has important postglacial paleoclimatic implications. Because the dam for the lake was composed of sediment rather than ice, and its spillway was across a bedrock drainage divide rather than the drift dam (fig. 1), the lake persisted for several thousand years. The life of the lake began with the ice-marginal emplacement of the Cromwell-Rocky Hill delta complex (i.e. drift dam) graded to glacial Lake Middletown, a precursor in the valley to Lake Hitchcock. The details of the early lake history have been presented in Stone and others (1982) and Koteff and others (1988). Correlation of regional $^{14}C$ dates (Stone and Borns, 1986) places the ice margin in central Connecticut and the inception of Lake Hitchcock at about 16 ka. Sequential ice-marginal deltas graded to Lake Hitchcock record its existence in the valley during the time of ice retreat from central Connecticut to northern Vermont, a distance of about 300 km. Ice-marginal deltas in the southern part of the lake record high lake levels, between 115 ft (35 m) and 90 ft (27 m) in altitude, at the spillway; meteoric-water-fed deltas in the south and ice-marginal deltas from Chicopee, MA northward record a stable lake level at 82-ft (25 m) in altitude at the spillway (Koteff and others, 1988). The lake drained due to breaching of the drift dam at a time after the ice margin had left the northerly reaches of the Connecticut valley.

For many years, the time of Lake Hitchcock drainage was accepted to be approximately 10.7 ka based on dates and interpretations presented by Flint (1956). The five $^{14}C$ dates >12.0-12.6 ka from basal organic fill of the rimmed depressions on the Hitchcock lake bottom clearly demonstate lake drainage before that time (12.6 ka); the depressions (regardless of their mode of origin) could not have formed until after the lake drained. The ~14-ka date on detrital wood in the eolian pingo-scar fill (I-91 cut, fig. 5) possibly indicates an even earlier time of drainage, but the twigs may have been redeposited in the fill from underlying lake beds. A recent exposure of upper lake beds 1 km north of the I-91 cut in Windsor, CT (Matianuck Ave. site, fig. 10) has provided more evidence for a ~14-ka time of lake drainage.

### Matianuck Avenue Site, Windsor, CT

Although the lake-bottom surface in the area of the Matianuck Ave. site displays many pingo-scar depressions, the section of the roadcut available at the time it was studied did not cross any of these features. It did reveal approximately 4 m of upper lake-bottom section (see fig. 10, Stop 4). Thin layers of plant debris interbedded with fine sand and silt layers were found 0.5 m above varved beds and 2.5 m below the top of the section. Small plant macrofossils in this material were exceptionally well preserved (see below). Two $^{14}C$ dates were obtained on the plant material, **14,120±90 yr BP** (Beta-52711) and **13,080±100 yr BP** (W-6397); both dates came from the same sample of material, so the reason for the 1,000-yr difference between the two dates is not clear at present; a willow twig submitted by Norton Miller (see below) for AMS dating should resolve the problem. Current directions of crossbeds and ripples and composition indicate that these beds are distal from a mapped post-stable-stage delta about 2 km to the northeast, built into Lake Hitchcock by the paleo Farmington River (Stone and others, 1992; Stone and Schafer, in prep.). This delta records lowering of the lake by about 5 m from the stable 82-ft (25-m) level; lowering of this amount produced a lake shoreline within a few hundred meters of the Matianuck site. It is probable that the post-stable delta represents lake-level lowering due to initial breaching of the Cromwell-Rocky Hill drift dam.

**A preliminary account of plant fossils from Matianuck Ave. site**
**by Norton G. Miller, Biological Survey, New York State Museum, Albany, NY 12230**

Lake-bottom sediments from the Matianuck Ave. site in Windsor CT contained abundant pollen and plant macrofossils. Two samples, mostly lacking large plant fossils and each limited to a single bed, one 5 cm below the other, were taken for pollen analysis from a clean face of a 20x20x20-cm block of sediment. The two samples were separated by four thin sandy beds. The samples were processed in the laboratory following standard procedures (HCl and HF digestion of inorganics, acetolysis, floatation in $ZnCL_2$). Both pollen residues contained large amounts of pollen of herbs, over 50% of a sum (percentage base) that included all terrestrial plants (trees, shrubs, and herbs; with pteridophytes, bryophytes, and unknowns and degraded pollen excluded from the sum).

The pollen spectra are unusual in having high percentages of grass (Gramineae) pollen (41%), a relatively modest representation of sedge pollen (14%, 16%), and the occurrence of pollen of many kinds of herbs and shrubs, including *Armeria* (0.4%, 0.6%); *Dryas* (0.3%, 1.4%); *Oxyria* (0.8%, 3%); *Saxifraga* (0.2%, 0.3%); *Thalictrum* (2%, 4%); undifferentiated Caryophyllaceae (4%, 4%), Compositae (2%, 3%; excluding *Artemisia* (0.4%, 0.2%)); Cruciferae (0.1%, 0.3%), Ericaceae (0.2%, 0.5%) and Rosaceae (0.1%, 0.1%); *Salix* (4%, 5%), and *Alnus* (including *A. crispa*). Pollen of aquatics was limited to single grains of *Hippuris* and *Nymphaea*.

Tree pollen percentages were 29% and 33% of the sum, with pine (*Pinus*) the predominant type (14%, 19%). Much of the pine pollen was of the jack pine (*P. banksiana*) type. Spruce (*Picea*) percentages were low (4%, 5%), and small amounts of fir (*Abies*) pollen were present (0.1%, 0.6%). Pollen from hemlock (*Tsuga*), beech (*Fagus*), sugar maple (*Acer saccharum*), ironwood (*Carpinus* and/or *Ostrya*), elm (*Ulmus*), ash (*Fraxinus*), basswood (*Tilia*), hickory (*Carya*), black walnut or butternut (*Juglans*), and aspen or poplar (*Populus*) occurred in trace amounts (generally less than 0.2%). Oak (*Quercus*) and birch (*Betula*) pollen was more frequent, 1-2% and 1% of the sum, respectively).

These pollen assemblages indicate an essentially treeless vegetation. The tree pollen present (boreal conifers, spruce, and jack pine) probably came from distant (i.e. not local) sources. Some of the other pollen in the sediment may have been deposited by wind, but most of it is probably of terrestrial detrital origin, deposited in Lake Hitchcock along with the plant detritus. Therefore, direct comparisons with the pollen spectra of the same age from small lakes are problematic. The terrestrial detrital origin of much of the pollen may explain why grass is so abundant, whereas it is poorly represented in the basal late-glacial silts and clays of small lakes basins in the New England-New York region.

Plant macrofossils were abundant in the sediment; these were disaggregated by swirling a sample in a dish of water until the individual leaves and other fossils were freed from one another. Inorganic sediment in suspension was decanted as required. In a second treatment, bulk sediment samples were washed free of small inorganics in nested 500-um- and 250-um-mesh sieves. The organic-rich residues were transferred to jars prior to examination under a dissecting microscope. To help free the fossils from adhering silt and clay, the samples were put in beakers, flooded with 5% $Na_2CO_2$ and heated to $50^0$ C on a hot plate for 3 hr (with periodic stirring). To help keep the fossils intact, vigorous agitation of the sample was avoided.

The plant macrofossil assemblages consisted of leaves, twigs, bud scales, fruits, and seeds of vascular plants and fragments of leafy moss plants. The assemblages have not been studied completely, but some of the species already identified include the dwarf willow, *Salix herbacea* L. (leaves of which are the most abundant plant macrofossil in the organic debris); alpine meadowrue (*Thalictrum alpinum* L.), *Sibbaldia procumbens* L., and *Oxyria digyna* L. (which are represented by fruits); *Armeria maritima* L. (represented by calyces); and the moss *Aulacomnium turgidum* (Wahlenb.) Schwaegr. Some of these macrofossils are illustrated in figure 7. In addition, fruits of sedges, *Potentilla* and Compositae and leaves of other species of willow and of blueberries (*Vaccinium*) are present, as are about 20 additional species of mosses.

The contemporary ranges of these species of seed plants and the moss are largely arctic-alpine, with southernmost occurrences in northeastern North America associated with the highest mountains in this region (e.g. the Adirondack High Peaks, (New York), the Presidential Range (White Mountains, NH), the Shickshock Mountains (Gaspe Peninsula, Quebec), and alpine areas of Newfoundland and Labrador). *Salix herbacea* and *Aulacomnium turgidum*, for example, now grow in the alpine tundra of the Adirondacks, White Mountains, Mt. Katahdin,
and in appropriate places northward, but *Thalictrum alpinum* occurs no farther south at present than in suitable habitats on the Gaspe Peninsula. Thus, these species do not presently share identical ranges but are similar in being arctic and alpine in distribution and occurring in areas that are treeless.

The pollen and plant macrofossils indicate a preforest stage in the late-glacial vegetation of New England. They are evidence of herb- and shrub-dominated plant communities on surfaces above the shore of Lake Hitchcock. The communities comprised species that today are common in subarctic and arctic Canada; they occur only sporadically in the northeastern United States and adjacent parts of Canada in edaphically specialized sites on cliffs that are microclimatically suited to arctic and alpine plants. The plant fossils at this site extend the record for such vegetation in the Connecticut River Valley 325 km south of the Columbia Bridge site, near Colebrook, NH (Miller and Thompson, 1979). At Columbia Bridge, a boreal plant assemblage is dated at 11.4-11.5 ka. This assemblage is present 2,500 years earlier at the Windsor, CT, site.

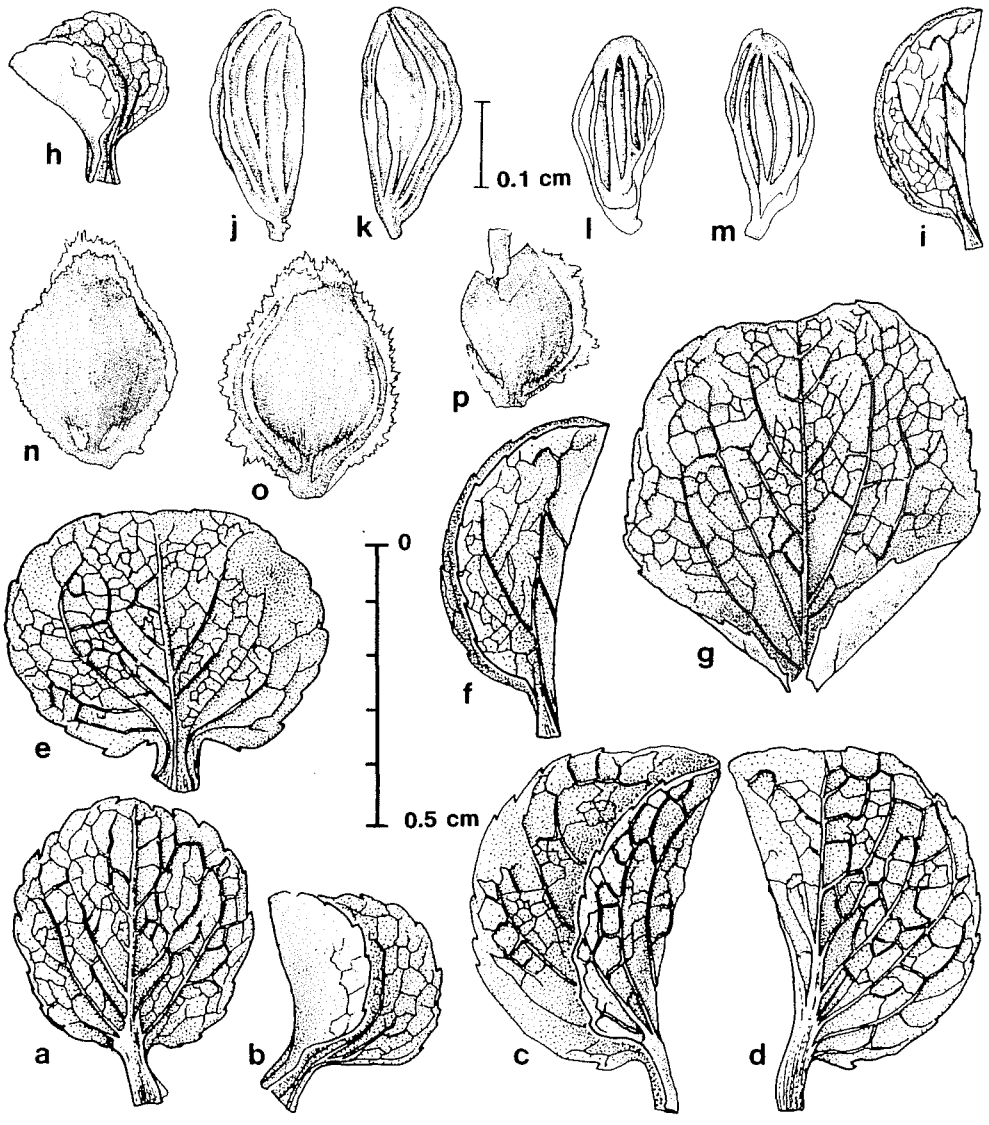


**Figure 7. Selected plant fossils from Matianuck Ave. site, Windsor, CT. a-i**, leaves of *Salix herbacea*, note that some of the leaves are folded in half, which is characteristic of fossils of this species; **j-m**, achenes (fruits) of *Thalictrum alpinum*, **j** and **k**, **l** and **m**, show both sides of the same fruit; **n-p**, achenes (fruits) of *Oxyria digna*, marginal wing eroded to different degrees. (**a-i**, lower scale; **j-p**, upper scale).

## Conditions at and following lake drainage

At the end of its existence Lake Hitchcock was very shallow in the southern part of the Valley; the lake bottom was only 10-12 m below lake level. North of the Holyoke Range in Massachusetts, the lake bottom was much deeper, 30-50 m below stable lake level (see fig. 4 in Koteff and others, 1988). When the the relatively narrow, easily erodable was breached, water levels

must have dropped relatively quickly. Once the lake level had lowered by 10-12 m, however, the lake-bottom surface was exposed, the lake in the south was gone, and the incipient Connecticut River began the more difficult task of entrenching the lake bed over a very long distance. Extensive fluvial terrace deposits at several levels (the highest is seen at Stop 5) record this entrenchment of the lake-bottom surface. At this time the lakebed in the southern part of the basin was acting as the dam for a lower-level remnant-lake that occupied the northern part of the basin. Water in this remnant lake would still have been 20-40 m deep. The Ridge and Larsen (1990) estimate of lake drainage at 12.4 ka at the Canoe Brook site in southern Vermont most likely applies to this northern remnant of Lake Hitchcock.

Evidence for an open landscape and strong northwesterly winds after the lake drained is preserved on the lakebed in the form of extensive longitudinal and transverse dunes especially on the east side of the Valley. Dunes up to 9-m high can be seen along the route between Stop 5 and 6 (fig. 11); many more are present on surfaces between Stops 5 and 7. The dune exposed at Stop 7 is on a high lake-bottom surface at 175-185 ft in altitude in front of a delta graded to a high level of glacial Lake Hitchcock (20 ft (6 m) above stable level); the stable shoreline at that locality was at 170 ft. The north to northeasterly wind directions indicated by the dune morphology and internal structure (see discussion by R.M. Thorson, Stop 7) record an early wind direction after the lake had dropped to stable level, and while the ice margin was still in southern Massachusetts. The upper section of this dune contained detrital pieces of charcoal which yielded a $^{14}C$ date of **11,485±115 yr BP** (AA-7154), (L. McWeeney, pers. comm.); thus, vigorous eolian activity continued at least until this time.

The incision of the lakebed by the early Connectiuct River down to the level of the modern floodplain was probably accomplished by about 8 ka. We have recently obtained vibracores from two low-terrace areas in Massachusetts, 2-3 m above the modern Connecticut River floodplain. Basal organics in a core from the Huntington Road site (Hadley, MA) yielded a date of **9370±100 yr BP** (Beta-52257) and 35 cm higher in the core, **8780±80 yr BP** (Beta-52256); basal orgainics in the Sanderson core near Stop 8 (Whately, MA) yielded a date of **8530±110 yr BP** (Beta-52258) (L. McWeeney, pers. comm.).

## CONCLUSIONS

Interpretations of the conditions under which the last ice sheet receded during the late Pleistocene vary with the evidence used, location of the study and the investigator. In southern New England the glacial sediment record indicates fairly rapid recession following the maximum extent of the Laurentide ice at about 21 ka (Sirkin, 1982; Stone and Borns, 1986). Deglaciation was characterized by systematic northward retreat of an active-ice margin fringed with a zone of stagnant ice and by morphosequence deposition (Koteff and Pessl, 1981; Stone and others, 1992; Stone and Schafer, in prep). In southern New England, ice retreat was dominated by meltwater (Gustavson and Boothroyd, 1987). By 12 ka, the ice margin was in southern Canada and marine waters occupied the St. Lawrence lowland (Dyke and Prest, 1987). The calculated retreat rate for southern New England is 50-75 m/yr, and the conventional wisdom of glacial geologists has been that rapid meltwater-dominated deglaciation occurred in a rapidly warming climatic regime. However, if we examine the paleobotanical literature for southern New England, we find that pollen records and plant macrofossils indicate the presence of cold-climate, tundra vegetation for several thousand years following deglaciation (see recent regional summaries by Gaudreau and Webb (1985) and Jacobson and others (1987)). Evidence for Connecticut comes from records at Rogers Lake (Davis and Deevey, 1964; Davis, 1965; 1969), Totoket Bog and Durham Meadows (Leopold, 1955; 1956), and Cedar Swamp, (Thorson and Webb, 1991). These records all indicate that between about 15 ka and 12.5 ka a treeless plant assemblage was present in southern New England characterized by sedges, shrubs and grasses (the "herb zone"). This assemblage is indicative of arctic to subarctic conditions.

Northward retreat of the ice margin from the Long Island terminal moraines through southern New England took place early (19-14 ka), before the great bulk of the Laurentide Ice Sheet began to dissipate. Retreat through southern New England was probably induced by changes in dynamic equilibrium within the ice sheet rather that climatic amelioration. A periglacial climate which supported permafrost, at least in some areas favorable to its development, persisted for several thousand years after the ice had left the area.

## FIELDTRIP STOPS

**STOP 1: HAIN BROTHERS SAND AND GRAVEL PIT**, town of Windham, CT, Willimantic quadrangle. From Rt. 32 and Rt. 203 junction, proceed 0.75 mi (1.2 km) northeast on Rt. 203. Entrance to pit is left off Rt. 203, 0.75 mi northeast of Rt. 32 intersection. Pit is areally extensive; we will examine sections at nine exposures lettered **a-i** in fig. 7.

***We will spend a couple of hours in this pit, and as we examine and perhaps argue about these intriguing "wedge structures", let us remember fondly Phil Schafer who spent much time here over the years, who first interpreted them as ice-wedge casts, and from whom we gained a great deal of our knowledge about the details of these structures.***

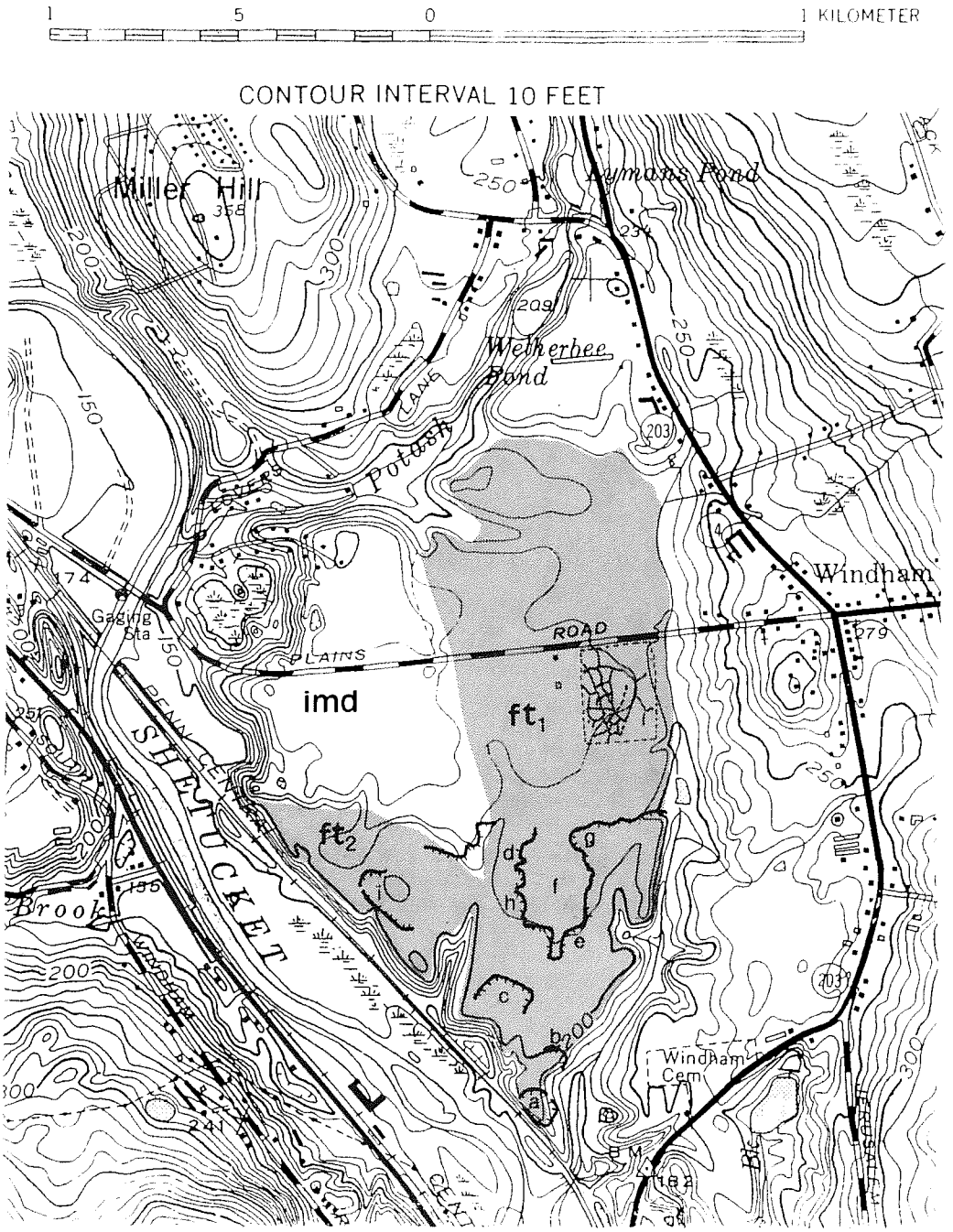


**Figure 8. Topography of Hain Pit area.** Pit excavation scarps (1990) shown by hachured lines. Polygonal ground pattern sketched from airphoto (fig.4). Shaded areas are fluvial terraces ($ft_1$, $ft_2$); these are inset into an ice-marginal delta (**imd**). Sections **a** through **i** described below.

The glacial geology in the vicinity of the Hain pit was mapped at 1:24,000-scale by Clebnik, (1980), who also recognized the occurrence of numerous wedge-shaped structures in the Hain pit. More detailed analysis of the area during investigation of the wedge structures, has shown that the area included in figure 8 consists of three different episodes of meltwater deposition. An ice-marginal delta (**imd**) was deposited into a small drift- dammed lake in the Shetucket valley (one of a series of such lakes in that valley). This delta has an ice-marginal slope on its NW side and a surface plain that slopes southeast from 255 ft altitude near the ice-margin position to 235 ft near its southern edge where it is cut by a 20-ft scarp; its eastern side is cut by a 10-15-ft scarp. Both scarps were incised during the succceeding episodes of meltwater erosion and deposition. The 245-250-ft surface SE of Wetherbee Pond is also part of the ice-marginal deltaic sequence; there are no exposures in the delta surface, but it is estimated that the delta was built into a lake with a water plane of about 230 ft altitude.

The second episode of meltwater deposition was initiated when lake levels in the Shetucket valley dropped (probably due to erosion of the drift dams) by at least 25 ft. This resulted in incision and reworking of the surface of the ice-marginal delta by meltwater that flowed down the Potash Brook valley east of Miller Hill from an ice margin position farther to the north. The surface of this fluvial terrace deposit (**$ft_1$**) slopes from 225 ft at the north end, through the patterned ground field (fig. 4), and the Hain pit area to about 215 ft where it is cut by a 10-ft scarp visible only on older (pre-gravel pit) airphotos. The deposits of this fluvial terrace are exposed in sections **d** through **h** in the Hain pit; they comprise ~3 m of horizontally layered and cross-bedded sand and pebble gravel. This material is the main source of gravel in the Hain pit, and most of the wedge structures deform these beds. Near the lower part of these sections, a disconformable contact between the fluvial sands and gravels and the underlying delta foreset and bottomset beds of the deltaic sequence (**imd**) is exposed in some places.

The third episode of meltwater deposition took place slightly later, perhaps at first contemporaneously with the **$ft_1$** terracing. Distal meltwater in the Shetucket valley terraced the southern edge of the ice-marginal deltaic deposits and constructed a 205-ft terrace surface (**$ft_2$**). Fluvial sand and pebble gravel beds exposed in section **i** are as much as 6 m thick.

**Section a**: reveals approximately 12 m of sandy delta forset beds; the altitude of the pre-excavation stream-terrace surface here was 195 ft; only a small remnant of the original surface remains; the floor of the excavation is now at an altitude of about 150 ft.

**Section b**: reveals sandy and silty lacustrine beds higher in section than section **a;** quite spectacular penecontemporaneous water-escape structures can be seen near the top of this section, which is not the original surface. Several meters of section have been stripped by excavation.

**Section c**: reveals about 5 m of sandy and silty lacustrine beds, still higher in section than previous section **b**; well-developed sets of climbing ripples can be seen here, and also more water-escape/load deformation in the form of flame, and ball-and-pillow structures. Periodically exposed near the top of this section is a 0.3 to 1.0-m-thick and 15-m-long lense of varved silt and clay in 10 couplets (each about 5-10 cm thick). These fine-grained beds result in a perched water table on the floor of much of the pit area to the north. About 3 m of fluvial sand and gravel beds have been removed from the top of this section; the original terrace surface was at 205 ft in altitude.

Sections **d** through **i** are cut into the original ground surface; wedge structures are exposed (or formerly were) in all of these sections. All sections except **f** have been stripped of 0.5-1.0 m of upper soil horizon prior to sand and gravel extraction.

**Section d**: fluvial terrace deposits ($ft_1$) are exposed in this area. At times the disconformity between the fluvial beds and the underlying deltaic/lacustrine beds is exposed near the base of this section. Lacustrine material underlies the pit floor; puddles of water on the pit floor indicate the presence of a perched water table above the underlying fine-grained lacustrine beds. Several wedges which deform the fluvial beds and penetrate into the lacustrine section, have been visible along this N-S trending face during the past 2 years, as the face has moved eastward. The most prominent wedge, near the center of this face, strikes nearly E-W and is exposed in section **g**, as well; its linear ground trace across the topsoil-stripped surface between **d** and **g** can be seen on the 1990 airphotos.

**Section e**: exposes about 3 m of fluvial section ($ft_1$), beds are generally finer-grained than in section **d**, consisting mostly of fine to medium sand, with only minor pebble gravel. Three wedges, all of which exhibit preserved upturned beds and pressure-effects in the wall strata are exposed in the pit face (see fig 3. and discussion in text). One wedge here strikes N40°W and intersects with its neighbor, which strikes N60°E, approximately 6 m in front of the pit face. This intersection can be seen by scraping the pit floor and following the deeply oxidized sand trace of the narrow "at depth" wedge fillings.

**Section f**: is a ~1.5-m-deep backhoe excavation in the only remaining non-pitted, non-topsoil-stripped surface in the Hain pit; the trench is cut perpendicular to a wedge fracture. Here we can see an intact modern soil horizon developed in the eolian mantle. "Pebbles" within the eolian fine sand have been incorporated from the underlying fluvial gravel and sand beds by cryoturbation. Note also that the "soil" is not deformed by the wedge structure. The following is a pedon description outside of the wedge structure by Dr. Harvey Luce, University of Connecticut, Storrs, CT:

Phase: Agawam gravelly fine sandy loam, 0-3% slope.

**Ap**- 0 to 23 cm; very dark brown (7.5YR2/2) gravelly fine sandy loam; weak fine subangular blocky structure parting to weak fine granular structure; very friable; abundant roots; abrupt smooth boundary.

**Bw1**- 23 to 37 cm; dark yellowish brown (10YR4/6) gravelly fine sandy loam; weak fine subangular blocky structure; very friable; many roots; gradual smooth boundary.

**Bw2**- 37 to 51 cm; yellowish brown (10YR5/6) fine sandy loam; weak medium subangular blocky structure; very friable; common roots; abrupt smooth boundary.

**2C**- 51 to 101 cm; brownish yellow (10YR6/6) gravelly sand; single grain; loose; very few roots.

**Section g**: Wedge intersections exposed in this face in the past have displayed extremely complex deformation structure. Presently exposed is a deep E-W striking wedge into which the eolian plug material penetrates as much as 3 m. As a result of the greater depth at which the eolian material lies in many of the wedges but particularly in this one, it is gray in color. This is because the deeper eolian material is not pervasively oxidized by B-soil horizon processes as it is in the shallower eolian mantle material.

**Section h**: This section has historically exposed numerous soil-involution (cryoturbation) structures, some of which are developed in the top of the wedge structures. The involutions appear to be of type 2 and 4 of Vandenberghe (1988), formed by loading due to cryohydrostatic pressures. These kinds of involutions are generated during the progressive degradation of permafrost, when a lot of water is present but an underlying impermeable horizon (the permafrost) is still present. As discussed by Van Vliet-Lanoe (1988), the presense of fine-grained eolian cover on coarser-grained fluvial materials, produces a positive frost susceptibility gradient. With progressive lowering of the water table associated with the downward migration of permafrost table, pot-shaped involutions form in ice-wedge cracks, especially at their intersections which are the lowest and wettest points of the microtopography.

**Section i**: For the past several years, this section has revealed about 6 m of fluvial pebble gravel and sand ($ft_2$); no wedges have been seen here during this time, however their presence here is indicated by Thorson and others (1986). At present this excavation has expanded northward by about 150 m from the 1990 scarp shown on figure 8. Excavation has begun to intersect the south end of unit **imd**.

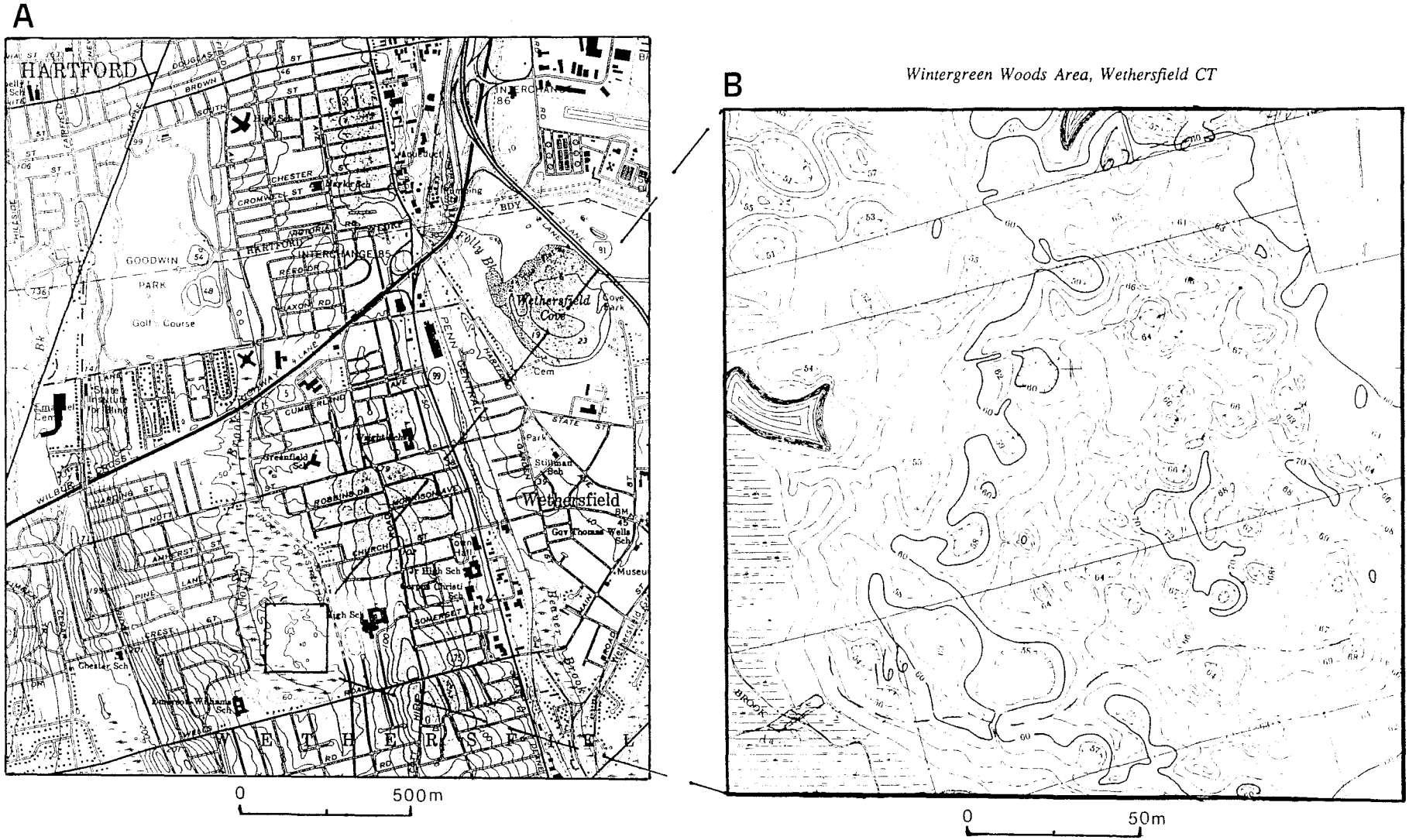


**Figure 9. Topography of Wintergreen Woods Area.** **A)** Section of Hartford South 1:24,000-scale quadrangle map with 10-ft contour interval (reduced). **B)** Section of Metropolitan District Commission (MDC) 1:2400-scale map 147 with 2-ft contour interval (reduced).

**STOP 2. WINTERGREEN WOODS park**, town of Wethersfield, CT, Hartford South Quadrangle. From its junction with Silas Deane Highway (Rt. 99), proceed 0.3 mi (0.5 km) west on Wells Road (Rt. 175). Turn right on Folly Brook Drive. Entrance to park is at the end of Folly Brook Drive, 0.15 mi (0.25 km) north of Wells Road.

This is the best known locality to see the landforms which we have interpreted as pingo scars; park nature trails wind around the rims of the many depressions that occur on this surface. The lake-bottom surface here is part of a finger of Lake Hitchcock that existed in a low area between two till/bedrock hills. Although it is not at all apparent on the quadrangle map (fig. 8A), the MDC map (fig. 8B) with a 2-ft contour interval shows that the entire surface is covered with shallow closed depressions and depressions with breached rims marked by C-shaped contours. As is generally the case, the water table is at relatively shallow depth beneath these lake-bottom surfaces, and its seasonal fluctuations result in alternately ponded (or marshy) and dry conditions in the lower part of the depressions. Many of them are in fact "vernal pools" which are homes to numerous amphibious creatures (some of which are endangered species). The pingo-scar depressions in this area are relatively small compared to other areas, ranging in diameter from 8-25 meters. In this area the features are present on surfaces between 50 and 70 ft altitude which is as high as the lake-bottom surface gets here; the stable level shoreline was at an altitude of 82 ft. The features are conspicuously absent on the lower surfaces incised into the lake-bottom by the north-draining Folly Brook. This general characteristic of their distribution demonstrates their early-postlake time of formation.

**STOP 3. I-91 CUT NORTH OF HARTFORD**, town of Windsor, CT, Hartford North Quadrangle. From Windsor Ave. (Rt. 159) at Wilson, proceed west on Matianuck Ave; 0.4 mi (0.6 km) north of Rt. I-91 underpass, turn right on Oak St. (cul de sac). Roadcut was formerly behind highway fence; neighboring pingo scar is in yard on right.

A 4-m-deep, 165-m-long roadcut along the west side of southbound Rt. I-91 between exits 34 and 35 was exposed during the summer of 1989. The cut is now graded over; we include it as a STOP because of its importance to the interpretation of the rimmed depressions as pingo scars; this road cut remains the only cross-sectional exposure we have seen. The section (fig. 5A) revealed upper Lake Hitchcock bottom sediments, about 3 m of rhythmically bedded vf-f sand and silt containing pervasive water-escape structures mostly in the upper 1-2 m overlying poorly exposed varved silt and clay. The section cut through a rimmed depression (fig. 5B) developed in the lake-bottom sediments. Topographically, the feature was a circular, 40-m-diameter, 3.5-m-deep wetland depression (fig. 5B) similar in morphology to many other features in the area. The cross section clearly showed that the surface depression was underlain by deformed lake sediments. The parallel-bedded lacustrine sediments could be traced toward the depression where they were upwarped beneath the surface rim, then intensely collapsed downward beneath the central part of the depression. The structural depression created by the collapsed beds was partially filled with a bowl-shaped body of eolian-supplied fine to medium sand overlain by a body of peat and gyttja (see previous discussion of internal structure and $^{14}C$ dates above). A feature we interpret as a small ice-wedge cast cut lacustrine sediments at the south end of the exposure (see fig. 5A).

Connecticut Department of Transportation test borings indicate that 15-18 m of varved silt and clay underlie this section. The presence of deformed lake deposits below the surface depression exposed in this roadcut eliminates the alternative hypothesis that these features are eolian deflation hollows.

**STOP 4. MATIANUCK AVE. ROADCUT**, town of Windsor, CT, Hartford North Quadrangle. From Oak St., proceed 0.4 mi (0.6 km) north on Matianuck Ave. to road construction for new entrance ramp for relocated Rt. 291 (Bissell Bridge connector). East-west trending roadcut passes beneath Matianuck Ave. Section on east side was still visible in August 1992.

This roadcut is slowly being graded over and may not be viewable in the near future. The lake-bottom section exposed here is similar to that formerly exposed at the I-91 cut, Stop 3. West of Matianuck Ave., the varved section was better exposed than anywhere at the I-91 cut. A composite sketch of sections east and west of Matianuck Ave. as seen in Oct. 1991 and March 1992, respectively, is shown in figure 10.

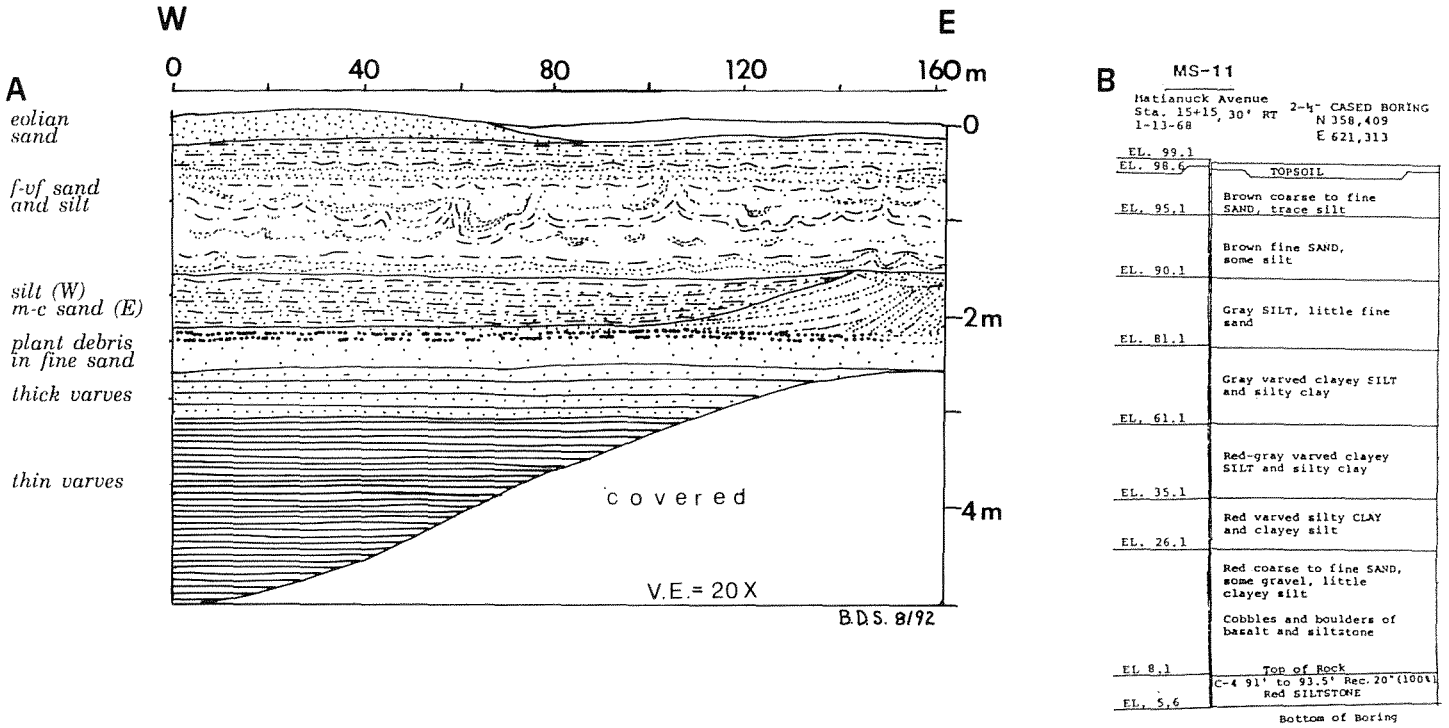


**Figure 10. A)** Composite sketch of Matianuck Avenue roadcut section **B)** Bridge test boring log for Matianuck site.

West of Matianuck Ave, the deeper section exposed 2.5 m of varved silt and clay. Two meters of thin gray varves averaging 1-cm in thickness were overlain by 0.5 m of thicker reddish gray varves averaging 2-3 cm in thickness. Test borings for bridge construction (fig. 10B) indicate the presence of approximately 20 m of varved silt and clay at depth below this section. The varved section recorded in boring logs shows typically red varves in the lower part, alternating red and gray varves in the middle of the section and gray varves in upper part. This color/mineralogic change within the varve section reflects the early presence of ice in the basin supplying local "red" (Mesozoic rock-derived) material, then changing with increasing distance of ice source to "gray" (crystalline rock-derived) material supplied by streams flowing from the highlands into the lake.

Approximately 2.5 m of fine sand and silt beds containing pervasive water-escape structures overlies the varved section. At the eastern end of the roadcut, a lense of medium to coarse sand in a 1.0-m-thick set of crossbeds (containing abundant large muscovite flakes) prograded westward intertonguing with thin (0.5-1.0 cm-thick) parallel-laminated fine sand beds. Two of these beds consisted predominantly of detrital plant debris. The organic material yielded a $^{14}C$ date of **14,100±110 yr BP**. The well-preserved state of plant macrofossils indicates that the plant debris had not been transported far. Local and regional stratigraphy indicates that the sandy beds containing the plant material were deposited in association with lowering of the lake level by breaching of the drift dam (see previous discussion).

**STOP 5. EAST HARTFORD SAND AND GRAVEL PIT and KELSEY-FERGUSON CLAY PIT**; town of South Windsor, CT, Manchester quadrangle. From junction of Rt. 291 (Bissell Bridge) with Rt. 5, proceed 2.5 mi (4 km) north on Rt. 5. Turn right at light onto Strong Rd., then immediate left. Turn right past the K-F Brick Co. buildings onto dirt road. Two exposures are present; the sand pit exposes stream terrace deposits and a deeper clay pit exposes lake-bottom sediments.

As seen in figure 11, stream terrace deposits (**Qst**), 3-4 m thick, overlie lake-bottom deposits (**Ql**) which are as much as 53-m thick in the vicinity of this Stop. The stream terrace deposits are pebbly, crossbedded sands laid down soon after the draining of Lake Hitchcock. The terrace surface is at 75-85 ft altitude here and grades to a 50-ft terrace surface cutting through the drift dam deposits to the south. Its surface has been modified by eolian activity. The terrace deposits are inset only slightly into the underlying lake-bottom surface which lies at 65-70 ft altitude beneath it. Lake-bottom surfaces just to the east (Stop 6) are at 85 ft altitude. Stable lake level here was at 115 ft altitude; after lowering of only 40 ft (12 m), the lake-bottom was exposed and terracing began.

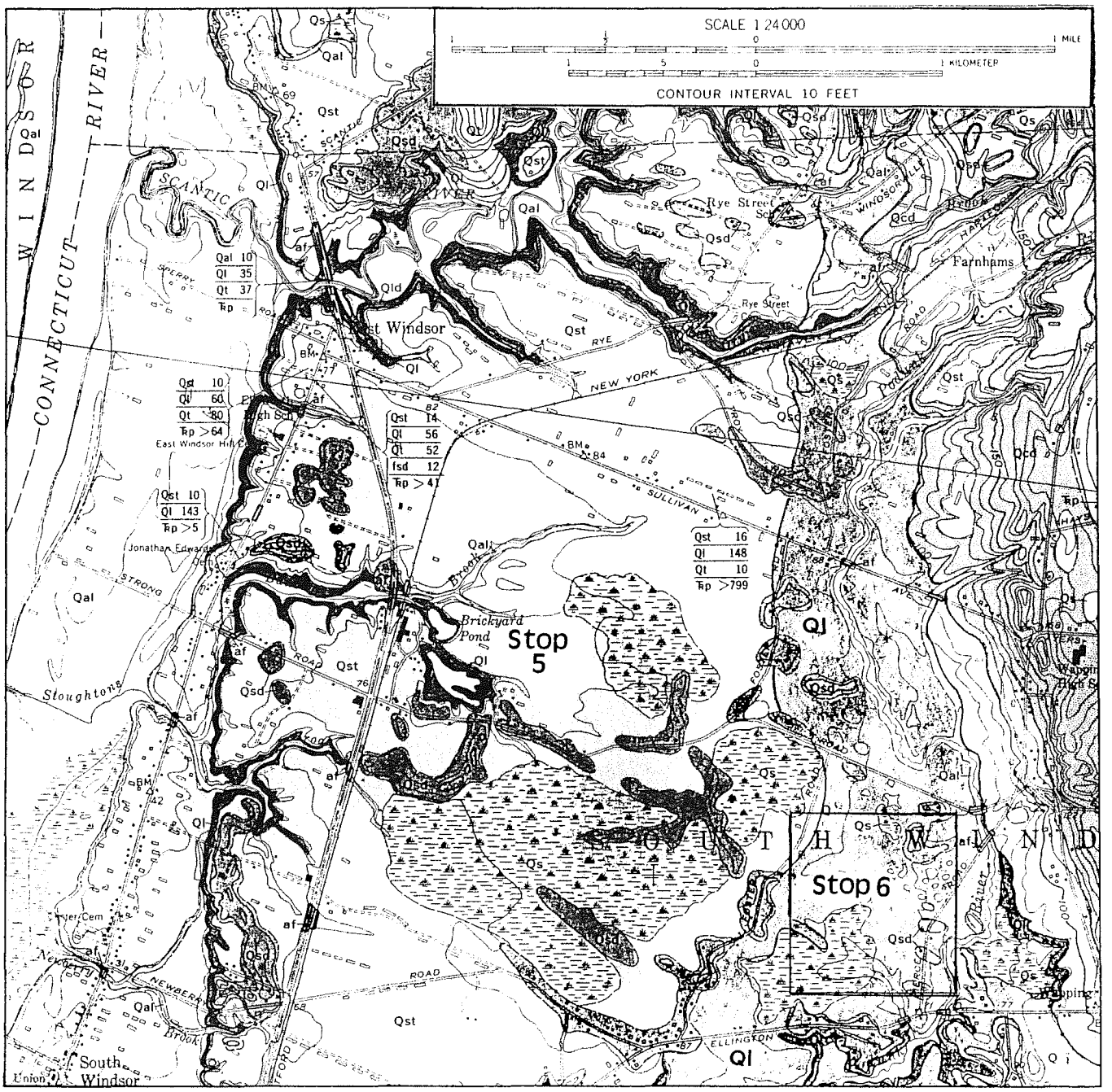


**Figure 11. Topography/Geology in the vicinity of Stops 5 and 6.** Modified from the geologic map of the Manchester quadrangle (Colton, 1965). Note that the contact between Qst and Ql has been changed. Qst-stream terrace deposits, Ql-lake-bottom deposits of glacial Lake Hitchcock, Qsd-eolian sand dune deposits, Qs-swamp deposits, Qal-floodplain alluvium. Area of Figure 12 is outlined.

About 4 m of varved sediments are periodically well exposed in the clay pit. This location is 2.5 km from the lake margin and is far removed from major deltaic sources of sediment. The lake at the time of deposition was 17-21 m deep. Sediments are rhythmically bedded dark clays and light colored multiple-graded silts. However, there is enough inconsistency in the bedding (i.e. thin clay bands in silt and thin silt layers within the clay) to present problems in interpreting these rhythmites as varves. It is difficult to determine which silt laminae represent the commencement of a yearly deposit and which are the result of a chance introduction of silt into the lake during clay deposition. It is likely that sediments were transported to the site by overflows and/or interflows.

**EN ROUTE TO STOP 6**: Longitudinal and transverse dunes up to 9 m high built by northwesterly winds occur on the terrace surface. Strong Road crosses several dunes and Foster Road follows the crest of one. The wetland areas (Qs) on the terrace surface are relatively large; these are deflation areas associated with the dunes. Near its junction with Foster Road, Strong Road crosses onto a nonterraced lake-bottom surface (hand shaded and label Ql on figure 11) that lies at 85-95 ft altitude. This surface has many small, shallow closed depression/wetlands and breached, "C"-shaped depressions (visible on the 2-ft contour interval map in figure 12), but not visible on the 10-ft contour interval map of figure 11. Small sand dunes may be present on this surface as well, but it is difficult to distinguish them from the pingo-scar rims.


**Figure 12. Topography in the vicinity of Stop 6.** Section of MDC map 467 showing detailed, large-scale topography of Wapping area pingo scars. See figure 11 for regional location. Path along power lines between Pierce Road and Foster Road provides access to view some of the best-developed of these features.

**STOP 6: WAPPING PINGO SCARS:** town of South Windsor (village of Wapping), CT, Manchester quadrangle. Proceed east on Strong Road, turn right on Foster Road. Proceed 0.25 mi (0.4 km) to Seventh Day Adventist Church on left. Trail begins at small outbuilding at edge of parking lot.

The trail between the church parking lot and Pierce Road to the east (see figure 12) follows a line of well-developed pingo scars. These features are larger on average than the ones at Stop 2. These are generally in the 30- 80-m diameter range. The ones to the east along the power line are deeper and contain perenially ponded water (they show up as small ponds on the 1:24K quadrangle map, fig. 11). We have obtained a number of vibracores from several of the depression nearest to the church parking lot; cores and GPR lines from pingo-scar **FLA** (fig. 12), are described in Oakes (1992).

**STOP 7: ENFIELD DUNE**: town of Enfield, CT, Springfield South quadrangle. Procced east on Rt. 220 (Elm St.) which becomes Shaker Road. Turn left on Washington Road; proceed 1.6 mi (2 km) and turn right on Brainard Road; dune lies along south side of Brainard Road; exposure is behind apartment complex now occupying a former sand pit where internal structure was once well-exposed.

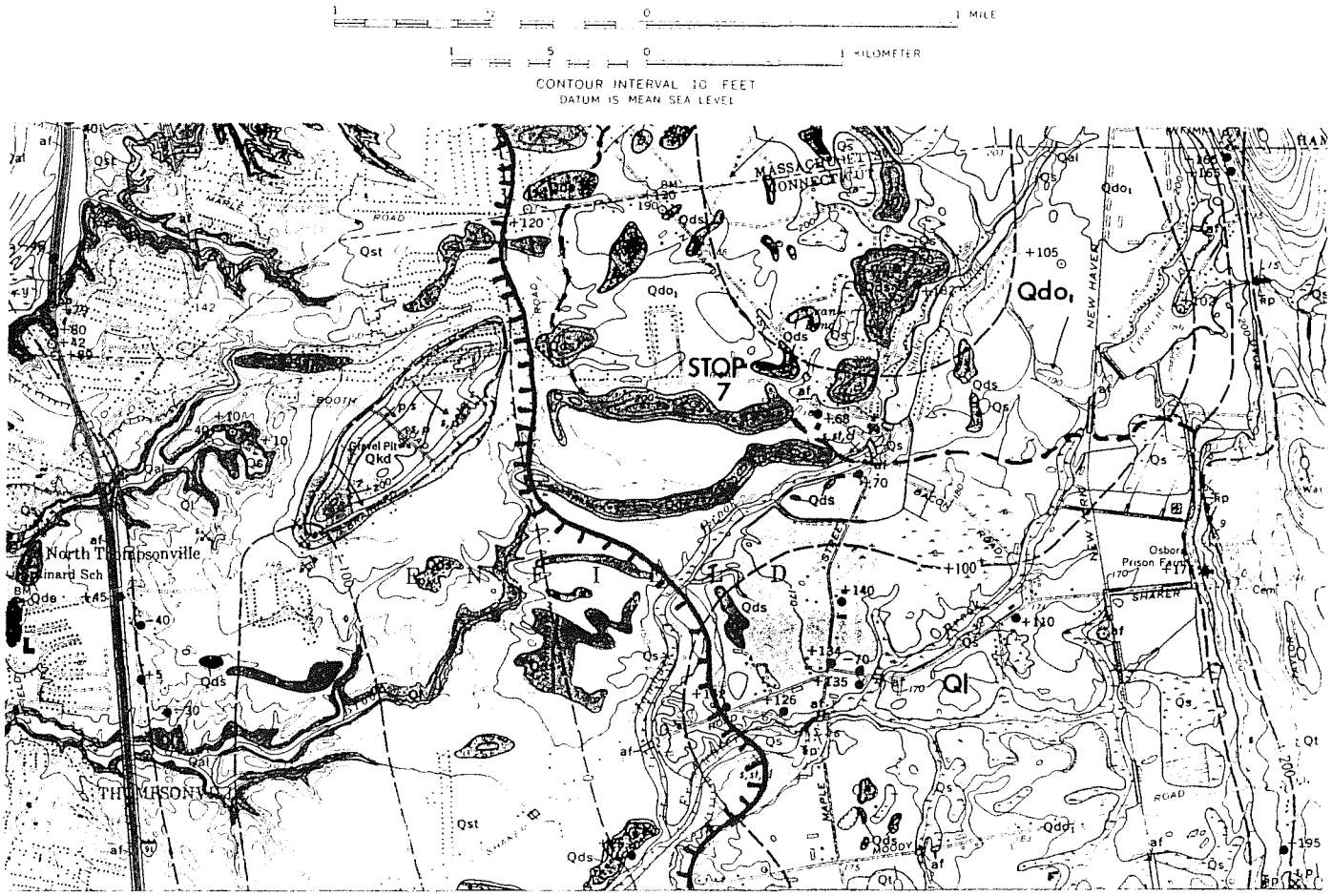


**Figure 13. Topography/Geology in the vicinity of Stop 7.** Modified from the geologic map of the Springfield South quadrangle (Hartshorn and Koteff, 1967). Lake Hitchcock stable-level shoreline is shown by heavy ticked line. Delta labeled $Qdo_1$ records a 190-ft lake-level, 20 ft (6.5 m) higher than stable level. The 175-185-ft surface to the south of the delta is a lake-bottom surface (Ql) associated with the higher lake level. The east-west trending transverse dunes lie on this surface; the high lake-bottom surface SE of the dunes displays many pingo scars (not visible at the scale of this map).

**Description by Robert M. Thorson, University of Connecticut, Storrs, CT**:

The "Enfield dune" is a complex elongate stationary dune that extends for nearly 1 km in an east-west alignment. The external form of the dune is irregular, but in its central section, the south-facing slope is consistently steeper than the northern one. This asymmetrical topographic profile suggests that it is a transverse dune, formed by northerly winds.

The internal structure of the dune was documented by Schile (1991). The bulk of the dune consists of parallel, inversely graded laminae that dip gently to the north-northeast, and which are interpreted as subcritical translatent strata formed by southeasterly migrating ripples on the northerly stoss side of the dune. The southerly part of the dune is dominated by grainfall and grainflow strata in steeply dipping beds, which formed on the lee side of the dune below its crest. Secondary structures are generally limited to burrow casts and root mottles near the present dune surface.

The so-called eolian mantle caps the dune with 1-2 m of silty unbedded sand, the bulk texture of which is similar to the underlying well-bedded material. Lying conformably below the eolian mantle and unconformably above the stratified sand is a complex transition zone ranging from a few centimeters to a meter thick. The base of the transition zone typically exhibits one or more horizons in which coarse sand and granules are intermixed with the sand. The sand above the granule concentration(s) is weakly and discontinuously bedded, but rarely contains shallow trough cross-stratified lenses of coarser sand interpreted as the fill of surface rills. At the crest of the dune, the transition zone contains an incipient soil (expressed as an oxidized, mottled, and bioturbated horizon) that is unconformably overlain by crosscutting beds of eolian sand 5-20 cm thick with trough-shaped basal contacts. One of these beds contained fragments of detrital charcoal collected by Schile. This charcoal was indentified by Lucinda McWeeney as *Pinus sp.* and yielded a date of **11,485±115 yr BP** (AA-7154).

We interpret the stratigraphy of the Enfield dune to indicate the progressive growth of a stationary transverse dune. This took place in front of a delta on a high lake-bottom surface after lowering of the lake to stable level. The north-northeasterly wind regime is consistent with eolian erosional surfaces reviewed by Schafer and Hartshorn (1965), and requires the continued presence of a strong anticyclonic circulation set up by the Laurentide Ice Sheet. The stratigraphy, sediment texture and sedimentary structures of the transition zone are consistent with a gradual stabilization of the area by vegetation. A discontinuous cover of plants in the interdunal environment probably restricted the supply of sand prior to a reduction in wind intensity, causing erosion of the dune surface and a lag concentrate, which in turn, would have led to edaphic conditions more favorable for plant growth on the dune. The remainder of the transition zone suggests the continued saltation of sand over a progressively vegetated surface. Local erosion and re-activation of sand accumulation, possibly by winds with a more westerly component, is suggested by the dated horizon.

**STOP 8: SANDERSON PIT**: town of Whately, Mass., Mt. Toby Quadrangle. Proceed north into Massachusetts on Rt. I-91. Pass through the Holyoke Range. Exit Rt. I-91 at interchange 24; turn right at end of ramp; proceed 0.1 mi (0.16 km), turn right (east) on Rt.116, proceed 0.15 mi (0.24 km), turn right on Long Plain Road; proceed 0.4 mi (0.65 km) to Sanderson farm nursery. Pit is down farm road on left at edge of terrace scarp.

A small pit cut into the upper lake-bottom section near its edge bounded by the Connecticut-River-incised scarps is the focus of this stop. The cut reveals 3-4 m of rhythmically bedded fine sand and silt, similar in sedimentary structure to the upper lake sections at Stops 3 and 4 in the southern part of the basin. Well logs indicate the presence of varves below the exposed section; the exposed fine-sand and silt beds are deformed by pervasive water-escape convolutions. We will use this location to discuss the origin of these water-escape structures, often seen near the top of the lacustrine section. Were these structures generated by syn-depositional loading?; are they cryoturbations generated by intense freezing or melting of permafrost?; or, were they generated by liquifaction induced by earthquakes associated with glacioisostatic rebound and lake drainage?

The lake-bottom surface (Qhl, figure 14) here has not been terraced; it lies at 185 ft altitude which is 120 ft (37 m) below the stable lake level recorded by the nearby Sunderland delta (topset/foreset contact at 295 ft). Stable Lake Hitchcock was much deeper here than it

was in the south of the Holyoke Range. When the drift dam at the southern end of the lake was breached, waters in the south lowered only 10-12 m before intersecting the lake bottom. North of the Holyoke Range, a remnant lake, 20-m deep remained after the drift dam was breached. This slowly lowering remnant lake may have lasted for a significant period of time as the outlet stream incised the 80-km length of the lake-bottom surface to the south.

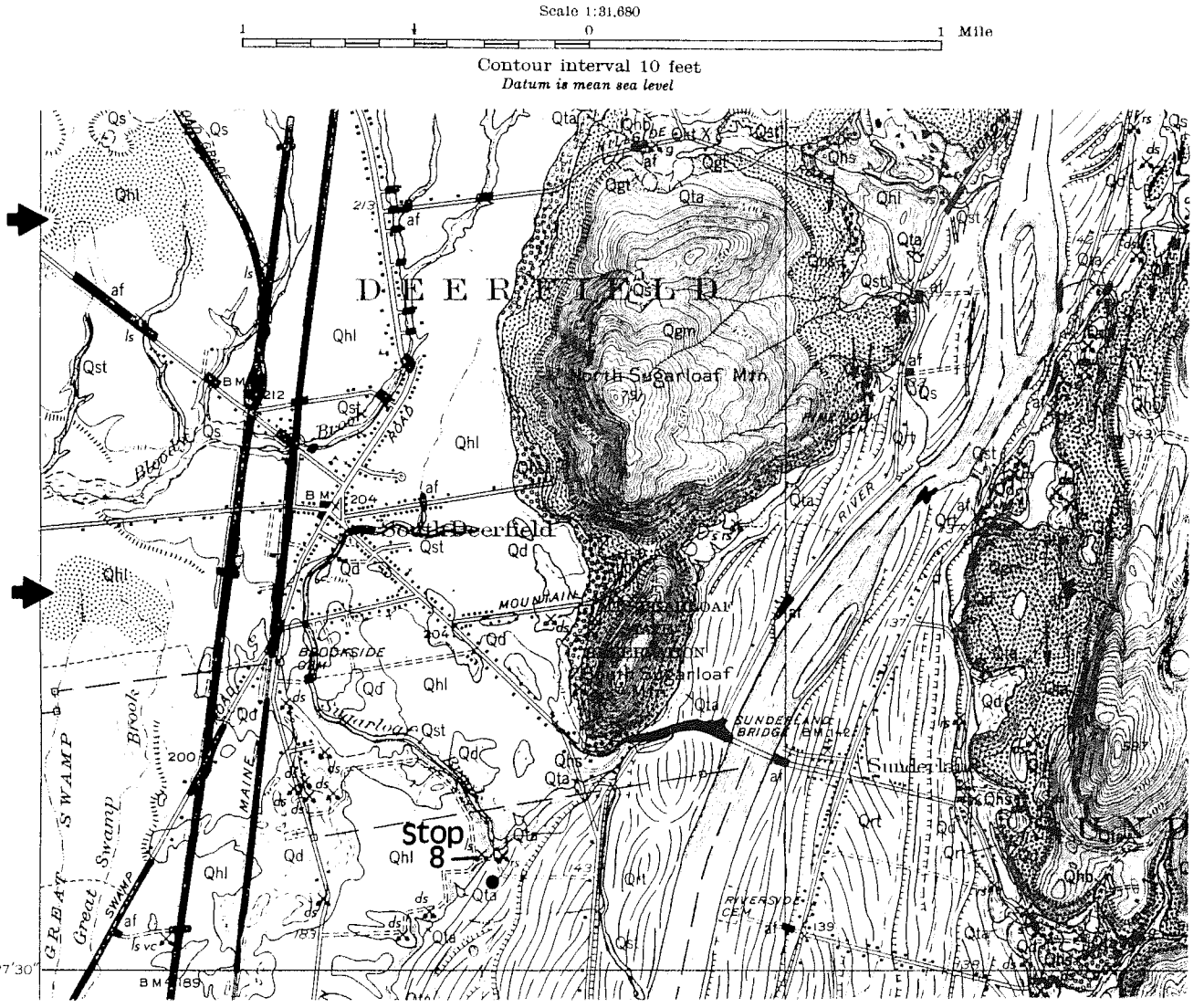


**Figure 14. Topography/Geology in the vicinity of Stop 8.** Section of the surficial geologic map of the Mt. Toby quadrangle, Massachusetts (Jahns, 1951). Arrows point to stippled areas mapped by Jahns as "kettled lake-bottom deposits"; these areas contain the small depressions we believe to be pingo scars. Dot is location of $^{14}C$ date on Connecticut River terrace.

The lake-bottom surface here also was subjected to intense eolian activity once the lake had drained. Dunes up to 6 m high built by northswesterly winds are common on this surface. Present also on this surface are the small rimmed depressions we believe to be pingo scars. These occur, in the area of figure 14 (see arrows), where Jahns (1951) shows a stippled pattern on unit Qhl to indicate "areas of kettled lake-bottom deposits".

A vibracore was taken on the 155-ft terrace surface (Qrt) below the pit; basal organic material lying on lake sediments yielded a $^{14}C$ dates (obtained by L. McWeeney) of **8530±110 yr BP** (Beta-52258); two other dates from basal organics on a similar low terrace on the east side of the river 11 km to the south were **9370±100 yr BP** (Beta-52257) and **8530±80 yr BP** (Beta-52256). Before 9 ka, the Connecticut River had incised down into the Lake Hitchcock lakebed nearly to the level of the modern floodplain.

## REFERENCES


Black, R. F., 1974, Ice-wedge polygons of northern Alaska: in Coates, D.R., ed., Glacial Geomorphology, Fifth Annual Geomorphology Conference, Binghamton, N.Y. p. 247-275.

Black, R. F., 1976, Periglacial features indicative of permafrost: Quaternary Research, v. 6, p. 3-26.

Black, R. F., 1983, Pseudo-ice-wedge casts of Connecticut, northeastern United States: Quaternary Research, v. 20, p. 74-89.

Boothroyd, J. C. and Lawson, D. A., 1989, Wedge structures in southern Rhode Island: Evidence for postglacial permafrost conditions: Geological Society of America Abstracts with Program, v. 21, p.5.

Clebnik, S. M., 1980, Surficial geology of the Willimantic quadrangle, with map; Connecticut Geological and Natural History Survey, Quadrangle Report QR-39, 45 p.

Colton, R. B., 1960, Surficial geologic map of the Windsor Locks Quadrangle, Connecticut: U.S. Geological Survey Geologic Quadrangle Map GQ-137, scale 1:24,000.

Colton, R. B., 1965, Surficial geologic map of the Manchester Quadrangle, Connecticut, U. S. Geological Survey Geologic Quadrangle Map GQ-433, scale 1:24,000.

Davis, M. B. and Deevey, E. S., Jr., 1964, Pollen accumulation rates: Estimates from sediment of Rogers Lake; Science, v. 145, p. 1293-1295.

Davis, M. B., 1965, Phytogeography and palynology of northeastern United States, in Wright H.E. Jr., and Frey, D.G. eds., The Quaternary of the United States: Princeton, N.J., Princeton University Press, p. 377-401.

Davis, M. B., 1969, Climatic changes in southern Connecticut recorded by pollen deposition at Rogers Lake: Ecology, v. 50, no. 3, p. 409-422.

De Gans, W., 1988, Pingo scars and their identification, in Advances in Periglacial Geomorphology, New York, N.Y., John Wiley & Sons, Inc., p. 279-322.

Denny, C. S., 1936, Periglacial phenomena in southern Connecticut: American Journal of Science, v. 32, no. 191, p. 322-342.

Dyke, A. S. and Prest, V. K., 1987, Late Wisconsinan and Holocene history of the Laurentide Ice Sheet: Geographie Physique et Quaternaire 41(2): p. 237-263.

Flint, R. F., 1956, New radiocarbon dates and late-Pleistocene stratigraphy: America Journal of Science, 254: 265-287.

Gaudreau, D. C. and Webb, T. III, 1985, Late-Quaternary pollen stratigraphy and isochrome maps for the northeastern United States: in Pollen Records of Late-Quaternary North American Sediments, A.A.S.P. Foundation, p. 247-279.

Gustavson, T. C. and Boothroyd, J. C., 1987, A depositional model for outwash, sediment sources, and hydrologic characteristics, Malaspina Glacier, Alaska: A modern analog of the southeastern margin of the Laurentide Ice Sheet: Geological Society of America Bulletin, v. 99, p. 292-302.

Hartshorn, J. H. and Koteff, Carl, 1967, Geologic Map of the Springfield South quadrangle, Massachusetts-Connecticut; U.S. Geological Survey Geologic Quadrangle Map GQ-678, scale 1:24,000.

Hartshorn, J. H. and Schafer, J. P., 1965, INQUA Field Conference A, pt. 1, New England-New York: Lincoln, Nebraska, Nebraska Academy of Sciences, International Association of Quaternary Research, VIIth Congress, 92 p.

Jacobson, G. L., Jr., Webb, T. III, and Grimm, E. C., 1987, Patterns and rates of vegetation change during the deglaciation of eastern North America; in North America and adjacent oceans during the last deglaciation; Boulder, Colorado, Geological Society of America, The Geology of NOrth America v. K-3, p. 277-288.

Jahns, R. H., 1951, Surficial geologic map of the Mt. Toby quadrangle, Massachusetts; U.S. Geological Survey Geologic Quadrangle Map GQ-9, scale 1:31,680.

Johnson, W. H., 1990, Ice-wedge casts and relict patterned ground in central Illinois and their environmental significance: Quaternary Research, v. 33, p. 51-72.

Koteff, Carl, Stone, J. R., Larsen, F. D., Ashley, G. M., Boothroyd, J. C., and Dincauze, D. F., 1988, Glacial Lake Hitchcock, postglacial uplift, and post-lake archaeology, in Brigham-Grette, J. (ed.), Fieldtrip Guidebook AMQUA 1988, p. 169-208.

Koteff, Carl, and Pessl, Fred, Jr., 1981, Systematic ice retreat in New England: U.S. Geological Survey Professional Paper 1179, 20 p.

Larsen, F. D., 1979, Retreat of ice left eastern upland of Massachusetts subject to deep freeze; U.S. Geological Survey Professional Paper 1150, p. 58.

Leopold, E. B., 1955, Climate and vegetation changes during an interstadial period in southern New England: Yale University, New Haven, unpublished PhD thesis, 99 p.

Leopold, E. B., 1956, Two late-glacial deposits in southern Connecticut; Proceedings of the National Academy of Sciences, v. 42, no. 11, p. 863-867.

Mackay, J. R., 1979, Pingos of the Tuktoyaktuk Penninsula area, Northwest Territories: Geographie Physique et Quaternaire, v. 33, p. 3-61.

Mackay, J. R., 1988a, Pingo collapse and paleoclimate reconstruction; Canadian Journal of Earth Sciences, v. 25, p. 495-511.

Mackay, J. R., 1988b, The birth and growth of Porsild Pingo, Tuktoyaktuk Penninsula, District of Mackenzie: Arctic, v. 41, p. 267-274.

Mackay, J. R., 1990, Seasonal growth bands in pingo ice: Canadian Journal of Earth Sciences, v. 27, p. 1115-1125.

Miller, N. G. and Thompson, G. G., 1979, Boreal and western North American plants in the late Pleistocene of Vermont: Journal of the Arnold Arboretum of Harvard University, v. 60, p. 167-218.

Mitchell, G. F. 1973, Fossil pingos in Camaross Townland, Co. Wexford: Proceedings of the Royal Irish Academy v. 73, Sec. B, p. 269-282.

Oakes, L. A., 1992, Ice-wedge casts and pingo scars: evidence for former permafrost in Connecticut: Wesleyan University, Middletown, CT, unpublished senior thesis, 164 p.

O'Leary, D. W., 1975, Surficial Geologic Map of the Moodus Quadrangle, Connecticut, U.S. Geological Survey Geologic Quadrangle Map GQ-1205, scale 1:24,000.

Oldale, R. N., 1974, Geologic Map of the Dennis Quadrangle, Barnstable, County, Massachusetts: U.S. Geological Survey Geologic Quadrangle Map GQ-1114, scale 1:24,000.

Ridge, J. C. and Larsen, F. D., 1990, Re-evaluation of Antev's New England varve chronology and new radiocarbon dates of sediments from glacial Lake Hitchcock: Geological Society of America Bulletin, v. 42, p. 889-899.

Schafer, J. P., 1968, Periglacial features and pre-Wisconsin weathered rock in the Oxford-Waterbury-Thomaston area, western Connecticut: in Orville, P.M. (ed.), Guidebook for fieldtrips in Connecticut, New England Intercollegiate Geological Conference 60th Annual Meeting, New Haven, CT; Connecticut Geological and Natural History Survey Guidebook no. 2 , p. 1-5.

Schafer, J. P. and Hartshorn, J. H., 1965, The Quaternary of New England: in H.E. Wright, Jr. and Frey, D.G. (eds.). The Quaternary of the United States: Princeton, N.J., Princeton University Press, p. 113-128.

Schile, C. A., 1991, Stationary dune growth in the Connecticut River valley (Springfield South Quadrangle); University of Connecticut, Storrs, CT, unpublished master thesis, 82 p.

Sirkin, Les, 1982, Wisconsinan glaciation of Long Island, New York, to Block Island: in Larsen, G.J. and Stone, B.D., eds., Late Wisconsinan glaciation of New England: Dubuque, Iowa, Kendall/Hunt Publishing Company, p. 35-59.

Stone, B. D., 1978, Surficial Geologic map of the Jewett City Quadrangle, New London County, Connecticut, U.S. Geological Survey Geologic Quadrangle Map GQ-1434, scale !:24,000.

Stone, B. D. and Borns, H.W., 1986, Pleistocene glacial and interglacial stratigraphy of New England, Long Island, and adjacent Georges Bank and Gulf of Maine: in Sibrava, Vladimir, Bowen D.Q., and Richmond, G.M., eds., Quaternary Glaciations in the Northern Hemisphere: Quaternary Science Reviews, v. 5, p. 39-52.

Stone, B. D., Lapham, W. W, and Larsen, F. D., 1992, Glaciation of the Worcester Plateau, Ware-Barre area, and evidence for the succeeding late Woodfordian periglacial climate; Trip C-7, this volume.

Stone, J. R. and Ashley, G. M., 1989, Fossil Pingos?: New evidence for active permafrost in southern New England during Late Wisconsinan deglaciation: Geological Society of America Abstracts with Programs, v. 21 (2), p. 69.

Stone, J. R., Ashley, G. M. and Peteet, D. M., 1991, Cross section through a post-lake Hitchcock surface depression: New $^{14}C$ dates and evidence for periglacial origin: Geological Society of America, Abstracts with Programs, v. 23 (1), p. 135.

Stone, J. R., Schafer, J. P., and London, E. H., 1982, The surficial geologic maps of Connecticut illustrated by a field trip in central Connecticut; in Guidebook for fieldtrips in Connecticut and south-central Massachusetts, Guidebook No. 5, Connecticut Geological and Natural History Survey, Hartford, Connecticut, p. 5-30.

Stone, J. R., Schafer, J. P., London, E. H., and Thompson, W. B., 1992, Surficial Materials Map of Connecticut: U.S. Geological Survey Special Map; scale 1:125,000.

Stone, J. R. and Schafer, J. P., (in preparation) Quaternary Geologic Map of Connecticut, U.S. Geological Survey Special Map; scale 1:125,000.

Svensson, H., 1969, A type of circular lakes in northernmost Norway: Geografiska Annaler 51A, p. 1-12.

Thorson, R. M. and Webb, R. S., 1991, Postglacial history of a cedar swamp in southeastern Connecticut: Journal of Paleolimnology v. 6, p. 17-35.

Thorson, R. M., Clayton, W. S. and Seeber, Leonardo, 1986, Geologic evidence for a large prehistoric earthquake in eastern Connecticut: Geology, v. 14 (6), p. 463-467.

Vandenberghe, J., 1988, Cryoturbations in Clark, M.J. (ed.), Advances in periglacial geomorphology, John Wiley & Sons Ltd. p. 179-198.

Van Vliet-Lanoe, Brigitte, 1988, The significance of cryoturbation phenomena in environmental reconstruction; Journal of Quaternary Science, v. 3, p. 85-96.

# PETROLOGY OF THE HIGH-ALUMINA HOOSAC SCHIST FROM THE CHLORITOID+GARNET THROUGH THE KYANITE+BIOTITE ZONES IN WESTERN MASSACHUSETTS

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## INTRODUCTION

The eastern limb of the Berkshire anticlinorium of western Massachusetts (Figure 1) is a complex, multiply-deformed, polymetamorphic, Taconian/Acadian orogenic terrane. The geologic framework of this area is well established, originally by the mapping of B.K. Emerson (1892, 1898, 1899) and Pumpelly *et al.* (1894), as summarized on the Massachusetts geologic map of Emerson (1917), and more recently by the mapping of L.M. Hall, N.L. Hatch, S.A. Norton, P.H. Osberg, N.M. Ratcliffe, and R.S. Stanley, as summarized on the Massachusetts geologic map of Zen *et al.* (1983). The summary reports of USGS Professional Paper 1366 in 1988 as well as the work of Hatch *et al.* (1984), Stanley and Ratcliffe (1985), and Sutter *et al.* (1985), among others, provide a provocative regional synthesis that brings into sharp focus a variety of interrelated structural, stratigraphic, petrologic, and geochronologic problems.

Despite vigorous efforts, our ability to constrain the timing of many fundamental events is still hampered by both the complexity of the terrane and a lack of data. As reviewed by Karabinos and Laird (1988), differentiating between the effects of different metamorphic events remains quite problematic in much of the terrane. The recent work of Hames *et al.* (1991) and Armstrong *et al.* (1992) emphasizes the problem of differentiating between Taconian and Acadian orogenic effects along the zone of maximum overlap, which generally coincides with the axis of the Berkshire massif. This field trip (see figure 10 for route) will review the nature of this polymetamorphism in a nearly continuous belt of high-alumina, Gassetts-like schists of the Hoosac formation that occurs along the eastern margin of the Berkshire massif. As a bonus we will have the opportunity to examine the nearly continuous prograde metamorphic evolution of a relatively unusual, but mineralogically interesting, bulk composition that has historically received much attention.

## REGIONAL SETTING

The Berkshire massif consists of "nested thrust slices" of Middle Proterozoic (~1 Ga), "Grenvillian", metamorphosed sedimentary, granitic, and volcanic gneisses and their unconformable cover rocks of the Cambrian Dalton Formation (Ratcliffe *et al.*, 1988). The antiformal structure of the Berkshire massif results from Acadian folding of Proterozoic to Lower/Middle Devonian age rocks that manifest a variety of stratigraphic, metamorphic facies, and structural relationships produced in part by the older Taconian orogeny (Robinson, 1986). As shown on the Bedrock Geologic Map of Massachusetts (Zen *et al.*, 1983), summarized by Hatch *et al.* (1984) and shown on Figure 2, the Berkshire massif and its eastern cover sequence can be represented by three lithotectonic assemblages: the Taconic-Berkshire Zone, the Rowe-Hawley Zone, and the Bronson Hill Zone.

The Taconic-Berkshire Zone includes an allochthon of ~1 Ga basement gneisses of the North American craton. West of this allochthon are autochthonous basal clastic rocks that rest unconformably on this same basement and that are succeeded by Cambrian to Lower Ordovician carbonate bank deposits. These miogeoclinal rocks are overlain by allochthonous Cambrian to pre-Middle-Ordovician clastic sediments and minor volcanic rocks of the Taconic allochthons. The Taconic allochthons are thought by Stanley and Ratcliffe (1985) to be the eastern facies of the autochthonous section and, thus, to have been transported from east of the continental basement. The Late Proterozoic to Lower Cambrian Hoosac Formation is believed to be a "western facies," equivalent to some of the allochthonous section, but remains east of the Berkshire massif where it is in fault contact with basement gneisses along the Middlefield-Hoosac Summit thrust (Stanley and Ratcliffe, 1985).

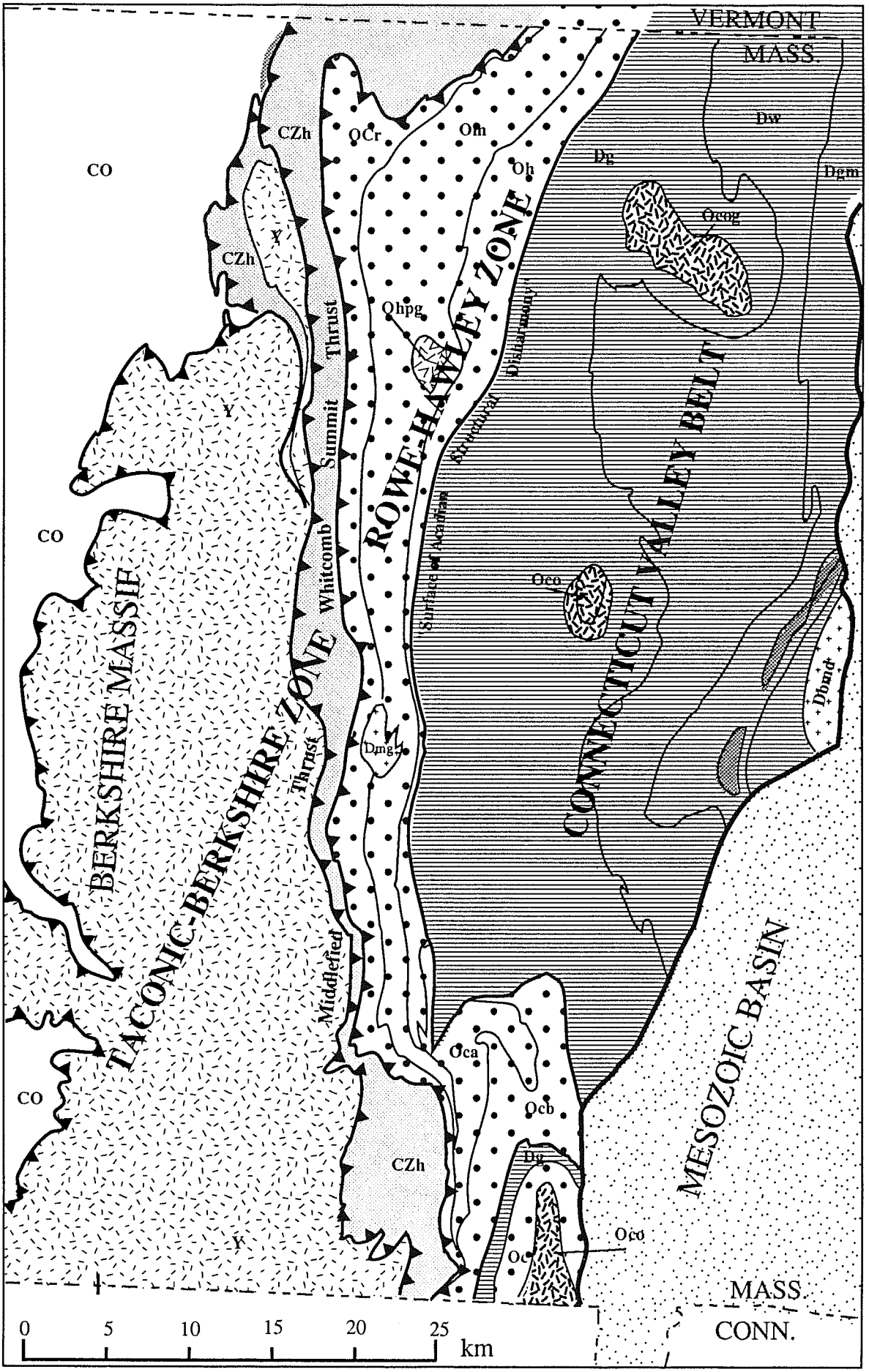


**FIGURE 1.** Generalized geologic map of the Berkshire massif, modified from Zen *et al.*(1983). The map symbols are as shown on the State Map of Massachusetts, see Zen *et al.*(1983) for details.

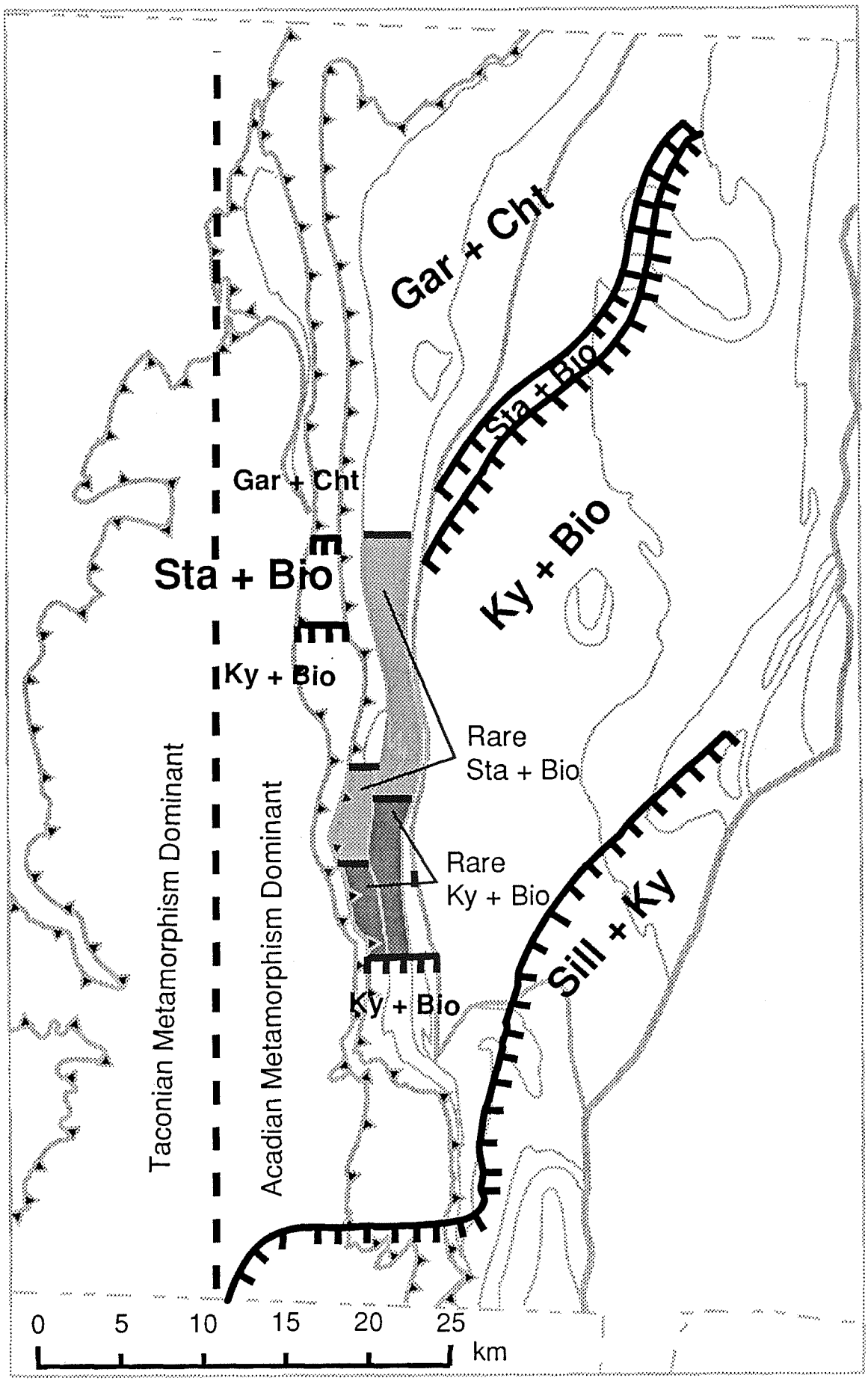


**FIGURE 2.** Metamorphic isograd map for the east limb of the Berkshire massif. Geology as shown on figure 1.

The Hoosac Formation, regarded by Stanley and Ratcliff (1985) as the source of the Taconic allochthons, is separated from Cambrian to Ordovician rocks of the Rowe-Hawley Zone by the Whitcomb Summit thrust of Taconian age. This fault is interpreted as the northern extension of Cameron's line in Connecticut. It separates the Hoosac Formation to the west, originally deposited upon Grenvillian age continental basement, from rocks of the Rowe, Moretown, and Hawley (or Cobble Mountain) Formations to the east, probably deposited upon oceanic crust (Hall and Robinson, 1982; Zen *et al.*, 1983). As developed by Stanley and Hatch (1988), the "Whitcomb Summit thrust carries the Rowe-Hawley Zone over the root zone of the Taconic allochthons." Hatch and Stanley (1988) have interpreted the Rowe Schist as a complex tectonite zone with displacements along internal thrusts of the same magnitude (tens to hundreds of kilometers) as estimated for the Whitcomb Summit and the Middlefield thrusts.

The border of the Rowe-Hawley Zone with the Bronson Hill Zone to the east is covered by the Silurian-Devonian Connecticut Valley Belt. The contact between the Hawley Formation to the west and the Goshen Formation of the Connecticut Valley Belt (locally called the RMC) is interpreted as a "surface of Acadian structural disharmony" (Hatch *et al.*, 1988), although other interpretations have been entertained (see Hatch and Stanley, 1988). The concealed contact between the Rowe-Hawley Zone and the Bronson Hill Zone is the "upper surface of the east dipping subduction zone along which the Iapetus ocean disappeared beneath the Bronson Hill plate" (Hatch *et al.*, 1984).

The Taconian and Acadian events that assembled the package of rocks now comprising the east limb of the Berkshire massif have imparted a complex structure to the area and a complex fabric to these rocks. A detailed Taconian-Acadian orogenic history has been established that includes up to five stages of folding and multiple pulses of metamorphism expressed by differential mineral growth relative to the resulting fabrics (Hatch, 1975; Norton 1975a,b; Stanley, 1975; Stanley and Hatch, 1988; Ratcliffe *et al.*, 1988). At least three distinct generations of folds are present in post-Taconian rocks and thus must be Acadian (Hatch and Stanley, 1988). The Acadian folding is time transgressive, especially along the north to south axis of the orogen. Although the intensity of Acadian deformation and associated schistosity diminishes to the west, Acadian fabrics clearly overprint and dominate Taconian fabrics at least to the western edge of the Rowe-Hawley Zone. In the Rowe Schist and Hoosac Formation the dominant schistosity is parallel or sub-parallel to Taconian age schistosity traced eastward from the Taconic-Berkshire Zone (Hatch and Stanley, 1988; Ratcliffe *et al.*, 1988). However, as reviewed by Hames *et al.* (1991) and Armstrong *et al.* (1992), Acadian folding is present well to the west of the Whitcomb Summit thrust and, indeed, to the west of the Berkshire massif.

Similarly, there is considerable uncertainty in the age of metamorphism and extent of polymetamorphism affecting the Cambrian-Ordovician rocks. In particular, the western extent of Acadian fabric development and mineral growth is difficult to ascertain and the separation of Acadian-dominant vs Taconian-dominant metamorphism has not been documented in detail. Karabinos and Laird (1988) point out that the uncertainty in the time of metamorphism is due to conflicting age dates, ambiguous rock fabrics, and the complex metamorphic/tectonic history of the area. However, recent geochronologic studies of Sutter *et al.* (1985) and Hames *et al.* (1991) have generally confirmed the overprint pattern shown on the Metamorphic Map of Massachusetts (Zen *et al.*, 1983). Specifically, the Acadian-dominant boundary seems to be within the Berkshire massif, west of the Whitcomb Summit thrust (Cameron's line), and perhaps even west of the Middlefield-Hoosac Mountain thrust.

## METAMORPHISM

The distribution of metamorphic isograds as shown on the Metamorphic Map of Massachusetts (Zen *et al.*, 1983) obscures the complex nature of the metamorphism resulting from the events chronicled above. As originally described by Hatch (1975), Norton (1975a), Hatch and Stanley (1976), and reviewed by Hatch and Stanley (1988), there has been some difficulty in mapping consistent metamorphic isograds across structural/stratigraphic contacts in the east limb of the Berkshire massif.

Figure 2 is a metamorphic map for the east limb of the Berkshire massif. An earlier version of this map was used in the compilation of the Metamorphic Map of Massachusetts (Zen *et al.*, 1983). Figure 2, based upon petrographic descriptions of over 2000 thin sections, represents results from ongoing, detailed petrographic studies aimed at refining our understanding of the complex sequence of tectonic/metamorphic events in this polygenetic

terrane. Much of this work has been done by seniors at Amherst College as honors theses or independent study projects. These field-oriented, petrographic studies by undergraduate students combined with the studies of of Abbott (1979),Downie (1975, 1979) and Pferd (1981) have been instrumental in building on earlier work to locate in detail several of the isograds shown on Figure 2. These isograds include: (1) the St+Bio isograd within the Hoosac, Moretown, and western portion of the Goshen Formations (Sander, 1977), (2) the Ky+Bio and St+Bio isograds in the Goshen (Dibble, 1981), Ashfield (Hudson, 1983), and Colrain (Pferd, 1981; Goeldner, 1983) Quadrangles, and (3) the "Sillimanite isograd" in the Hoosac Formation (Bryan, 1978). In addition, these studies and those of Conner (1979) in the Westhampton Quadrangle, Britt (1980) in West Granville, Cohan (1980) and Handy (1980) in Woronoco, Hickmott (1982) in Worthington, and Maggs (1984) in South Sandisfield have expanded coverage of the large area comprising the east limb of the Berkshire massif.

As shown on Figure 2, the east-west trending St+Bio and further south Ky+Bi isograds established during bedrock mapping (see Zen *et al.*, 1983 for references) are difficult to trace across and are conceivably offset along the north-south trending RMC (Taconic Line) and Whitcomb Summit thrust faults. These structural discontinuities separate Cambrian-Ordovician units from the older Hoosac and younger Goshen Formations. The resulting pattern of apparent metamorphic grades consists of a north-south-trending Gar+Cht Zone "core" comprised of the Rowe, Moretown and Hawley Formations that is flanked by successively higher grade rocks, from north to south, of the Hoosac and Goshen Formations. Approximately 20 km south of the Ky+Bio isograds mapped in the flanking units, the transition from Gar+Cht Zone to Ky+Bio Zone occurs within the "barren core" over a short distance of 1-2 km. The initial occurrences of staurolite and kyanite within these barren units are "so complexly distributed that consistent isograds could not be mapped" (Hatch and Stanley, 1976).

Post metamorphic faulting might explain the differential nature of apparent metamorphic grade characterizing various stratigraphic portions of the terrane including the apparently offset isograds. However, a more plausible explanation for these complexities and the resulting map pattern is based upon the effect of spessartine and grossular components of garnet on the AFM discontinuous reaction Gar+Cht = St+Bio. All available mineral assemblage and analytical data are generally consistent with a model, described by Cheney *et al.* (1980), that involves differential stabilization of the Gar+Cht join by the spessartine and/or grossular components in garnet (Figure 3). Specifically, the near-rim analytical sum CaO+MnO is systematically higher in garnets from the Cambrian-Ordovician "barren core" (Rowe, Moretown, and Hawley Formations) relative to garnets from the flanking Hoosac and Goshen Formations (Figure 4). As shown (Figure 3) by the biotite projections of Cheney *et al.* (1980; see also Spear and Cheney, 1989), the addition of MnO and/or CaO causes the AFM discontinuous reaction Gar+Cht = St+Bio to become a continuous reaction. Thus, in rocks containing MnO+CaO-richer garnet, the coexistence of staurolite with biotite should not occur until higher temperatures are reached. Assuming that the observed concentration of these extra components in the garnet rims is proportional to the increased temperature of the isograd reaction, the St+Bio isograd should occur at lowest grade in the Goshen and Hoosac Formations and at higher grades in the "barren core." These results are consistent with similar conclusions based on different approaches of Hatch and Stanley (1988), Sutter and Hatch (1985) in Massachusetts, and Laird *et al.* (1991) in southern Vermont.

Sources of uncertainty in this interpretation include the relatively small sample population, the extreme compositional zoning characteristic of most garnets, and the uncertain status of chlorite in many rocks. An additional complexity pertaining to the significance of metamorphic isograds in this terrane is the uncertainty in the extent of polymetamorphism and age of metamorphism affecting even the Silurian-Devonian rocks in this area. As pointed out by Sander (1977), Abbott (1979), Pferd (1981), and others, the Garnet, Staurolite and Kyanite isograds are multiple-event phenomena as indicated, for example, by initial occurrences of garnet, and at higher grade staurolite, in mappable zones of chlorite pseudomorphs.

## HIGH-ALUMINA SCHISTS OF THE HOOSAC FORMATION

The Hoosac Formation in western Massachusetts is a Proterozoic Z to Cambrian allochthon, isolated by Taconian thrust faults, that forms part of the eastern cover of the Berkshire massif. The rocks comprising the Hoosac Formation have been significantly affected by both the Taconian and Acadian orogenies and the rocks now reside in the zone of maximum overlap between these events. As shown on the Bedrock Geologic Map of Massachusetts (Zen *et al.*, 1983) and described in some detail by Norton (1976) and more recently by Ratcliffe *et al.*

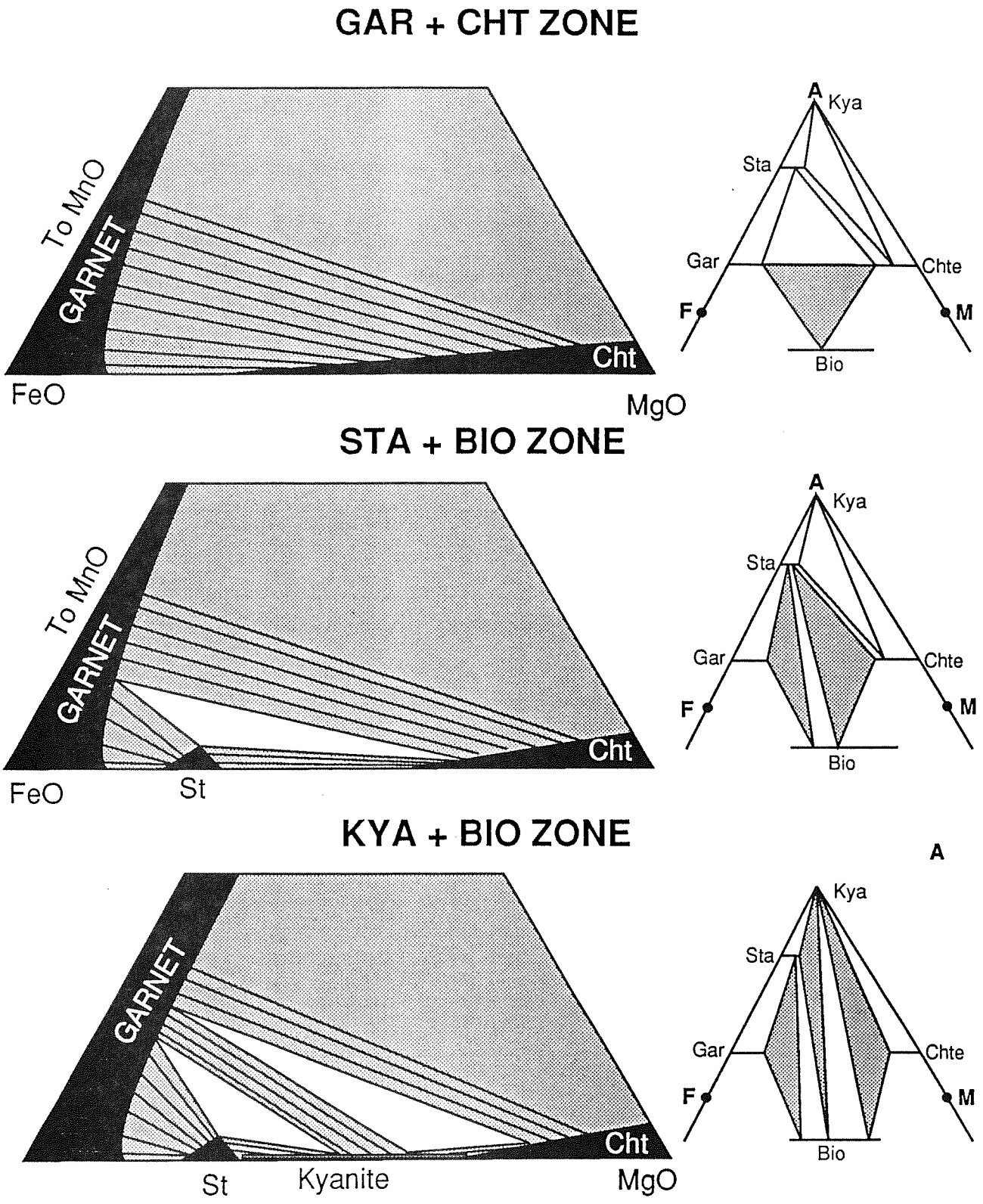


**FIGURE 3.** Schematic projections of phase relationships from "Biotite", Quartz, Muscovite, and $H_2O$ onto the FeO-MgO-MnO plane of the AKFM-Mn subsystem. AFM projections are shown for reference.

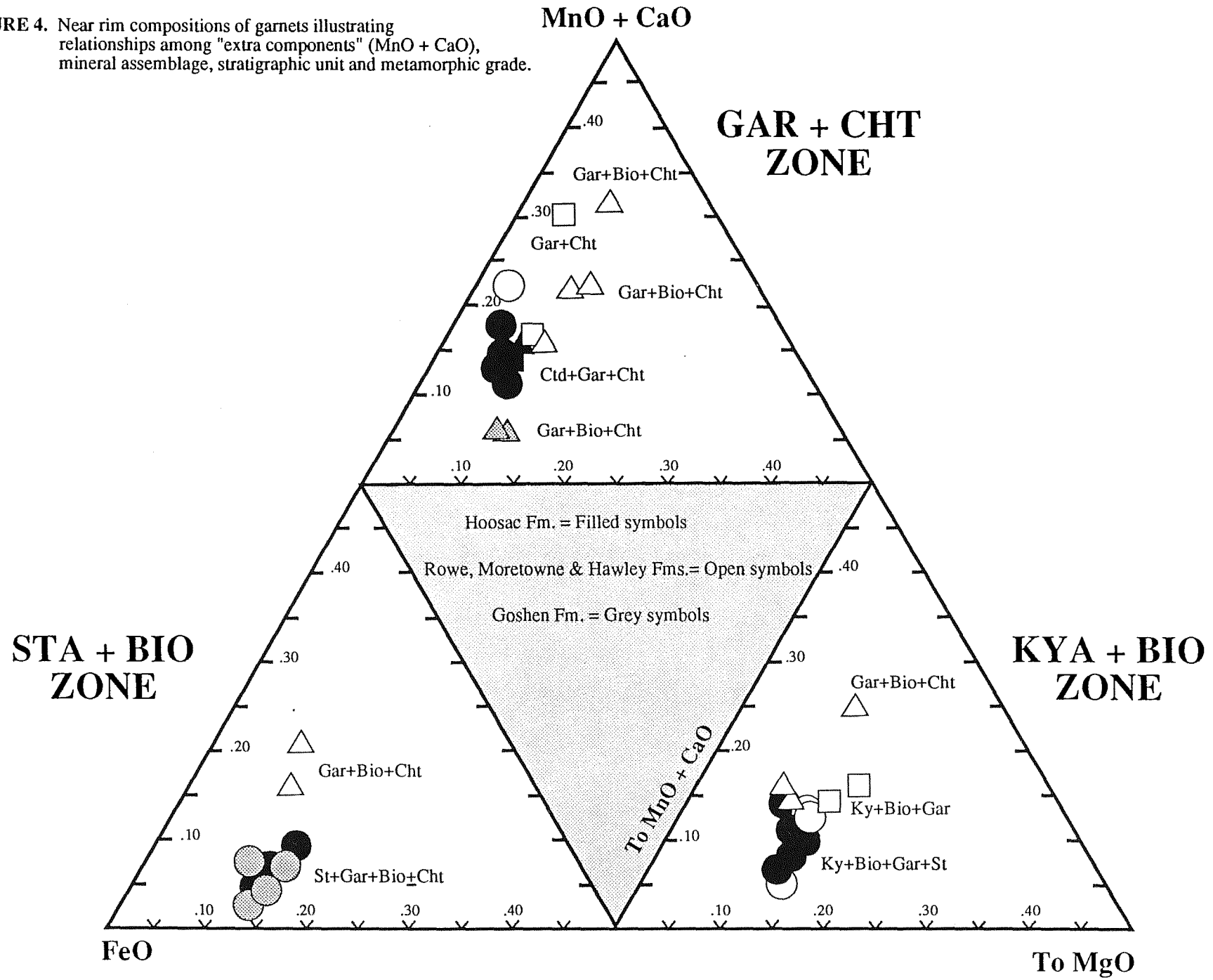


**FIGURE 4.** Near rim compositions of garnets illustrating relationships among "extra components" (MnO + CaO), mineral assemblage, stratigraphic unit and metamorphic grade.

(1988), the Hoosac formation can be divided into at least seven units. As observed by Norton (1976), most of the formation consists of "rather monotonous" grey to rusty, medium-grained Mus-Ab-Qtz schist to granofels that may include a variety of accessory minerals such as chlorite, biotite, garnet, staurolite, kyanite, and sillimanite, depending upon metamorphic grade and bulk composition. Of particular interest here are the high-alumina pelitic schists (CZhgt of Zen *et al.*, 1983) characterized by large (1 cm) garnets in a Qtz-Mus±Paragonite-rich matrix that extend in mappable bands from Vermont to near the Connecticut border. Even at the lowest metamorphic grade the overall coarse grain size of these rocks is striking. However, as pointed out by Norton (1976) and emphasized by Ratcliffe *et al.* (1988), there are several different high alumina lithologies mapped within the Hoosac Formation. Ratcliffe *et al.* (1988) suggest that the upper large garnet schist (CZhg of Zen *et al.*, 1983) may be equivalent to high-alumina compositions in the Rowe Formation (Pinney Hollow Formation in Vermont). The stratigraphically lower, big garnet (aluminous) unit (CZhgt of Zen *et al.*, 1983) has been correlated with the Gassetts schist in Vermont as indicated on the Bedrock Geologic Map of Massachusetts. The spatial extent of metamorphic zones (Figure 5), compositions of porphyroblasts, distribution of inclusions within garnet, and the compositional zoning of garnet from these aluminous rocks in Massachusetts are generally consistent with the prograde evolution of the Gassetts, Vermont location described by A.B. Thompson and co-workers (Thompson *et al.*, 1977a,b) as well as subsequent investigations on similar rocks in southeastern Vermont including those of Downie (1980, 1982), Crowley (1989), Karabinos (1984, 1985), Cook (1988), Cook and Karabinos (1988), and Giaramita and Day (1991).

### AFM Assemblages and Isograds

The high-alumina Hoosac schists of western Massachusetts contain mineralogic and textural evidence of polymetamorphism. Compositionally-zoned megacrysts of white mica (phengite cores and muscovite rims) and, in rocks of appropriate grade, homogeneous and largely undeformed porphyroblasts of muscovite, staurolite, kyanite, biotite, chlorite, and chloritoid crosscut highly folded schistosities defined by phengitic muscovite and, in some rocks of appropriate grade, paragonite, chlorite, chloritoid, staurolite and kyanite. The mineral assemblages based upon the overprinting porphyroblasts are generally consistent with the four metamorphic zones that reflect an increase of metamorphic grade from north to south as shown on Figure 5. These isograds are very similar to those of the Metamorphic Map of Massachusetts (Zen *et al.*, 1983), except there is more detail shown here. For example, due to the restricted compositions of the rocks and detailed sample coverage it has been possible to separate the Mg/Fe-dependent Kyanite isograd from the Staurolite isograd.

The mineral assemblages of the Mus+Qtz-bearing schists are generally consistent with the AFM mineral assemblage diagrams of Figure 5. Because these rocks have mineral assemblages that reflect bulk compositions above the Gar+Cht join on the AFM projection, Zone I, the Garnet Zone, can be divided into three prograde "sub-zones": (Ia) Cht+Gar+Ctd; (Ib) St+Cht+Gar (with extra components)+Ctd; (Ic) St+Cht+Gar. Due to the lack outcrops in the critical area (see Figure 5), assemblage Ic rocks have not yet been observed. Although Zone II is the traditional Staurolite Zone defined on the basis of St+Bio compatibility, the "staurolite" isograd is actually a "biotite-in" isograd and biotite typically occurs only at the higher grade in high alumina rocks as textually late laths and in low modal amounts (<10%). The Kyanite Zone, Zone III, can also be divided in to two subzones, although some overlap exists due to the stabilizing effect of calcium in feldspar (see Cheney and Guidotti, 1979): (IIIa) Ky+Bio+Par and (IIIb) Ky+Bio+Pla. Zone IV, the Sillimanite Zone (Sil+Bio), as is typical in this terrane, is marked initially by rocks containing fibrolite with kyanite as discussed by Hames *et al.* (1991). The sillimanite isograd occurs in the Hoosac Formation south of the southern most occurrences of Gassetts like schist (Figure 5) and will not be considered further.

### Mineral Chemistry:

The compositions of minerals from 15 samples of high-alumina rock as shown on Figure 5 have been determined by electron microprobe analysis. This data base typically consists of 50-100 point analyses of large (>1.0 cm) zoned garnets and 30-100 point analyses of complex white mica assemblages in addition to a minimum of 3-10 point analyses of most other minerals comprising each sample. Various aspects of the chemical data in conjunction with the petrographic work on some 150 Hoosac samples have been partially presented by Cheney *et al.* (1980) and Cheney (1980; 1986). Mineral composition data are summarized on Figure 6, a semi-log diagram (an Albee plot; see Albee, 1972) with the logarithm of compositional variables plotted along the X-axis against a Y-axis

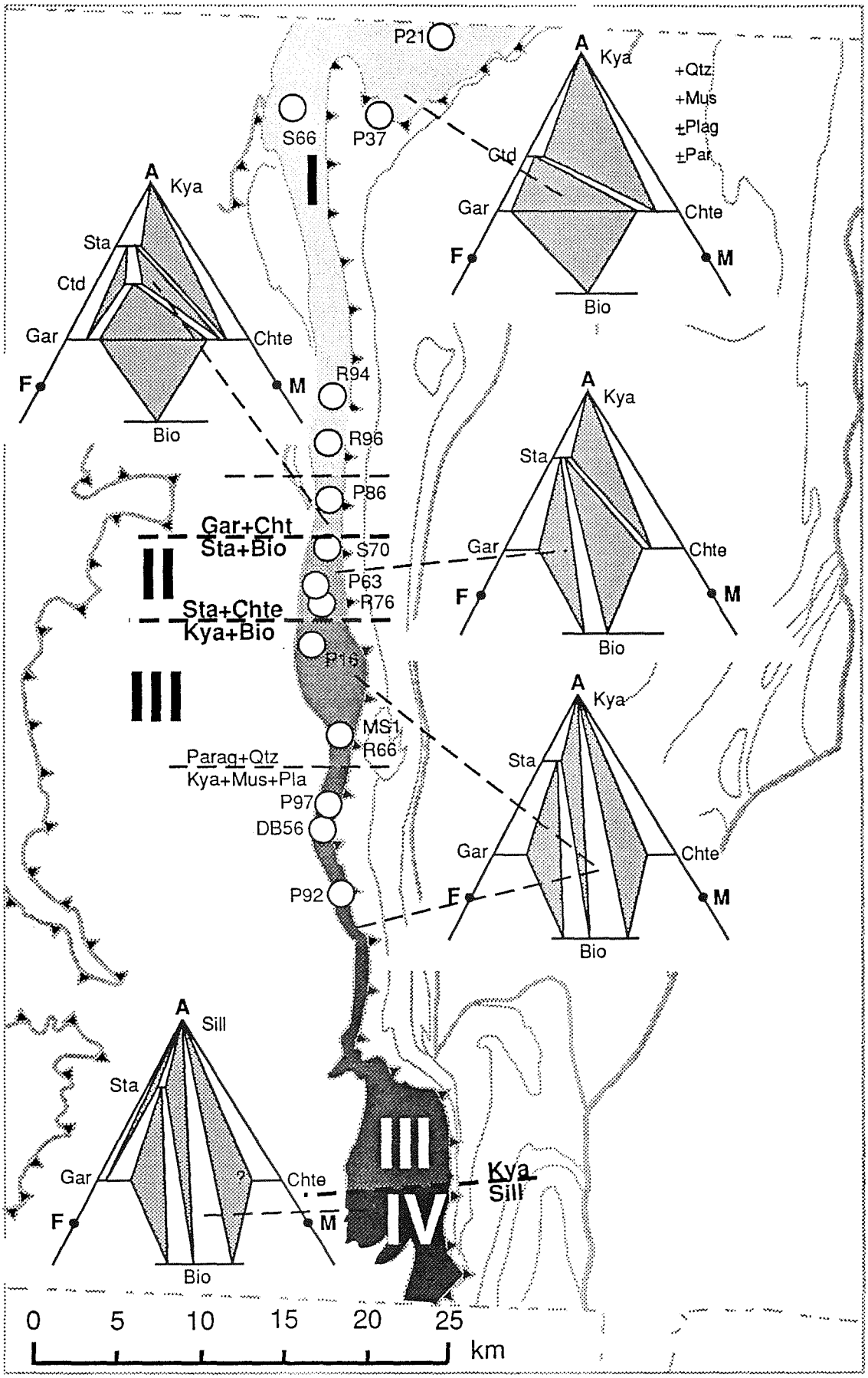


FIGURE 5. Metamorphic map for the Hoosac Formation showing the distribution of analyzed samples, metamorphic zones, and schematic AFM projections. The geology is as shown on figure 1.

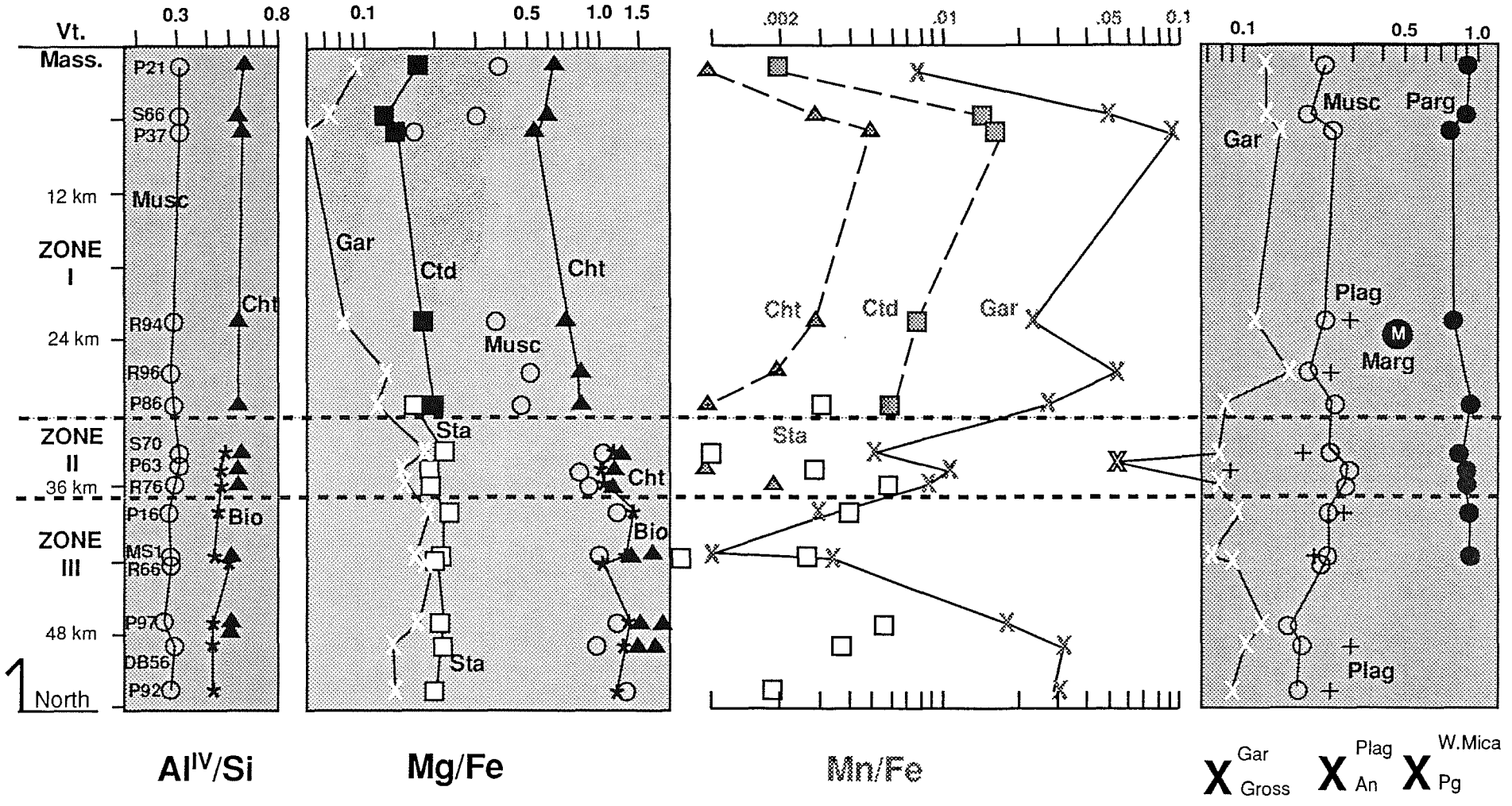


**FIGURE 6.** Semi-log diagram summarizing electron probe data for minerals from high-alumina schists of the Hoosac Formation. The logarithms of compositional variables (horizontal axis) are shown as a function of distance south from the Vermont border (vertical axis). The mineral assemlage for each sample plots along a horizontal line. As discussed by Albee (1972), the horizontal separation of two points (*e.g.* Mg/Fe) for a mineral pair is related to the Kd (*e.g.* $(Mg/Fe)^{Gar}/(Mg/Fe)^{Bio}$).

that records distance in kilometers south from the Vermont-Massachusetts border. These data are generally from fabric-cutting porphyroblasts and the compositions of most minerals change in a manner consistent with prograde evolution as observed in many terranes. Of particular note is that:

1) The Mg/Fe ratios of chloritoid and chlorite increase with increasing grade. The chlorite in some St+Bio Zone and higher grade samples clearly post-dates most of the other porphyroblasts and is "retrograde," although other hypotheses have been advanced for similar chlorite occurrences in similar grade rocks by Karabinos (1984) in the Jamaica, Vermont area and Giaramita and Day (1991) at Gassetts, Vermont (compare Guidotti, 1974).

2) The Mg/Fe ratio of biotite increases with increasing grade. The Mg/Fe ratio of garnet increases from Zone I through Zone II and then decreases in the Kyanite Zone.

3) The element partitioning among phases ($K_D$), as indicated by the horizontal separation between the same ratio for different minerals, is generally systematic and consistent with prograde evolution. However, in detail the compositions of some porphyroblasts may reflect equilibration at different conditions during P-T-t evolution.

4) The Pg (paragonite) content of the muscovite porphyroblasts and megacryst rims increases from the Gar+Cht Zone (I) through the St+Bio Zone (II). Within the Ky+Bio Zone (III), the Pg content of muscovite attains a maximum of ~33% Pg, coincident with the disappearance of paragonite from the groundmass, and then decreases with further rise of grade.

5) Of particular interest is that the tschermak content of these muscovites, as shown by $Al^{IV}/Si$, is very nearly constant from the Gar+Cht Zone to the Ky+Bio Zone.

## Fabrics, Garnets and Multiple Events

Relationships summarized heretofore suggest that muscovite, the other porphyroblasts, and some of the paragonite reflect a prograde metamorphism that was superimposed on a pre-existing-fabric (possibly Taconian). In most samples this fabric is defined by finer-grained laths of celadonitic (low-Na, low-Al) muscovite and even finer-grained laths of paragonite. Low grade rocks (Zone I) also have fine-grained laths of chloritoid and chlorite in these complexly folded fabrics. At higher grades (Zones II and III), the fabrics are defined primarily by white mica, although rare fine-grained staurolite and kyanite do occur aligned in the fabrics and may be deformed (broken /bent). These are most common in Zone II (St+Bio), just north of the initial occurrence of kyanite porphyroblasts.

Additional indications of an early event, superseded by a later event of similar grade, include the occurrence of garnets with unconformity textures similar to those described by Rosenfeld (1968, see also Rosenfeld *et al.*, 1988). The Hoosac garnets from Massachusetts are remarkably similar in both appearance and chemical zoning (see Figure 7 for some representative zoning patterns) to unconformity garnets from high-alumina schists of the Hoosac Formation near Jamaica, Vermont described by Karabinos (1984). Karabinos suggested that two stages of garnet growth, separated by a retrogressive resorption event, are required to explain the chemical zoning that accompanies the overgrowth of inclusion-free rims on inclusion-riddled cores. The validity of this model has received recent support from the isotopic studies of Christensen *et al.* (1989), Chamberlain and Conrad (1991), and Young and Rumble (1992, in preparation). Moreover, multiple sizes of garnet porphyroblasts occur in many of these rocks as described by Bashir (1989) and Crowley (1989) in samples from southeastern, Vermont. In the Vermont occurrences, the smaller garnets have compositions similar to the rims from the larger garnets (Bashir, 1989). An additional complexity mandating caution in the interpretation of both optical and chemical data from large garnets is that some of the larger garnets appear to have originated from the coalescence of several smaller garnets and that other large garnets seem to have formed from the apparent fracturing of early garnets and subsequent annealing. Similar textures were reported by Crowley (1989) for garnets from the Gassetts schist and Pinney Hollow Formations of the Star Hill sigmoid in Cavendish, Vermont.

### Thermobarometry

Due to the small modal amounts of biotite in these rock, the use of biotite geothermometers is problematic at best due to the expected tie line rotation on cooling noted by Tracy *et al.* (1976). Two additional complexities are the chloritization effect and the garnet resorption effect reviewed by Spear (1989, 1991). Bearing in mind these problems, Figure 8 shows some remarkably consistent geobarometric results from all four of our Ky+Pla rocks. These results were obtained from an updated version (by P. Crowley) of the Hodges and Crowley (1985) calibration. Of interest is that when combined with the Ky+Bio stability curve of Spear and Cheney (1989), the resulting minimum pressures for Zone III rocks in this terrane are on the order of 8 kilobars for the crosscutting porphyroblast assemblages. These pressures are consistent with pressures of 7 to 10 kilobars obtained by similar methods from clearly Acadian minerals of the Goshen Formation in Western Massachusetts (Goeldner, 1983; Hudson, 1983) and southeastern Vermont (Davidow, 1989) as well as with the more recent work summarized in Hames *et al.* (1991) and Hickmott and Spear (1992, in press).

### Muscovite and Multiple Events

Textually complex white mica assemblages can be divided into at least three types: (1) foliation-defining laths of celadonitic muscovite and some paragonite, (2) crosscutting porphyroblasts (coarse lath and tablets) of high-Na normal muscovite (Si≈6.1 atoms/22 oxygens), and (3) compositionally-zoned megacrysts with celadonitic core and "normal" rim compositions (documented at least into Zone IIIB, see Figure 9). As shown on Figure 9 there is a remarkable range in the compositions of muscovites from these rocks that correlates with these textural habits. Moreover, as has been commonly observed (*e.g.* Guidotti, 1984; Dempster, 1992), the tschermak and paragonite contents of the white micas are apparently coupled in such a way that there is a systematic relationship between high sodium (paragonite component) and high-alumina (tschermak component) contents.

The compositional variation of zoned megacrysts and/or porphyroblasts and groundmass laths from individual samples define systematic paths characterized by decreasing celadonite and decreasing Mg/Fe ratio with increasing Na/K ratio on the muscovite plane of the AKFM tetrahedron (Figure 9). As developed by Thompson (1979), the AKFM composition of muscovite is buffered, at constant P-T-$\mu_{H2O}$, in three-phase AFM assemblages (*e.g.*, Ky+St+Bio or Ctd+Cht+St). Hence, the composition of muscovite in these assemblages will vary in a predictable fashion and define paths on the muscovite plane of the AKFM tetrahedron that reflect variation in these intensive parameters. Muscovite compositional ranges are comparable to those discussed by Dempster (1992, and references therein) for normal (i.e. biotite-bearing) bulk compositions.

## TECTONIC IMPLICATIONS OF HOOSAC PETROLOGY

The observation that from Zone I through IIIb the tschermak content of the muscovite porphyroblasts remains relatively constant is of particular importance. Zone I contains the same matrix assemblage, Gar+Cht+Ctd, as the cores of many garnets. Thus, the garnets probably grew in this assemblage. However, the groundmass muscovite in nearly all rocks from this terrane, independent of "current" grade, has a very different composition than even the lowest grade muscovite porphyroblast. The data shown on Figure 9 clearly indicate that the groundmass phengitic muscovites and the muscovite porphyroblasts have formed at very different conditions. Based on the exploratory experimental work of Massone and Schreyer (1987), the phengitic micas must reflect either a temperature substantially lower than the Zone I mineral facies or pressures greater than those at which the muscovite porphyroblasts formed. Because the groundmass phengitic micas occur with other groundmass minerals consistent with at least the Zone I mineral facies, it is very plausible that the early event recorded by these rocks is a higher-pressure event than that recorded by the overgrowth assemblages. Thus, the core-to-rim compositional paths shown on Figure 9 generally reflect the prograde adjustment of lower T/higher P rocks to new metamorphic conditions. Additional support for this model includes the occurrence of irregular polycrystalline zones of plagioclase on the edges of some garnets, the sharp decrease in the grossular content in the outer rims of zoned garnets (see figure 7), and the restriction of rutile to the cores of these garnets. Ilmenite occurs in the garnet rims and it is the groundmass Ti -phase.

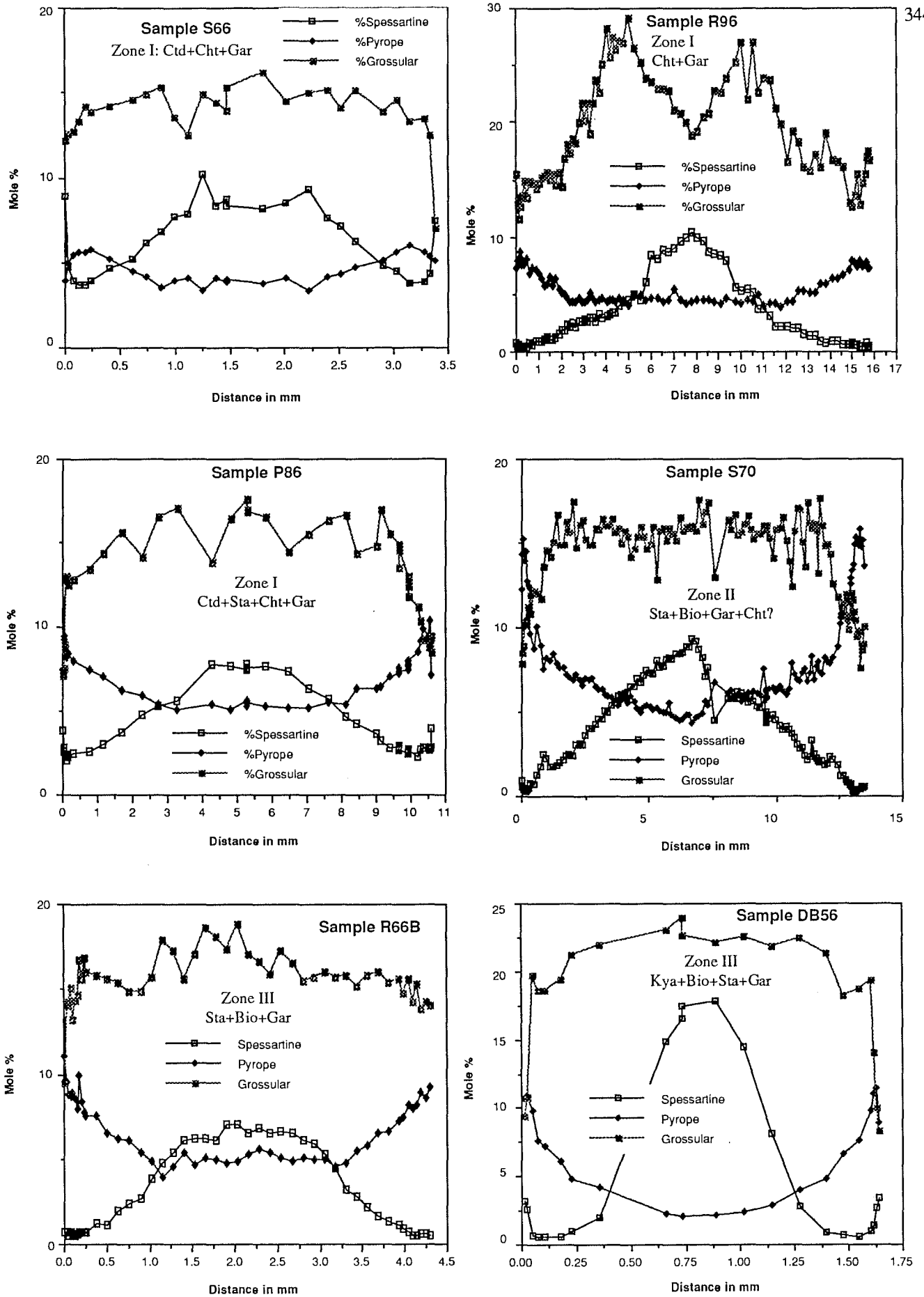


**FIGURE 7A.** Garnet zoning profiles for mole % spessartine, pyrope and grossular from representative Hoosac high-alumina schists. Almandine zoning for these same samples is shown on figure 7B.

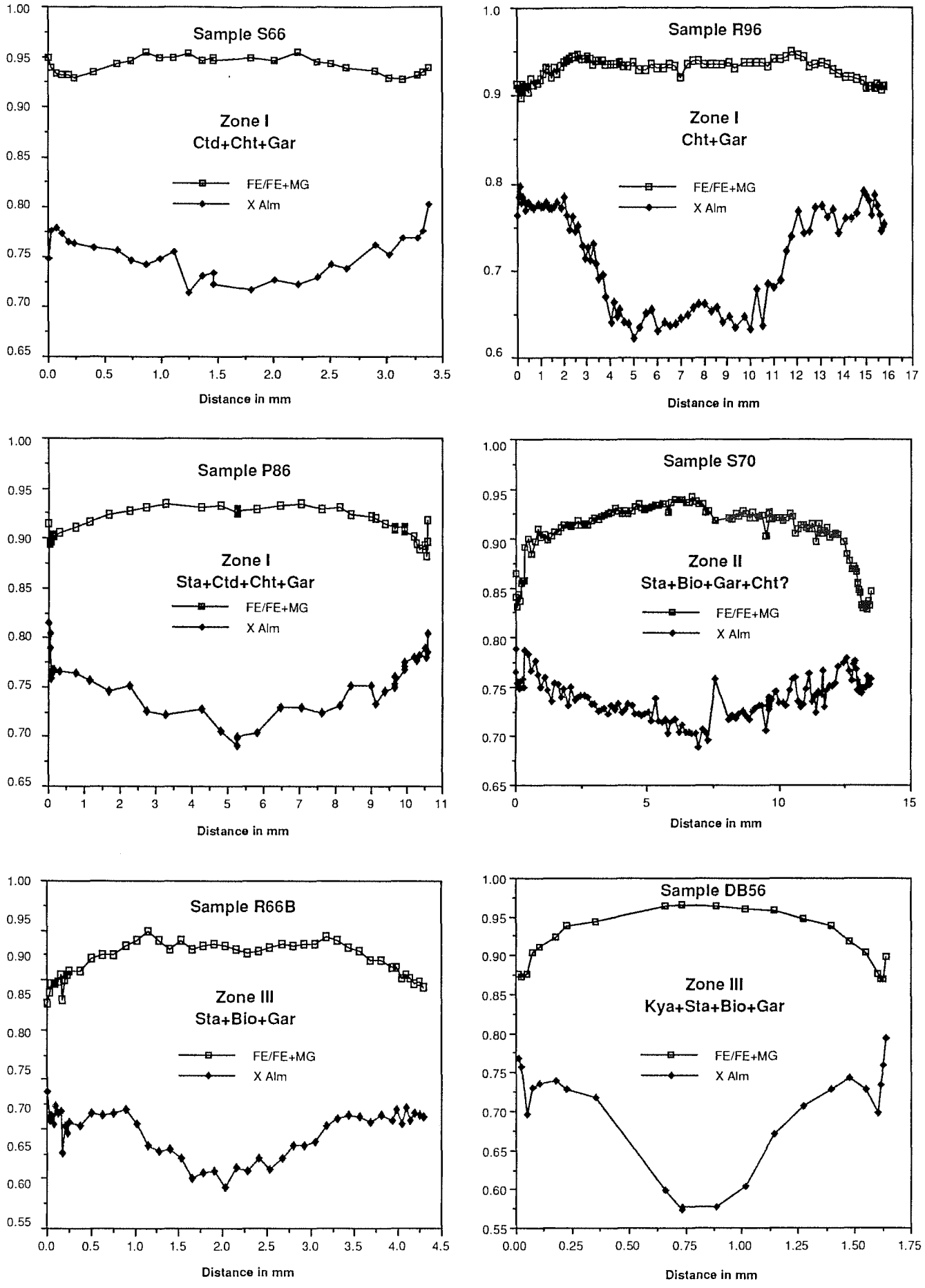


FIGURE 7B. Zoning profiles for the almandine mole fraction and Fe/Fe+Mg ratio from the same garnets as shown in figure 7A.

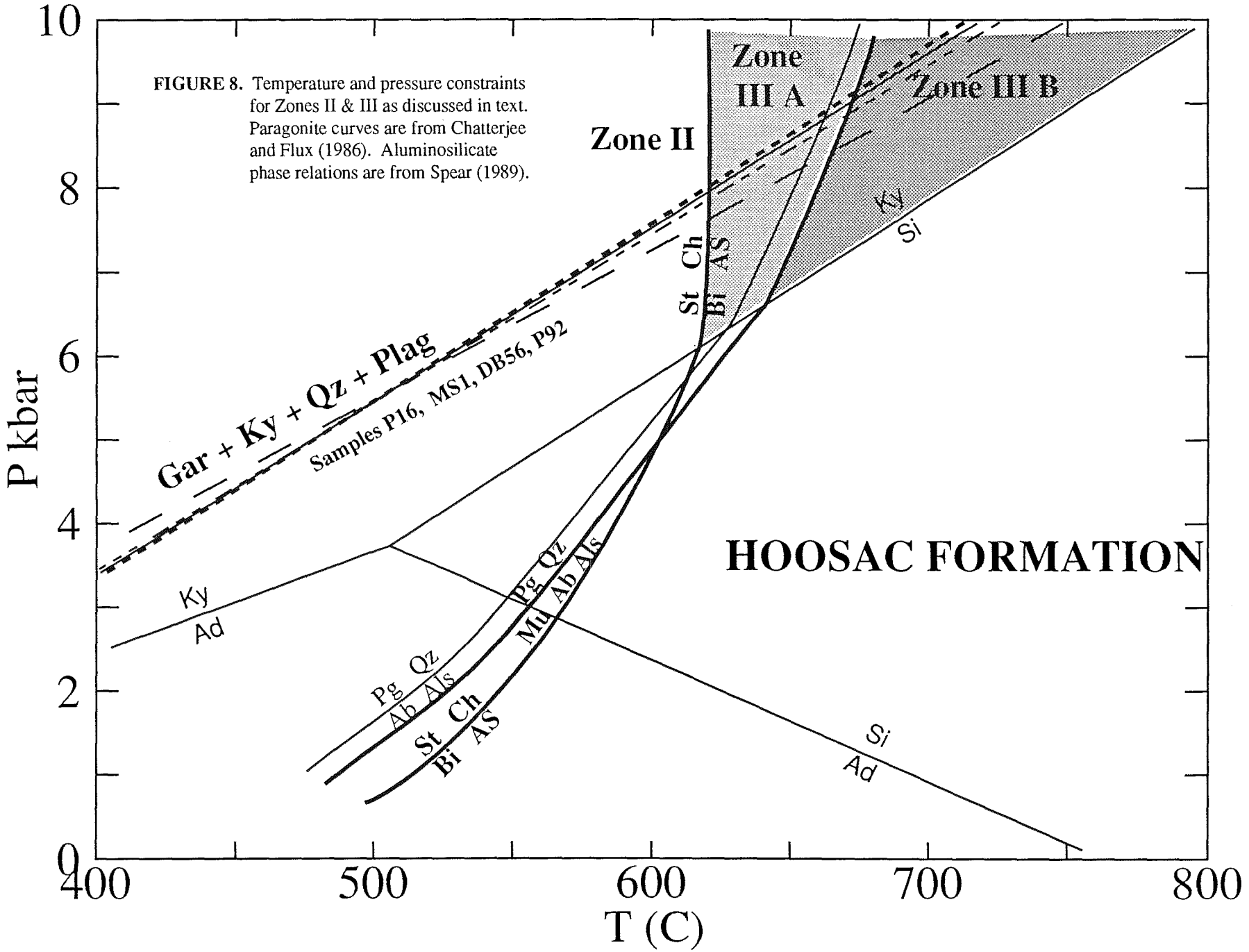


**FIGURE 8.** Temperature and pressure constraints for Zones II & III as discussed in text. Paragonite curves are from Chatterjee and Flux (1986). Aluminosilicate phase relations are from Spear (1989).

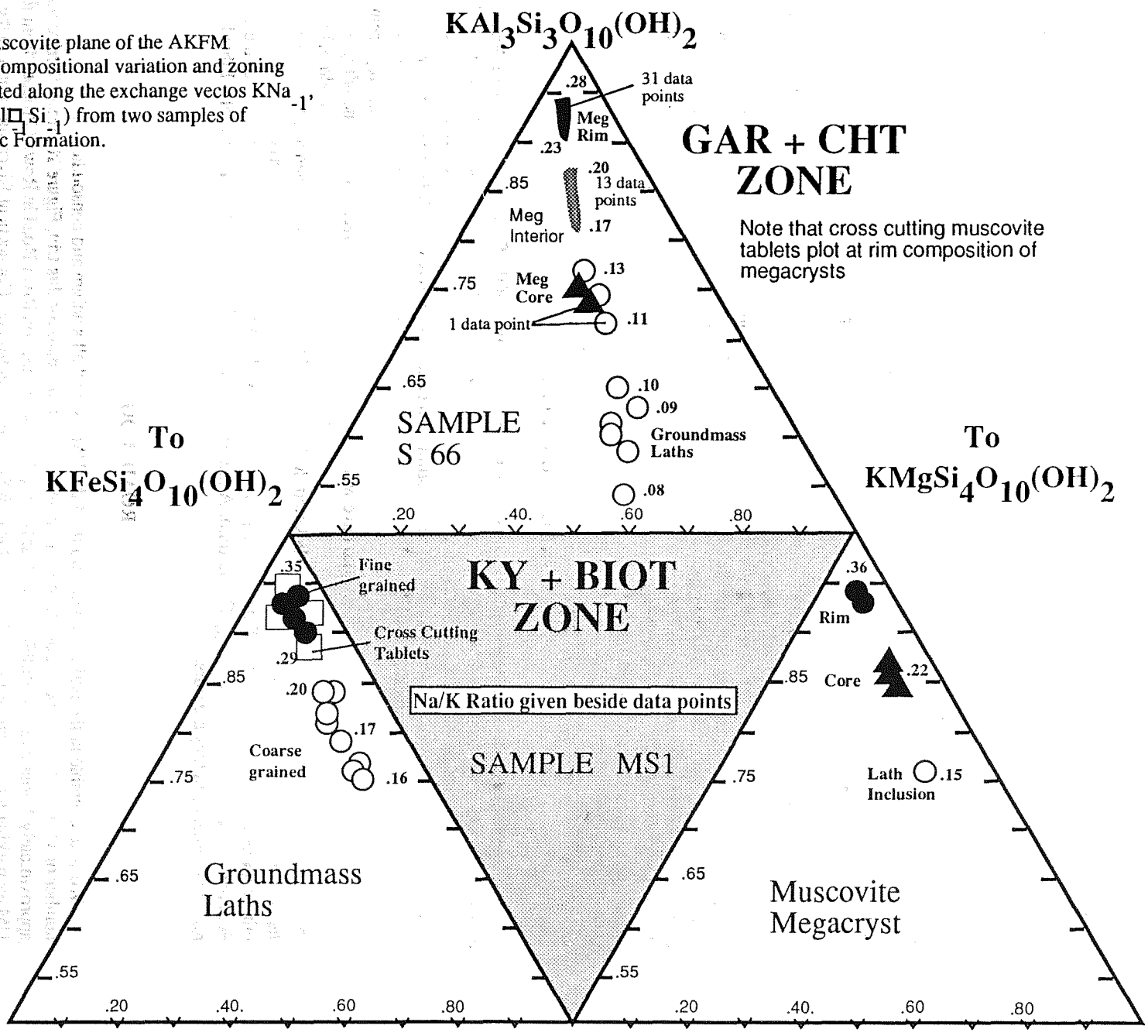


**FIGURE 9.** The muscovite plane of the AKFM tetrahedron and the compositional variation and zoning of muscovite (projected along the exchange vectos $KNa_{-1}$, $Al_2Ti_{-1}Fe_{-1}$, and $KAl\square_{-1}Si_{-1}$) from two samples of high- alumina Hoosac Formation.

There are two obvious models for this polymetamorphism: (1) Acadian overprint of an early Acadian terrane and/or (2) Acadian overprint of a Taconian terrane. Armstrong *et al.* (1992) have shown that the Acadian event recorded higher pressures (~12 kbars, maximum), by some 4 kilobars, than the Taconian event (~ 8 kbars, maximum) to the west of the Berkshire Massif. As pointed out to us by Karabinos (personal communication, 1992) this is in fact consistent with the allochthonous nature of the Hoosac rocks and with their source east of their current location. Because the Taconian subduction zone was probably east-dipping as shown by Stanley and Ratcliff (1985), rocks transported from the east during the Taconian orogeny could in fact reflect deeper burial, as is the apparent case. This of course implies that the Hoosac rocks were undergoing metamorphism or had already been metamorphosed at the time of thrusting. A similar conclusion is required to account for the high pressure "eclogitic" rocks of the Cannon Mountain Formation (Harwood, 1976, 1979; Maggs, 1984; Maggs *et al.*, 1986). However, bear in mind that although we have assembled an elegant mansion of cards, existing data do not preclude the possibility that the early event recorded in these rocks is Acadian, as Karabinos and Laird (1988) carefully point out. It is even possible that the early white mica assemblages in these rocks simply reflect very low temperatures and are recording the compression and early heating portion of a clockwise P-T-t loop in the sense of England and Thompson (1984).

Complex white mica assemblages have been observed so far only in high-alumina rocks from the Hoosac and Rowe Formations. Thus, discrimination among the polymetamorphic models requires additional data regarding the compositions of white mica from the low-alumina rocks of the Hoosac Formation and the Silurian-Devonian rocks from the east limb of the Berkshire massif. In short, the compositional variation of white micas from pelitic schists in western Massachusetts may help to discriminate between Taconian and Acadian events in the same way that complex amphibole assemblages from Cambrian-Ordovician meta-volcanic rocks of northern Vermont (*e.g.* Laird and Albee, 1981) and garnet zoning (Cook and Karabinos, 1988) have been used to look at possible Taconian metamorphic events through Acadian overprinting.

## ACKNOWLEDGEMENTS

Much of the work reported here was fostered by the interest and guidance of the field geologists who mapped the rocks. In particular we are indebted to the late Norm Hatch and the late Leo Hall for their curiosity, generosity and support. We are also thankful for the continued interest and support of Rolfe Stanley, Nick Ratcliff and Steve Norton. In addition we have benefitted from numerous stimulating and provocative exchanges about these rocks with many colleagues, in particular Peter Crowley, Liz Downie, Peewee Guidotti, Paul Karabinos, Peter Robinson, John Rosenfeld, Frank Spear, Walt Trzcienski, and Jim Thompson. Clearly the results summarized here represent, in part, the efforts of many former students who have worked with one or both of us over the years including, but not limited to Neena Bashir, Linda Boop, Aaron Britt, Deen Bryan, Dan Cohan, Jim Crowley, Joel Davidow, Liz Dibble, Carolyn Goeldner, Mark Handy, Don Hickmott, Mike Hudec, Sheila Hudson, Mark Sander, Mark Wick. Jackie Newberry cheerfully assisted with the preparation of the manuscript and Dick Stedman was turning out new thin sections from stop 9 yesterday. Financial support has been provided by NSF Grant 78-03636 to JTC, the Keck Foundation through the Keck Geology Consortium, Amherst College, and Smith College. In spite of all this help, mistakes remain and they are ours.

## ROAD LOG

We will assemble in the parking lot due north of the UMass football stadium and consolidate into a minimum number of vehicles for an 8:00 am departure. As shown on the route map for this trip, Figure 10, we will drive for approximately 1 hour and reassemble at the Bear Swamp Visitors' Center on River Road in Rowe. Proceed from the UMass parking lot west and north on MA 116 through Sunderland, across the Connecticut River, and north to the entrance on I-91 in South Deerfield. Take I-91 north to Greenfield and exit to follow MA 2 some 17 miles west to Charlemont. 1.5 miles west of the junction of MA 8A (south) and MA 2 in Charlemont, turn right (north) on Zoar Road. A rest area on the left and a blue sign for the Yankee Atomic Vistor Center signal the turn. Follow the main road 10.1 miles north through Monroe Bridge to the Bear Swamp Visitors' Center and its restrooms. From the parking lot at the Visitors' Center we will proceed north 3.6 miles to our first stop.

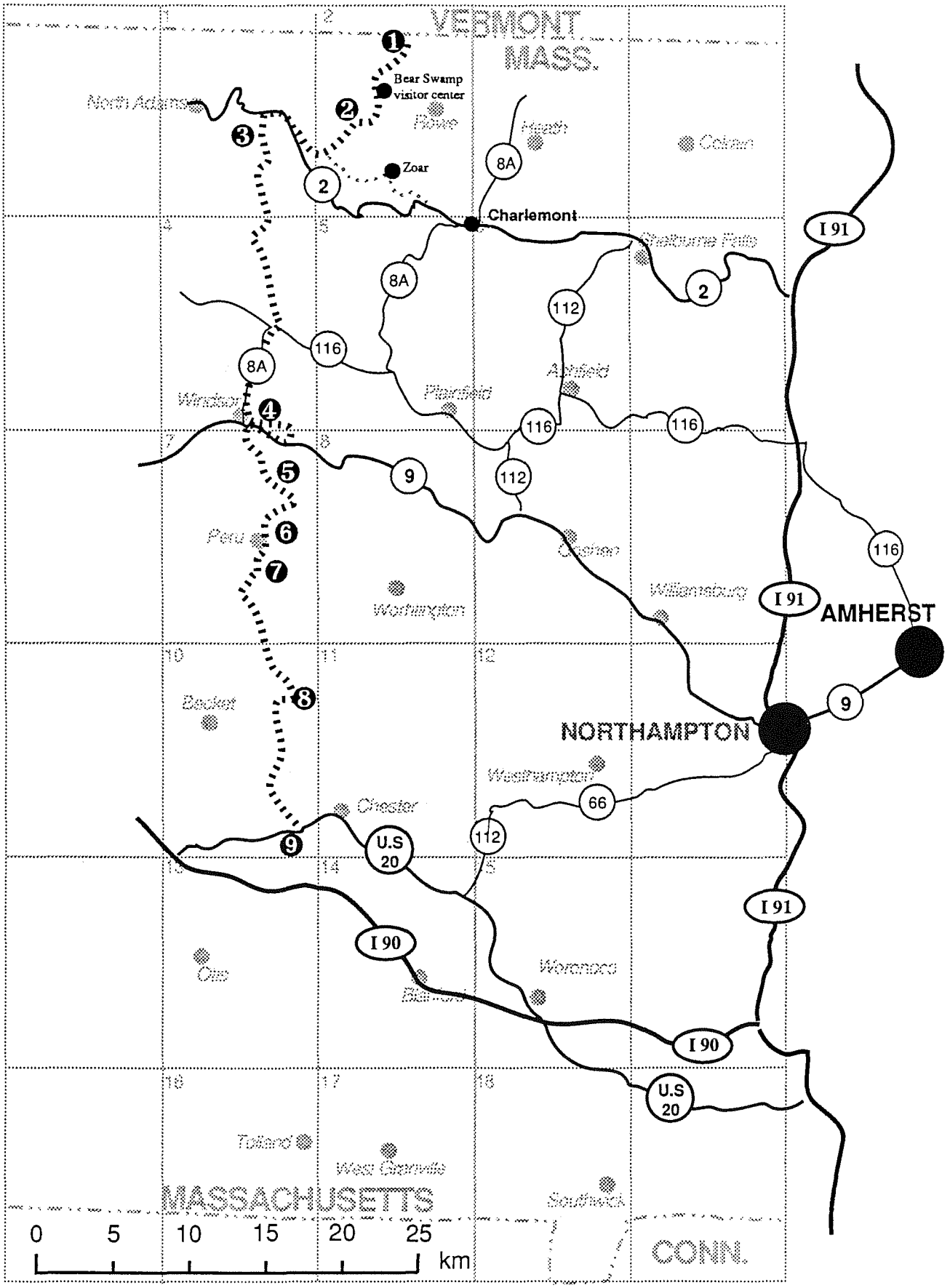


**FIGURE 10.** Field trip route and stop locations. Towns shown are for 7 1/2 minute quadrangles as indicated.

**Mileage:**

0.0 Road log begins at the pull-off on the west side of River Road, 0.4 miles north of turnoff to the dam for the now shut down Yankee Atomic Power Plant. Turn around and park heading south.

**STOP 1. BIG GARNET SCHISTS OF THE HOOSAC FORMATION** (30 MINUTES) (Rowe Quadrangle). The outcrops on the hillside to the west, under the power lines, consist of Cht+Gar+Mus+Qtz schists of the Hoosac formation that locally contain dark green, almost black, layers rich in chloritoid. The chloritoid-bearing rocks also contain paragonite. None of the seven samples from this outcrop examined petrographically contained biotite. Common accessory minerals include tourmaline, epidote/clinozoisite, rutile and or ilmenite, and locally magnetite. This is locality P21 on Figures 5 and 6. The garnets in some layers are very coarse-grained, despite the relatively low metamorphic grade indicated by the mineral assemblage. Moreover, the textural unconformity in these garnets can be seen with a hand lens as sharp rims on inclusion-riddled cores. Common garnet core inclusions are chloritoid, chlorite, very fine-grained rutile, quartz and less commonly white mica and/or epidote-group minerals. As is the case at Gassetts and other localities in S.E. Vermont, one can commonly distinguish two size fractions of white mica. The coarser-grained version that can be resolved into discrete grains is typically muscovite, whereas the much finer-grained variety tends to be paragonite.

0.0 Drive south on River Road.

0.4 On the left (east) is a road to the Yankee Atomic dam. The 186 MW Yankee Atomic Power Plant operated from 1960 until 1991. It was the third commercial nuclear power plant commissioned in the United States and the first in New England. Although this relatively small facility had an excellent safety record, the utilities that own it decided, amid controversy over the possible embrittlement of its steel confinement dome, that it would not be economic to continue operation and to seek the new license required when their original license expires in 1997. Decomissioning plans are being formulated now, but are not likely to be approved until the mid-1990's. Actual dismantling of the plant is unlikely to begin until at least 2000. One holdup on the process is the stock of highly radioactive spent fuel rods that is stored on site awaiting the completion of a national high-level waste repository.

1.0 Town center of Monroe Bridge. Continue south on River Road.

2.7 Monroe/Florida town boundary (yellow sign).

3.6 Bear Swamp Visitors' Center on left. The Bear Swamp Pumped-Storage Hydroelectric Station was built as a companion facility to the Yankee Atomic Power Plant. It is not easy or efficient to start and stop nuclear reactors, so the standard operating procedure is to run them continuously at full power. However, demand for electric power is episodic, matching the daily rhythms of people. When more power is available than is necessary to meet demand (usually at night), the Bear Swamp Station uses the excess power to pump Deerfield River water up to the Bear Swamp Reservoir. When demand for electric power exceeds supply (usually during the day), Bear Swamp Reservoir water is permitted to flow back through the pumps, turning them into electric generators, and down to the Deerfield River. The flow of the Deerfield River is, therefore, episodic and high flow rates can be expected for a few hours each day -- to the delight of white-water enthusiasts.

5.1 Park on right shoulder next to blasted cliffs, as far off the main road as possible, directly across from the gate to Lower Reservoir and the road to Bear Swamp Pump station. **Watch for falling rocks!**

**STOP 2. GARNET SCHISTS OF THE ROWE FORMATION** (30 MINUTES) (Rowe Quadrangle) This is locality P37 on Figures 5 and 6. The vertical surfaces of this blasted roadcut and the next roadcut to the south (locality P38) contain a variety of Rowe Formation lithologies that include a much finer-grained version of the aluminous mineral assemblages seen in the Hoosac Formation at Stop 1: Ctd+Cht+Gar+Mus+Par+Qtz phyllites. Also occurring at these outcrops are Gar+Cht+Mus+Qtz+Pla phyllites. Biotite or paragonite are common additions to the latter assemblage. Biotite and paragonite generally do not occur in the same rocks in this terrane until much higher grade. Additional rocks of interest from these outcrops are amphibole-bearing Gar-Cht-Mus-Qtz phyllites.

5.1 Continue south on River Road.

6.7 Crossing over the Boston and Maine Railroad tracks that lead to the Hoosac Tunnel (100 meters west). This tunnel runs for an amazing 7.6 km beneath the Hoosac Range. It was built at enormous expense,

including nearly 200 human lives, between 1851 and 1875.

7.5 Whitcomb Hill Road. Turn right and follow this newly repaved road past fresh Rowe schist outcrops, past Church Road, past Monroe Road to a T-intersection with MA 2.

10.0 MA 2. Turn right and proceed west 1.8 miles west.

10.4 Whitcomb summit.

11.3 Olson Road (sign to Monroe).

11.8 Turn left onto the second road to the south after (west of) Olson Road. The first road to the south is a very sharp turn (120°). The second road south (Phelps Road) is a 90° turn and is unmarked, except by a stop sign. Proceed south 0.5 miles to where the road crosses the Cold River.

12.3 Just prior to the bridge pull off the road and park, on either side, as best you can. Walk back (north) along the road until you are upstream of the small waterfall (~50 meters) and then proceed west (left) to the outcrops creating the waterfall.

**STOP 3. "THE BEST GARNET-BEARING HOOSAC LOCALITY IN THE NORTH ADAMS QUADRANGLE."** (25 MINUTES) This locality (S66 of Figures 5 and 6), along with most of the other localities we have sampled in this area, was suggested by Nick Ratcliffe, who has mapped in the North Adams Quadrangle. The outcrops of Hoosac Formation along this stream are on strike with rocks of similar composition mapped as nearly-continuous lenses of high-alumina garnet schist south through the Windsor, Peru, and Becket quadrangles by Norton (1976 and references therein). A variety of mineral assemblages can be found at this location, ranging from "normal" Bio-Pla-Gar-Mus-Qtz phyllites through Gar-Cht-Pla-Pgt-Mus-Qtz phyllites to high-alumina Cht-Gar-Ctd-Qtz-Pgt-Mus phyllites. The biotite-bearing varieties generally contain very small garnets compared to the more aluminous bulk compositions. Of some interest is that many Bio+Pla-bearing Hoosac rocks contain garnets that have inclusions of chloritoid in their cores. This outcrop is also the home of some of the most strongly zoned muscovite "Megacrysts" yet analyzed in this terrane. Crosscutting muscovites here have cores with paragonite contents of ~10% whereas the discontinuous rims have paragonite contents of close to 25%!.

12.3 Continue south on Phelps Road.

12.4 Shaft Road, turn left.

13.3 The central air shaft of the Hoosac Tunnel for the Boston and Maine railroad.

14.0 Intersection. Turn right following Shaft Road and the brown sign pointing to Savoy State Forest.

14.6 Entering Savoy Mountain State Forest (brown sign).

15.3 North Pond picnic and swimming area on right. Restrooms in the summer.

16.9 Sharp left turn with (rough!) main road (now Burnett Road).

17.4 New State Road. The intersection is just past a large cleared area on the left (snowmobile parking). Turn right and continue south, into the Windsor quadrangle.

17.6 Tannery Road on your left.

18.3 Parking area on the right for Burnett Pond.

19.0 T-intersection with Adams Road. Turn left and head east.

19.1 Intersection. Turn sharply right with the pavement and follow Center Road south to MA 116.

22.0 T-intersection with MA 116. Turn right and proceed 0.5 miles east to junction with Rt. 8A.

22.4 This road cut was Stop 5 of Norton's (1975b) NEIGC trip.

22.5 Junction of MA 116 with MA 8A. Turn left on MA 8A and proceed south 4.3 miles to Windsor.

26.9 Intersection with MA 9. Turn left and drive east 0.6 mile.

27.5 Savoy Hollow Road. Turn left onto Savoy Hollow Road and drive north 0.2 mile.

27.7 Intersection with Shaw Road (sign on left for Cemetery Road). Turn right and drive east 0.8 miles to the height of land.

28.5 Stop and park at the top of the hill on this right-of-way through Notchview Reservation, owned by the Trustees of Reservations.

**STOP 4. GARNET SCHISTS AT THE NOTCHVIEW RESERVATION.** (30 MINUTES) (Windsor Quadrangle). Despite our drive south of some 16 miles, the large outcrops in the woods on both sides of the road are still in the "Garnet Zone" and contain the same range of assemblages as our last stop. Grain size, especially of the garnets is coarser and reminiscent of garnets from the same assemblages 25 miles to the north at Stop 1. Hence, for all practical purposes the metamorphic grade has not changed significantly since our first stop

today!! Here we are between localities R94 to the north and R96 to the south as shown on Figures 5 and 6. Locality R94 is of particular interest as we have verified the occurrence of margarite with paragonite and muscovite (see Figure 6) in the matrix assemblage of this Ctd-Cht-Gar schist. We have also found, with the electron microprobe, a very minor amount of plagioclase (~An30) in this rock. The very fine-grained nature of the margarite coupled with the minor amount of matrix plagioclase makes their relationship unclear, but margarite has been commonly reported as an inclusion in garnets from Gassetts-like schists (*e.g.* Thompson *et al.*, 1986).

28.5 Continue east on Shaw Road 1.7 miles to its intersection with High Hill Street.
30.2 Intersection with High Hill Street. Turn right and proceed 0.4 miles to MA 9.
30.6 Junction with MA 9. Turn right and head west 3.1 miles to the Peru Road turnoff in Windsor.
32.8 Official entrance to Notchview Reservation on the right. Over 3000 acres of woods, fields, and trails make this property a great destination for hiking (see Brady and White, 1992) and especially cross-country skiing.
33.7 Intersection with Peru road (just east of the intersection of MA 8A north). Turn left and proceed south into the Peru quadrangle. The road becomes Beauman Road when you enter the town of Peru. Avoid this road during inclement weather as it degrades and has a tendency to become quite muddy and slippery when wet.
36.3 Major power line crosses road.
37.3 Stop and park cars as directed and then walk northeast through the upper end of a small clearing ~0.5 km to the steep northwest-trending ledges.

**STOP 5. STAUROLITE (FINALLY)-CHLORITOID-CHLORITE-GARNET SCHISTS FROM DEEP PERU.** (60 MINUTES) (Peru quadrangle). The spectacular, corrugated, coarse-grained nature of the garnet and quartz ribbons in the schists that make up the ledges obscure the fine-grained staurolite that occurs in the matrix with chloritoid, chlorite, paragonite, and muscovite. Although staurolite rarely exceeds 2% in the mode of these rocks, it has been identified in many of the 17 samples from this outcrop studied petrographically. We have not found kyanite in these high-alumina rocks (hence the large number of thin-sections) nor have we found the AFM assemblage St+Cht+Gar. This is locality P86 on Figures 5 and 6. The garnets here are sufficiently coarse that the textural unconformities are particularly well-developed in many samples from these outcrops.

37.3 Continue 0.7 miles southeast on Beauman Road to its intersection with East Windsor Road.
38.0 Y-junction with East Windsor Road. Turn right and follow the pavement to intersection with North Road.
39.9 Y-junction with North Road. Bear left and follow North Road south 1.1 miles to MA 143 in Peru.
41.0 Intersection with MA 143. Turn left and drive east on MA 143 for 0.7 miles.
41.7 Park on right (south) shoulder and proceed north, across road, and gently downhill to low outcrops in the Peru Wildlife Management Area.

**STOP 6. BIOTITE IN HIGH-ALUMINA COMPOSITIONS FROM DEEPER PERU.** (30 MINUTES) (Peru Quadrangle). These outcrops of Hoosac Formation contain St+Bio in Gar-Mus-Pgt-Qtz schists. Here the biotite is intergrown with coarse muscovite and it is difficult to see in most hand samples due to the small modal amount, typically less than 5%. This is locality S70 of Figures (5 and 6). Despite extensive searching, we have as yet to find rocks between Stop 5 and here that contain the AFM assemblage St+Cht+Gar. This assemblage is characteristic of Zone IC, which is "missing" in Peru -- probably due to the sparse outcrop in the critical area (note the swamp to our north).

41.7 After turning vehicles around, proceed west on Rt 143, back to the main intersection in the center of Peru.
42.4 North/South Road Intersection. Turn left and follow South Road 1.0 mile to where the road makes a sharp right turn.
43.4 Entrance to the Dorothy Frances Rice Wildlife Sanctuary. Turn left (almost straight) onto Rice Road and proceed 0.3 mile to end of road.
43.7 Park as appropriate. Dorothy Francis Rice was a 1923 graduate of Smith College who spent summers here as a child. When she died of tuberculosis in 1925, her family decided to create this sanctuary in her memory.

**STOP 7. HIGH-ALUMINA BIG GARNET SCHISTS FROM DEEPEST PERU.** (60 MINUTES) (Peru Quadrangle). A color-coded map of the trails in the sanctuary is posted at the parking area. We

will follow the pink trail clockwise beginning with the path that leads south, just to the left of the stone well. The pink markers are hard to see because they are positioned for a counterclockwise loop, but the path is well-worn. Follow the trail up and along the south slope of French Hill, ~0.75 km, to the point where the trail turns sharply downhill and south. Proceed a few dozen meters further south to a west-directed overlook. Now proceed due east into the brush. Outcrops in this area are coarse-grained Bio-St-Gar-Mus-Pgt schists. Plagioclase occurs in many of these rocks. It is likely stabilized with paragonite by calcium. Of particular interest here is that many of the samples from the east side of French Hill contain a fine-grained, fabric-aligned, commonly-deformed generation of staurolite and even kyanite, in rare cases. However, only staurolite occurs here as second-generation, coarser-grained, crosscutting porphyroblasts in a biotite-poor matrix dominated by white micas. These outcrops are just south of the P63 locality and near the R76 locality of Figures 5 and 6. Moreover, the intercalated sequence of gneisses and schists we passed along the trail are discussed as Stop 6 by Norton (1975b). Return due west to the "pink trail" and retrace the route back to the vehicles. This trail is one of 50 in Massachusetts recommended for pleasure hiking by Brady and White (1992).

43.7 Drive west back along Rice Road to South Road.
44.1 Intersection with South Road. Turn left and follow South Road 1.5 miles south to Middlefield Road.
45.1 Turn sharply right and then turn sharply left staying on South Road.
45.6 Intersection with Middlefield Road. Bear left and follow Middlefield Road 5.2 miles to the village of Middlefield.
47.8 Middlefield Town line.
50.8 Middlefield Village. Park on the right at the Town Hall and walk due west to low pavement outcrops behind the playground and in the woods beyond.

**STOP 8. THE KYANITE + BIOTITE ZONE AND THE MIDDLEFIELD THRUST.** (30 MINUTES) (Becket quadrangle). This is location MS1 and R66 of Figures 5 and 6. These outcrops contain coarse-grained Bio+St assemblages in coarse-grained Gar-Mus-Pgt-Qtz schists that are locally intercalated with less common but relatively coarse-grained, kyanite-bearing versions of this same assemblage. Hence we are at Ky+Bio grade. This is the highest grade occurrence of paragonite that we have so far confirmed. To the south, the aluminous rocks of Zone IIIB contain Ky+Pla and are apparently above the terminal stability of paragonite. As shown on Figure 7, sample MS1 also contains zoned muscovite megacrysts, similar to those observed in lower-grade rocks. This locality was Stop 8 of Norton (1975b) and is the home of the Middlefield thrust.

50.8 Turn around and drive north 0.1 mile.
50.9 Town Hill Road. Turn left and proceed downhill west and then south on Town Hill Road 7.7 miles through Bancroft to US 20 in Becket.
52.0 Sharp bend left.
54.5 Arched Penn Central railroad underpass just before a bridge over the Westfield River to the village of Bancroft.
57.7 T-junction. Follow the main (Bancroft) road left.
58.6 Intersection with US 20. Turn sharply right and drive 0.4 mile uphill to Quarry Road.
59.0 Quarry Road. Park or turn around here being watchful of fast-moving vehicles on US 20. Walk downhill along MA 20 to newly-created (1992) outcrops.

**STOP 9. VERY COARSE-GRAINED KYANITE-STAUROLITE-GARNET-SCHISTS.** (20 MINUTES) (Becket Quadrangle). This is our last stop, just north of locality P92 on Figures 5 and 6. The rocks contain particularly fresh, coarse-grained aluminous minerals (for all of you with little faith) and are especially rich in kyanite. Just as at our first stop, however, the inclusion assemblage in the cores of the garnets is Ctd+Cht. There is staurolite and even kyanite in the outer part of some of these garnets. We have even encountered rare biotite inclusions in the rims of a few garnets from locality P92.

**TO RETURN TO AMHERST:** As shown on figure 10, proceed east on US 20 for 8.9 miles through Chester to Huntington where you should turn left on MA 112 north and east across the river. Follow MA 112 for 3.4 miles to the junction with MA 66. Turn right onto MA 66, which will take you in 13.4 miles to MA 9 in Northampton. Turn right and follow MA 9 east to Amherst and the Banquet!

## REFERENCES


Abbott, R.N., Jr. , 1979, Polymetamorphism in the Silurio-Devonian Goshen Formation of western Massachusetts, Geological Society of America Bulletin, v. 90, Pt. 1, p. 871-876.

Albee, A.L., 1972, Metamorphism of pelitic schists: Reaction relations of chloritoid and staurolite: Geological Society of America Bulletin, v. 83, p. 3249-3268.

Armstrong, T.R., Tracy, R.J. and Hames, W.E., 1992, Contrasting styles of Taconian, Eastern Acadian and Western Acadian metamorphism, central and western New England, Journal of Metamorphic Geology, v. 10, p. 415-426.

Bashir, Neena, 1989, Grain size variation in garnet in pelitic schists from the Chester Dome, Vermont. B.A. Honors Thesis, Amherst, MA, Amherst College, 178p.

Brady, J.B. and White, Brian, 1992, Fifty hikes in Massachusetts. Backcountry Publications, Woodstock, Vermont, 223p.

Britt, Aaron, 1980, Geology of the Prospect Hill area, west Granville Quadrangle, western Massachusetts: B.A. honors thesis, Amherst, Mass., Amherst College, 81p.

Bryan, Deen, 1978, Petrography and petrology of the Mt. Cobble area, Beckett Quadrangle, western Massachusetts: B.A. honors thesis, Amherst, Mass., Amherst College, 62p.

Chamberlain, C.P. and Conrad, ME., 1991, Oxygen isotope zoning in garnet. Science, v. 254, p. 403-406.

Cheney J.T., 1986, Disequilibrium muscovite compositions from single crystals and samples as indicators of P-T paths in western Massachusetts. Geological Society of America, Abstracts with Programs, v. 18, p. 8.

Cheney, J.T., 1980, Chloritoid through sillimanite zone metamorphism of high-alumina pelites from the Hoosac Formation, western Massachusetts: Geological Society of America Abstracts with Programs, v. 12, no. 7, p. 401.

Cheney, J.T., Sander, M., and Bryan, D., 1980, Influence of garnet composition on the distribution of metamorphic isograds in the east limb of the Berkshire anticlinorium: Geological Society of America Abstracts with Programs, v. 12, no. 2, p. 28.

Cheney, J.T. and Guidotti, C.V., 1979, Muscovite-plagioclase equilibrium in sillimanite + quartz bearing metapelites, Puzzle Mountain area, N.W. Maine: American Journal of Science, v. 279, p. 411-434.

Christensen, J.N., Rosenfeld, J.L. and DePaolo, D.J., 1989, Rates of tectonometamorphic processes from rubidium and strontium isotopes in garnet. Science, v. 244, p. 1465-1469.

Conner, T.G., 1979, Geology of the Tob Hill Area, B.A. honors thesis, Amherst, Mass., Amherst College, 44p.

Cook, S.M., 1988, Polymetamorphism in southeastern Vermont: B.A. Honors Thesis, Williams College, Williamstown, MA, 86p.

Cook, S.M., and Karabinos, Paul, 1988, Polymetamorphic isograds in southeastern Vermont: Geological Society of America, Abstracts with Programs, v. 20, p. 13.

Crowley, James, 1989, Metamorphic petrology of high aluminous schists in the Star Hill Sigmoid, B.A. Honors Thesis, Amherst, MA, Amherst College, 225p.

Davidow, Joel, 1989, Geochemical characterization of the Acadian orogeny in the Chester Dome region, southern Vermont, B.A. Honors Thesis, Amherst, MA, Amherst College, 101p.

Dempster, T.J., 1992, Zoning and recrystallization of phengitic micas: implications for metamorphic equilibration, Contributions to Mineral Petrology, v. 109, p. 526-537.

Dibble, E.C., 1981, A new position for the kyanite isograd: Mineralogy and Petrology of metasediments from the Goshen Quadrangle, western Massachusetts: B.A. honors thesis, Amherst, Mass., Amherst College, 74p.

Downie, E.A., 1982, Structure and metamorphism in the Cavendish area, north end of the Chester dome, southeastern Vermont. Ph.D. thesis, Harvard University, Cambridge, Massachusetts, 291p.

Downie, E.A. , 1980, Prograde and retrograde reaction sequence and condition of metamorphism, north end of the Chester dome, southeastern Vermont. Geological Society of America, Abstracts with Programs, v. 12, p. 415.

Downie, E.A., 1979, Metamorphic phase relations at staurolite and staurolite-kyanite grade in the Goshen Formation, western Massachusetts: Geological Society of America Abstracts with Programs, v. 11, no. 1, p. 18.

Downie, E.A., 1975, Mg/Fe distribution in coexisting garnet, biotite and staurolite from the kyanite and staurolite grade rocks of the Goshen Formation, western Massachusetts: B.A. honors thesis, Northampton, Mass., Smith College, 76p.

Emerson, B.K., 1917, Geology of Massachusetts and Rhode Island, U.S. Geological Survey Bulletin 597, 289p.
Emerson, B.K., 1899, The geology of eastern Berkshire county, Massachusetts, U.S. Geological Survey Bulletin 159, 139p.
Emerson, B.K., 1898, Geology of old Hampshire county, Massachusetts, comprising Franklin, Hampshire and Hampden counties, U.S. Geological Survey Monolith, 29, 790p.
Emerson, B.K., 1892, Outlines on geology of the Green Mountain region in Massachusetts: Hawley sheet, descriptive text: U.S. Geological Survey Geologic Atlas, Hawley sheet, 3p.
England, P.C., and Thompson, A.B., 1984, Pressure-temperature-time paths of regional metamorphism I. Heat transfer during the evolution of thickened continental crust, Journal of Petrology, v. 25, p. 894-928.
Giaramita, M.J. and Day, H.W., 1991, Buffering in the assemblage staurolite-aluminum silicate-biotite-garnet-chlorite, Journal of Metamorphic Geology, v. 9, p. 363-378.
Goeldner, C., 1983, Deformation and metamorphism in the Connecticut Valley thermal trough, Colrain-Bernardston quadrangles, western MA. B.A. Honors Thesis, Amherst, MA, Amherst College, 130p.
Guidotti, C.V., 1984, Micas in metamorphic rocks. In: Bailey , S.W. (ed). Micas. Reviews in mineralogy, Mineralogical Society of America, v. 13, p. 357-467.
Guidotti, C.V., 1974, Transition from staurolite to sillimanite zone, Rangeley Quadrangle, Maine, Geological Society of America Bulletin, v. 85, p. 475-490.
Hall, L.M., and Robinson, Peter, 1982, Stratigraphic-tectonic subdivisions of southern New England. In St.-Julien, P., and Béland, J., eds., Major structural zones and faults of the northern Appalachians, p. 15-41. The Geological Association of Canada Special Paper Number 24.
Hames, W.E., Sutter, J.F. & Ratcliffe, N.M., 1990, Evidence for a 15 Ma duration of the Taconian collisional episode. Geological Society of America Abstracts with Programs, v. 22, p. A368.
Hames, W.E., Tracy, R.J., Ratcliffe, N.M. & Sutter, J.F., 1991, Petrologic, structural and geochronologic characteristics of the Acadian metamorphic overprint on the Taconian zone in part of southwestern New England. American Journal of Science, v. 291, p. 887-913.
Handy, M., 1980, Geology of the Russell Mountain area, Woronoco quadrangle, western Massachusetts: B.A. honors thesis, Amherst, Mass., Amherst College, 108p.
Harwood, D.S., 1979, Bedrock geologic map of the South Sandisfield quadrangle, Massachusetts and Connecticut, U.S. Geological Survey Quadrangle Map GQ-1519, scale 1:24,000.
Harwood, D.S., 1979b, Bedrock geologic map of the Norfolk Quadrangle, Connecticut. U.S. Geological Survey Quadrangle Map GQ-1520, scale 1:24,000.
Harwood, D.S., 1976, Garnet + clinopyroxene plagioclase symplectites in southern Berkshire Massif (abs.). Geological Society of America, Abstracts with Programs, v. 8, no. 1, p. 149.
Hatch, N.L., Jr. 1988, Some revisions to the stratigraphy and structure of the Connecticut Valley trough, eastern Vermont. American Journal of Science, v. 288, p. 1041-1060.
Hatch, N.L., Jr., 1975, Tectonic, metamorphic, and intrusive history of part of the east side of the Berkshire massif, Massachusetts: U.S. Geological Survey Bulletin 888-B, p. 51-62.
Hatch, N.L., Jr., Robinson, P. and Stanley, R.S., 1988, Stratigraphy of the Connecticut Valley belt. In: The Bedrock Geology of Massachusetts (ed. Hatch, N.L., Jr.), U.S. Geological Survey Professional Paper 1366, B1-B34.
Hatch, N.L., Jr., and Stanley, R.S., 1988, Post-Taconian structural geology of the Rowe-Hawley Zone and the Connecticut Valley belt west of the Mesozoic Basins, in: U.S. Geological Survey Professional Paper 1366-A-C, p. C1-C36.
Hatch, N.L. and Stanley, R.S., 1976, Geologic map of the Blanford quadrangle, Hampden and Hampshire counties, Massachusetts: U.S. Geological Survey Geology Quadrangle Map GQ-1312.
Hatch, N.L., Jr., Zen, E-an, Goldsmith, Richard, Ratcliffe, N.M., Robinson, Peter, Stanley, R.S. and Wones, D.R., 1984, Lithotectonic assemblages as portrayed on the new bedrock geologic map of Massachusetts: American Journal of Science, v. 284, p. 1026-1034.
Hickmott, Donald, 1982, The mineralogy and petrology of the Betts Mine area, Plainfield, MA. B.A. honors thesis, Amherst, MA, Amherst College, 105p.
Hickmott, Donald and Spear, F.S., 1992, (in press), Major- and trace-element zoning in garnets from calcareous pelites in the NW Shelburne Falls Quadrangle, Massachusetts: Garnet growth histories in retrograded rocks. Journal of Petrology, v. 33, part 5.

Hodges, K.V., and Crowley, P.D., 1985, Error estimation and empirical geothermobarometry of pelitic systems, American Mineralogist, v. 70, p. 702-709.

Hudson, Sheila, 1983, The transition from garnet zone to staurolite-kyanite zone in the pelitic schists of the Ashfield quadrangle, western MA., B.A. honors thesis, Amherst , MA, Amherst College, 98p.

Karabinos, Paul, 1985, Garnet and staurolite producing reactions in a chlorite-chloritoid schist. Contributions to Mineralogy and Petrology, v. 90, p. 262-275.

Karabinos, Paul, 1984, Polymetamorphic garnet zoning from southeastern Vermont. American Journal of Science, v. 284, p. 1008-1025.

Karabinos, Paul and Laird, Jo, 1988, Structure and metamorphism from Jamaica to the Athens Dome, Vermont, In: Guidebook for Field Trips in Southwestern New Hampshire, Southeastern Vermont, and North-central Massachusetts (ed. Bothner, W.A.), Keene State College, Keene, New Hampshire, p. 281-292.

Laird, Jo, Kunk, M.J., Boxwell, M.A. and Bothner, W.A., 1991, Acadian metamorphism about the Taconian Line, southeastern Vermont, Geological Society of America, Abstracts with Programs, v. 23, p. 55.

Laird, Jo and Albee, A.L., 1981, Pressure, temperature and time indicators in mafic schists; their application to reconstructing the polymetamorphic history of Vermont: American Journal of Science, v. 281, p. 127-175.

Laird, Jo, Lanphere, M.L., and Albee, A.L., 1984, Distribution of Ordovicianand Devonian metamorphism in mafic and pelitic schists from northern Vermont, American Journal of Science, v. 284, p. 376-413.

Maggs, William, 1984, Garnet+clinopyroxene-plagioclase symplectite lenses in the Berkshire massif, northwest Connecticut. B.A. Honors Thesis, Amherst, MA, Amherst, College, 141p.

Maggs, W.W., Cheney, J.T., and Spear, F.S., 1986, Probable rtrograded eclogites in the Berkshire massif. Geological Society of America, Abstracts with Programs, v. 18, p. A32.

Massonne , H-J, Schreyer, W., 1987, Phengite geobarometry based on the limiting assemblage with K-feldspar, phlogopite, and quartz. Contributions to Mineral Petrology, v. 96, p. 212-224.

Norton, S.A., 1976. Hoosac formation (early Cambrian or older) on the east limb of the Berkshire massif, western Massachusetts: Geological Society of America Memoir 148, p. 357-371.

Norton, S.A., 1975b, Stratigraphic and structural relationships along the east side of the Berkshire massif, Massachusetts, in New England Intercollegiate Geological Conference, 67th annual meeting, Great Barrington, Mass., Oct. 10-12, 1975, Guidebook for field trips in western Massachusetts, northern Connecticut and adjacent areas of New York: City College of City University of New York, Department of Earth and Planetary Sciences, p. 122-143.

Norton, S.A., 1975a, Chronology of Paleozoic tectonic and thermal metamorphic events in Ordovician, Cambrian, and Precambrian rocks at the north end of the Berkshire massif: U.S. Geological Survey Professional Paper 888-B, p. 21-31.

Pferd, J.W., 1978, Bedrock geology fo the Colrain quadrangle: Amherst, Mass., University of Massachusetts, Ph.D. thesis, 277 p.

Pumpelly, R., Wolff, J.F., and Dale, T.N., 1894, Geology of the Green Mountains in Massachusetts: U.S. Geological Survey Monolith 23, 192 p.

Ratcliffe, N.M., 1988, Along strike variations in tectonics of the Taconian orogen: Geological Society of America Abstracts with Programs, v. 20, no. 1, p. 64.

Ratcliffe, N.M. Burton, W.C., Sutter, J.F. and Mukasa, S.A., 1988, Stratigraphic, structural geology, and thermochronology of the northern Berkshire massif and southern Green Mountains. Part II—Internal structure of the Green Mountain massif and Wilmington Dome. In: Guidebook for Field Trips in Southwestern New Hampshire, Southeastern Vermont, and North-central Massachusetts (ed. Bothner, W.A.), Keene State College, Keene, New Hampshire, p. 126-135.

Robinson, Peter, 1986, Introduction, in: Field Trip Guidebook: Regional Metamorphism and Metamorphic Phase Relations in Northwestern and Central New England, International Mineralogy Association 14th General Meeting, p. 1-10.

Rosenfeld, J.L., 1968, Garnet rotation due to major Paleozoic deformations in southeastern Vermont, in: Studies of Appalachian Geology, Northern and Maritime, E-an Zen, W.S. White, J.B. Hadley, and J.B. Thompson, Jr. (eds.), Intersicence, New York, p. 185-202.

Rosenfeld, J.L., Christensen, J.N. and DePaolo, D.J., 1988, Snowball garnets revisited, southeast Vermont, in: W.A. Bothner (ed.), New England Intercollegiate Geological Conference, 80th Annual Meeting, Keene, N.H.

Sander, Mark, 1977, The staurolite isograd of the east flank of the Berkshire anticlinorium, western Massachusetts, B.A. honors thesis, Amherst, Mass., Amherst College, 76 p.

Spear, F.S., 1991, On the interpretation of peak metamorphic temperatures in light of garnet diffusion during cooling. Journal of Metamorphic Geology, v. 9, p. 379-388.

Spear, F.S., 1989, Petrologic determination of metamorphic pressure-temperature-time paths, in: Metamorphic Pressure-Temperature-Time Paths (eds. Spear, F.S. & Peacock, S.M.), p. 1-55. Short Course in Geology, 7, American Geophysical Union.

Spear, F.S. & Cheney, J.T., 1989, A petrogenetic grid for pelitic schists in the system $SiO_2$-$Al_2O_3$-FeO-MgO-$K_2O$-$H_2O$, Contributions to Mineralogy and Petrology, v. 101, p. 149-164.

Spear, F.S. & Selverstone, J., 1983, Quantitative P-T paths from zoned minerals: theory and tectonic applications, Contributions to Mineralogy and Petrology, v. 83, p. 348-357.

Stanley, R.S., and Hatch, N.L., Jr., 1988, The pre-Silurian geology of the Rowe-Hawley zone, in: Hatch, N.L., Jr., ed., The bedrock geology of Massachusetts: U.S. Geological Survey Professional Paper 1366, p. A1-A39.

Stanley, R.S. & Ratcliffe, N.M., 1985, Tectonic synthesis of the Taconian orogeny in western New England. Geological Society of America Bulletin, v. 96, p. 1227-1250.

Sutter, J.F. and Hatch, N.L., 1985, Timing of metamorphism in the Rowe-Hawley Zone, western Massachusetts. Geological Society of America, Abstracts with Programs, v. 177, p. 65.

Sutter, J.F., Ratcliffe, N.M. and Mukasa, S.B., 1985, 40Ar/39Ar and K-Ar data bearing on the metamorphic and tectonic history of western New England. Geological Society of America Bulletin, v. 96, p. 123-136.

Thompson, A.B., Tracy, R.J., Lyttle, P.T. and Thompson, J.B., Jr., 1977a, Prograde reaction histories deduced from compositional zonation and mineral inclusions in garnet from Gassetts schist, Vermont: American Journal of Science, v. 277, p. 1152-1167.

Thompson, A.B., Lyttle, P.T. and Thompson, J.B., Jr., 1977, 1977b, Mineral reactions and A-Na-K and A-F-M facies types in the Gassetts schist, Vermont: American Journal of Science, v. 277, p. 1124-1151.

Thompson, J.B., Jr., 1979, The tschermak substitution and reactions in pelitic schists: (In Russian) in V.A. Zharikov, V.I. Forarev, and V.A. Korikovshii, eds.: Problems of Physiochemical Petrology, 1, 146-159. Academy of Sciences of the U.S.S.R., Moscow (translated by J.B. Thompson, Jr.).

Thompson, J.B., Jr., Cheney, J.T., and Robinson, Peter, 1986, Metamorphism of the east flank of the Green Mountain massif and Chester dome, in: Field Trip Guidebook: Regional Metamorphism and Metamorphic Phase Relations in Northwestern and Central New England, International Mineralogy Association 14th General Meeting, p. 94-119.

Tracy, R.J., Robinson, P. and Thompson, A.B., 1976, Garnet composition and zoning in the determination of temperature and pressure of metamorphism, central Massachusetts: American Mineralogist, v. 61, p. 762-775.

Young, E.D. and Rumble, D., III, 1992, (in prep.), Correlated variations in *in situ* 18O/16O and elemental concentrations in metamorphic garnet, Geochim. Cosmochim. Acta (in review).

Zen, E-An (ed.), Goldsmith, R., Ratcliffe, N.M., Robinson, P. and Stanley, R.S. (compilers), 1983, Bedrock Geologic Map of Massachusetts. U.S. Geological Survey, Reston and Commonwealth of Massachusetts, Department of Public Works, Boston.

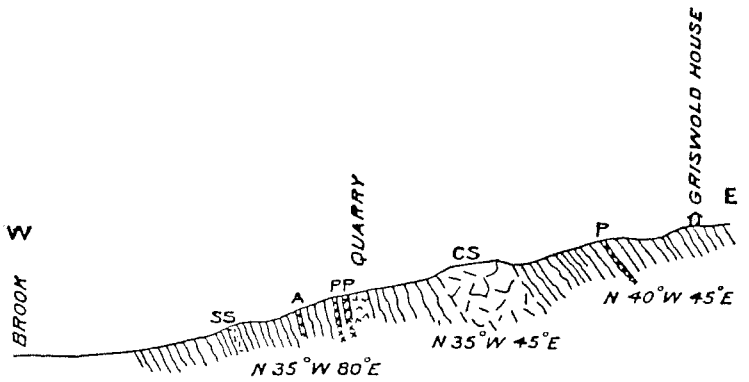


FIG. 5.—Section at Osborn soapstone quarry, Blandford. S S = Sahlite-serpentine; S = Steatite and enstatite-serpentine; O S = Olivine-serpentine; A = Amphibolite; P = Pegmatite; country rock = sericite-schist.

# MESOZOIC PALEOMAGNETISM, AND DIABASE PETROGRAPHY AND GEOCHEMISTRY IN CENTRAL MASSACHUSETTS

by
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## PURPOSE OF TRIP

Paleomagnetic poles from the Newark Supergroup rocks have formed the cornerstone for Apparent Polar Wander Paths (APWP) for North America in the Early Jurassic. New work in Massachusetts and Connecticut has led to questions about the validity of older pole positions and adds new data, that, combined with recently determined Cretaceous data from New England, can be used to construct a revised Jurassic-Cretaceous APWP. The purpose of this trip will be to show participants some of the localities where thermoremanent magnetic signatures were obtained in four suites of diabase intrusions, and where detrital and chemical remanent magnetic signatures have been obtained in Jurassic mudstones and arkoses (Figure 1). In addition there will be an opportunity to discuss the relationships between diabase petrography and geochemistry, and how geochemistry may relate to the Mesozoic tectonic history of the region. Many aspects of the diabase intrusions in central Massachusetts are covered in greater detail in a volume by McEnroe (1989) to which participants are referred.

## GEOLOGIC SETTING

The diabase dikes and sills intrude highly deformed metamorphic rocks of Late Precambrian through Devonian age (Zen et al.,1983) arranged in three north-northeast trending belts, from west to east, the Connecticut Valley synclinorium, the Bronson Hill anticlinorium, and the Merrimack synclinorium. The regional geologic setting is amply covered elsewhere in this guidebook, and the Mesozoic setting particularly by Wise et al. From the point of view of paleomagnetic studies, the important facts are that ductile deformation of the Paleozoic crystalline rocks ended in the Late Paleozoic, and that Mesozoic brittle faulting, although involving major vertical and horizontal movements in the Triassic and Early Jurassic, was virtually over by the Middle Jurassic.

In Massachusetts the eastern margin of the Connecticut Valley synclinorium and the western margin of the Bronson Hill anticlinorium are unconformably overlain by Upper Triassic and Lower Jurassic sedimentary rocks and basaltic volcanics of the Connecticut Valley Mesozoic basins, which are localized mainly along the west side of the Connecticut Valley border fault. This major west-dipping listric normal fault is estimated to have had about 5 km of vertical displacement in northern Massachusetts (Robinson,1986; Robinson et al., 1986) and as much as 10 km near the Connecticut State line. In matching the metamorphic rocks on opposite sides of the listric normal fault, Elbert (1986) suggests a 16 degree rotation of the western block relative to the eastern block. Based on slickensides the net slip was approximately down the dip of the fault (Ashenden, 1973) and the rotation axis essentially horizontal and trending N30E parallel to the fault trace. This appears to be the maximum Mesozoic rotation that has been proposed for any large block of basement rock in this part of New England. Because of the low paleolatitude in the Early Jurassic, resulting in low paleomagnetic inclinations, paleomagnetism does not give a very sensitive method for measuring rotations about horizontal axes during Jurassic faulting. In view of the north-trending rotation axis, which is subparallel to the magnetic declinations of most of the Mesozoic intrusions, this rotation has only a minor effect on paleomagnetic orientations. Knowing that 16° represents an extreme example for rotation of basement rock, it is probable that the Jurassic dikes and sills away from the border fault have not been rotated by an amount that would be significant to paleomagnetic studies, and the Cretaceous intrusions would have been affected even less. The 16° extreme rotation near the border fault, results in the negligible difference in magnetic directions of 2° to 3°. Therefore, all paleomagnetic directions of the diabases are reported in situ.

## PALEOMAGNETISM

### Introduction to Paleomagnetic Studies

The paleomagnetic research on the Jurassic mudstones and arkoses deposited in the Hartford and Deerfield Basins started over a decade ago (Brown, 1979). Many controversial questions are still unanswered about the timing of acquisition of the primary and secondary magnetizations found in the sedimentary rocks. The diabase dikes and sills in west-central Massachusetts were sampled for paleomagnetic and geochemical studies in 1986 and 1987. Until this research there had been surprisingly little paleomagnetic or geochemical work on these intrusions even

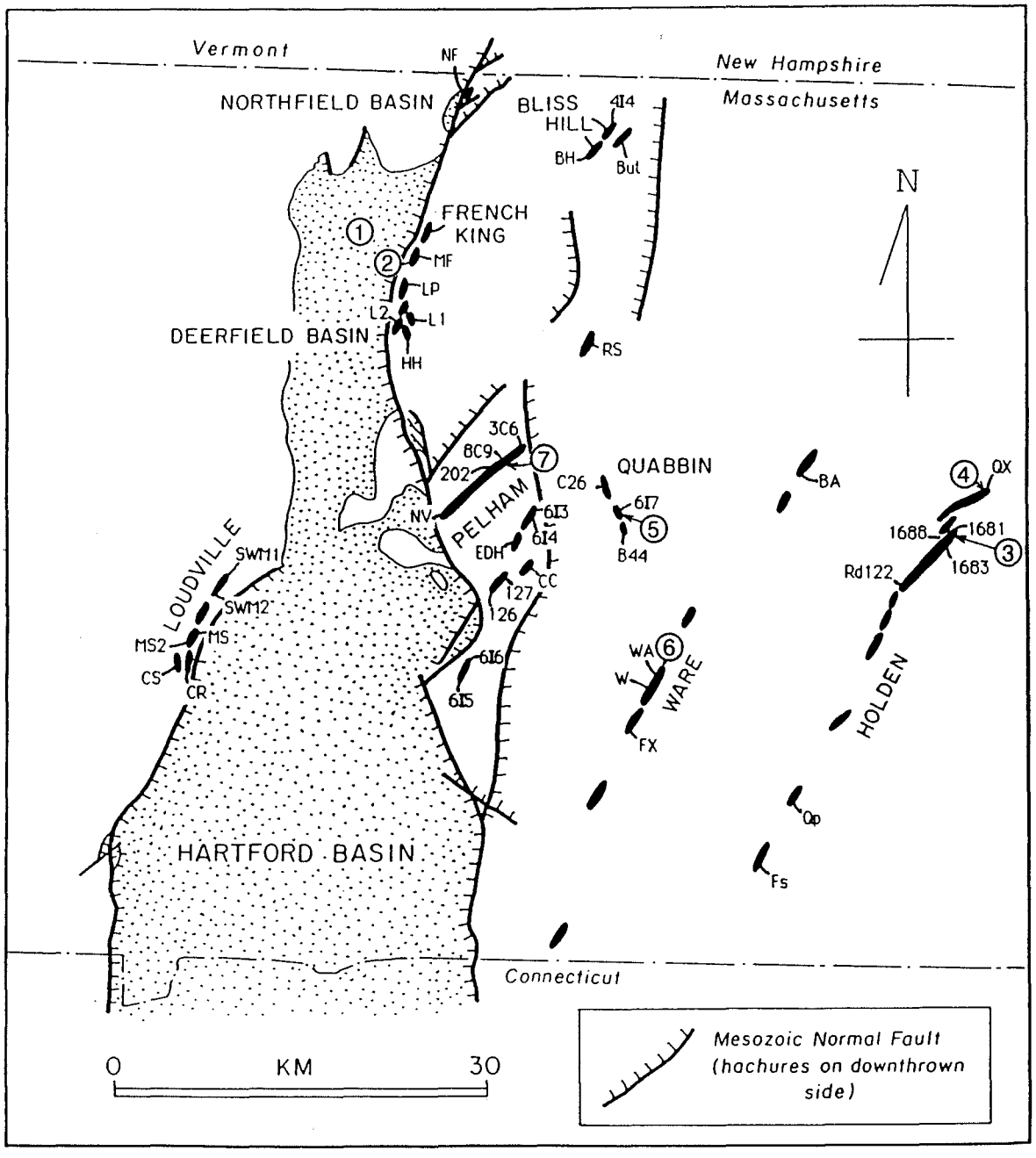



Figure 1. Map of central Massachusetts showing locations of Mesozoic basins, dikes and sills intruded into crystalline rocks, and numbered field-trip stops. Jurassic intrusions fall into three systems: Holden, Pelham-Loudville, and Ware including the French King sill. Cretaceous intrusions are in the Quabbin and Bliss Hill areas. All small intrusions without name labels west and north of the Quabbin group belong to the Pelham-Loudville system. Individual localities are labelled as in McEnroe (1989).

though there are abundant outcrops, albeit few roadcuts. Paleomagnetic sites were located on the basis of the Bedrock Geological Map of Massachusetts (Zen et al., 1983) and detailed field maps and notes kindly supplied by various geologists in the region.

The primary goal of the paleomagnetic part of this research has been a careful and modern reexamination of the mid- to late-Mesozoic segment of the APWP for North America. The accurate definition of apparent polar wander paths is crucial to our understanding of plate tectonics, especially during the Mesozoic when major rifting and

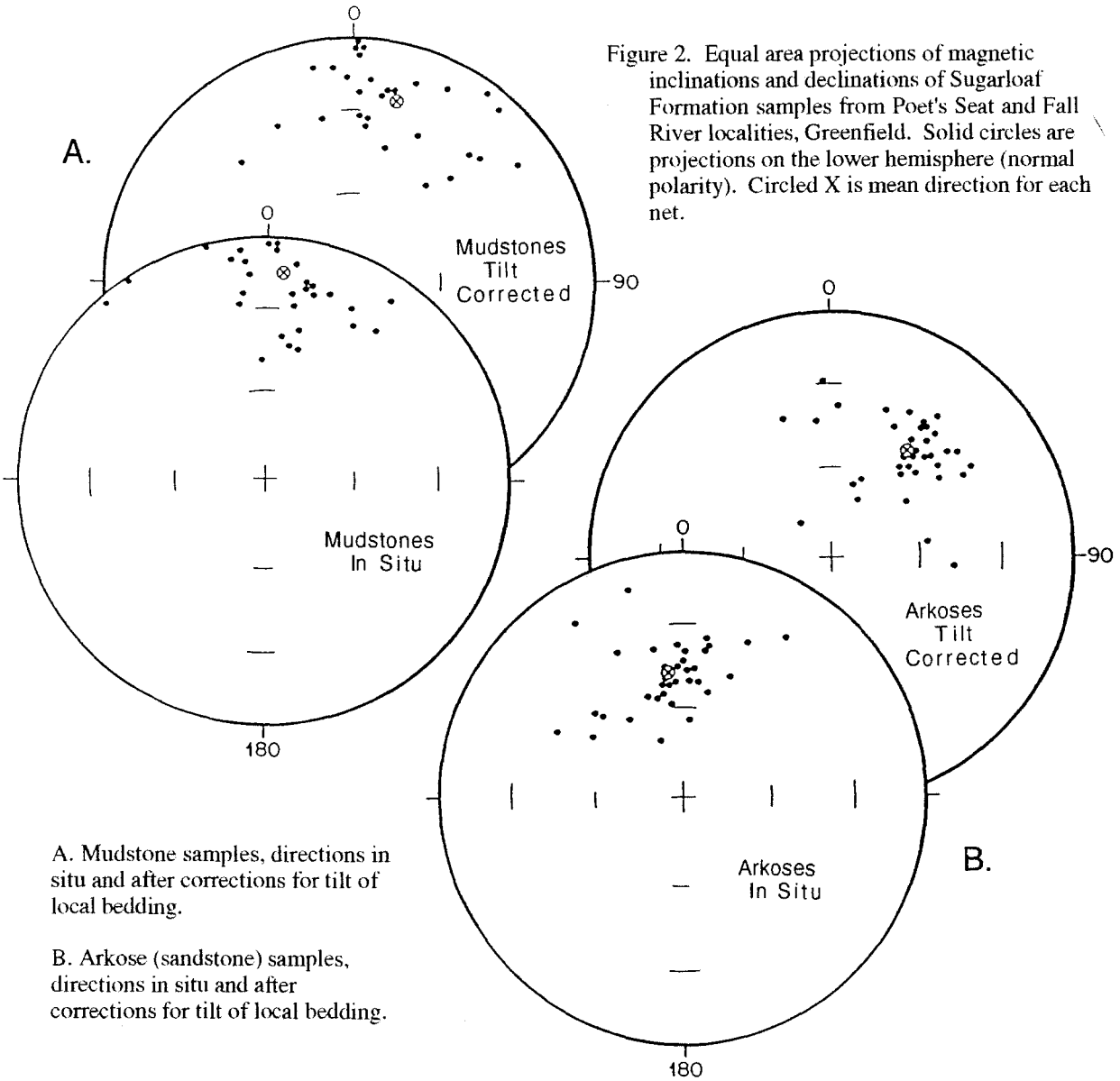



Figure 2. Equal area projections of magnetic inclinations and declinations of Sugarloaf Formation samples from Poet's Seat and Fall River localities, Greenfield. Solid circles are projections on the lower hemisphere (normal polarity). Circled X is mean direction for each net.

A. Mudstone samples, directions in situ and after corrections for tilt of local bedding.

B. Arkose (sandstone) samples, directions in situ and after corrections for tilt of local bedding.

continental reorganization occurred. Unfortunately the Jurassic and Cretaceous portion of the APWP for North America is poorly understood for several reasons. Though there is abundant data for the late Triassic-early Jurassic, it is predominantly derived from two locations, the Colorado Plateau and the east coast Mesozoic basins. Results from units of similar age from the two locations are in poor agreement, and shed suspicion on all the results. In addition, the poles from units in the Newark Supergroup represent complex magnetizations, where the problems of remagnetization, secondary overprinting, and improper age assignments are strongly suspected. In the Cretaceous, the number and quality of reliable paleomagnetic poles from any region of North America are inadequate to detect systematic motion along this portion of the APW path.

**Paleomagnetic Results**

The natural remanent directions (NRM) of the Jurassic sites show considerable variation with intensities ranging from 0.3 to 4.0 A/M. Under alternating field (AF) demagnetization the characteristic magnetic direction for igneous units appears after approximately 50% of the NRM has been removed in these sites, usually between 20 mT and 40 mT. In general the normalized intensity plots of the Jurassic diabases compare well with other Jurassic tholeiitic diabases from eastern North America (Smith, 1987). The reversed Cretaceous Quabbin Reservoir diabases show a different decay curve. These samples increase in intensity at low mT levels, as a soft component is removed that is

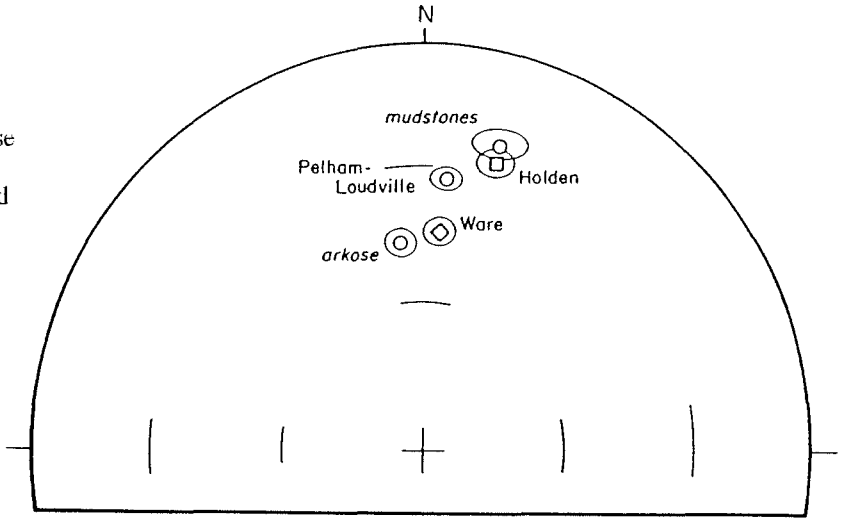



Figure 3. Site mean magnetic directions of central Massachusetts Jurassic diabase systems and Jurassic mudstones and arkoses plotted on an equal area net with a95 cones of confidence. All symbols are projections onto the lower hemisphere.

in the opposite direction from the remanent direction. This soft component is probably related to the present-day normal field. Thermal demagnetization studies yield results similar to the AF studies. The sedimentary rocks have a more complicated magnetization, with two or three components evident with thermal demagnetization. These include a present-day overprint on all samples, a steep normal component younger than tilting of the beds in the arkoses, and a shallow normal direction older than the tilting seen only in the mudstones (Figure 2).

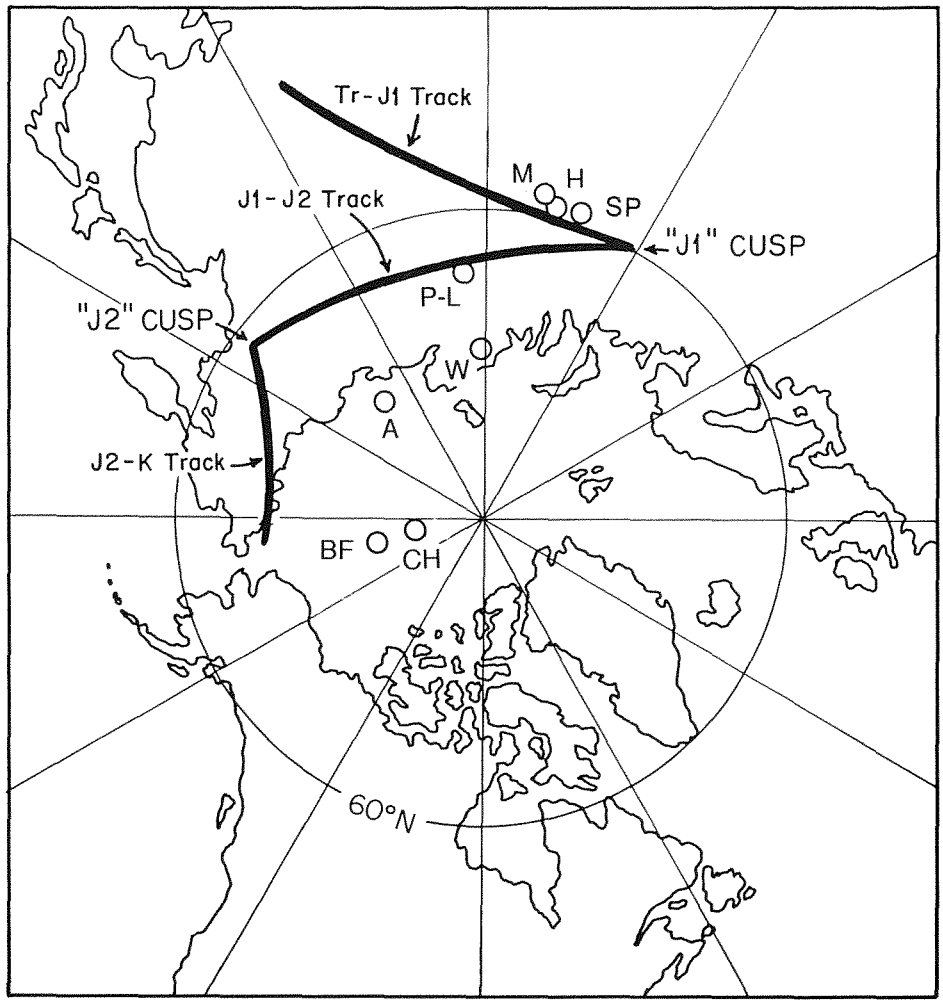
**Magnetic Site Directions**

Magnetic directions for the diabases were calculated by averaging sample directions for each site and then averaging all sites for each of the different chemical systems. When treated this way, each chemical system yields statistically different magnetic directions. The three diabase groups are statistically distinct, with no overlap at the 95% confidence level, indicating that each system recorded a different magnetic direction upon cooling. Mudstones from the uppermost section of the Sugarloaf Formation yield a primary direction that persists to high temperatures of thermal demagnetization, as well as an overprint, removed at lower temperatures, that is also seen as the major magnetization in the coeval sandstones. All mean paleomagnetic directions for the Jurassic systems, both sediments and diabases, are of normal polarity and are shown in Figure 3. These are: Sugarloaf mudstones D=14.0°, I=24.3°, a95=9.1°; Holden system D=15.2°, I=28.1°, a95=3.3°; Pelham-Loudville system D=5.7°, I=32.9°, a95=2.2°, Ware system D=5.4°, I=44.8°, a95=2.3°; Sugarloaf sandstones D = 353.1°, I=47.7°, a95=5.9°. The magnetic directions may be considered as a list ordered with decreasing age. Steeper mean magnetic directions translate into higher paleopole latitudes, and hence younger pole positions reflecting the northwestward motion of the North American plate. The paleomagnetic directions obtained for the Cretaceous diabases are of mixed polarity. The Bliss Hill north site (4I4, normal polarity) and the Quabbin Reservoir group (reversed polarity) show a steeper magnetic inclination which yields a significantly higher paleolatitude than the Jurassic directions.

A very significant unanswered question is: What are the ages of the high -Fe-Ti quartz tholeiite and the low-Ti quartz tholeiite intrusions, which yielded a significant part of the paleomagnetic data base used to calculate the Newark trend N1 and N2 pole positions? These tholeiite types in Massachusetts and in Connecticut (McEnroe, 1992) have significantly distinguishable paleomagnetic poles at the 95% confidence level, indicating significant intervals of time between episodes of magmatic activity.

**Pole Positions and Apparent Polar Wander Paths**

The paleopoles from this study are plotted on Figure 4 and are compared to the Mesozoic apparent polar wander curve for North America by May and Butler (1986). The mean poles for the Sugarloaf mudstones and the Holden group (59°N, 93°E) plot near the Triassic-Jurassic track prior to the J1 cusp. The Pelham-Loudville system, interpreted as intermediate in age (65°N, 95°E), plots near the J1-J2 track, and the Ware system, the youngest of the Jurassic dike systems, has a high-latitude pole (73°N, 91°E), plotting north of the J1-J2 track. The pole from the




(May & Butler 1986)

Figure 4. Mesozoic apparent polar wander path for North America, after May and Butler (1986). Paleopoles determined in this study shown by circles:
M - mudstones;
H - Holden;
SP-Seaward Point, Maine;
P-L - Pelham-Loudville;
W - Ware:
A - Arkoses;
BF - Baffle Dam Island sill of Quabbin group;
CH - Chapman Island sill of Quabbin group.

Sugarloaf sandstones also plots at high latitude (75°N, 132°E). The Massachusetts Cretaceous diabases plot well north and east of the J2-K track.

As a comparison, other Mesozoic pole positions for the Appalachians are listed in Table 4.4 of McEnroe (1989). Smith and Noltimier (1979) presented paleomagnetic data on the Newark Supergroup intrusive rocks and basalt flows in Connecticut and related this to $^{40}Ar/^{39}Ar$ age determinations (Sutter and Smith, 1979). They calculated two pole positions for the early Jurassic, now known as the Newark trend N1 and N2 poles, , suggesting two separate periods of igneous activity at 190 m.y. and 175 m.y. These poles have been widely used in constraining the Early Jurassic portion of Mesozoic polar wander paths for North America.

Field and geochemical work on diabase dikes and sills, and basalt flows in and near the Connecticut Valley Mesozoic basins suggests that some of the dikes are feeders to the flows (Philpotts and Martello, 1986). This casts doubt on the grouping of Connecticut data used by Smith and Noltimier (1979) to calculate the Newark N1 and N2 poles. They calculated N1 pole (63°N, 83°E) based on the sills and flows. All the narrow dikes were considered to post-date the volcanic flows and in Connecticut, Pennsylvania and Maryland were grouped together to calculate the younger N2 pole (65°N , 103°). Limited alternating field demagnetization treatment, on average only to 10mT, and oversimplified structural corrections for diabases intruded into the Mesozoic basins may also have led to erroneous N 1 and N2 pole positions. Such corrections are minimal for the diabase intrusions in crystalline bedrock outside of the basins, though locally there are Mesozoic faults.

Smith (1976) averaged the southern extensions of the Holden, Pelham-Loudville and Ware systems together, leading to an erroneously young pole position for the Holden and Pelham-Loudville systems and an older pole position for the Ware system. Sutter and Smith (1979) determined an age for the Buttress (Ware) dike of 176.0 ± 3.8 by the $^{40}Ar/^{39}Ar$ method, and Smith and Noltimier (1979) inferred this age for Higganum and Bridgeport systems. Philpotts and Martello (1986) correlated the Massachusetts diabases with those in Connecticut, showing the identity of the Holden with the Higganum, Pelham with Bridgeport and Ware with Buttress..

Unfortunately, a direct comparison of the pole positions from the present study with those determined for the dikes in Connecticut in earlier studies cannot be made. Most of the work of DeBoer (1967) on intrusive diabases along the length of the Appalachians is not considered valid due to his excessive AF demagnetization (Smith and Noltimier 1979) but the results from Connecticut Valley flows (DeBoer,1968) are still used. The earlier work by DuBois et al. (1957) and Bowker (1960) were largely reconnaissance surveys, with inappropriately low levels of demagnetization.

The virtual geomagnetic pole (VGP) positions determined for the Cretaceous intrusions in Massachusetts are displaced farther north in latitude than previously accepted pole positions for North America. These VGP's are in excellent agreement with recently determined poles on nine well dated Cretaceous intrusions in New Hampshire, Vermont and Maine (McEnroe, 1991; McEnroe and Brown, 1992). These indicate a smoother and more northerly apparent polar wander path for North America from the Jurassic into the Early Cretaceous than presently accepted. Previous studies of the apparent polar wander path for North America by Irving and Irving (1982), Gordon et al. (1984) and May and Butler (1986) all indicate the critical need for additional paleomagnetic data for the Cretaceous. The new data from the Massachusetts diabase intrusions indicate a more northerly latitude for the North America plate at 120 m.y. than previously thought and point to the critical need of additional paleomagnetic research from this period in areas such as New England that have been tectonically inactive after the Mesozoic. These results indicate that the apparent polar wander path for North America in the Jurassic - Cretaceous needs to be reevaluated.

## DIABASE PETROGRAPHY

### Introduction

The Mesozoic dike systems are distinct in their petrographic character, based on examination of sixty-eight thin sections, twenty-four polished-thin sections and nine polished cores. Plagioclase compositions were determined using the Michel-Levy method or estimated on the basis of CIPW norms calculated from chemical analyses (McEnroe, 1989, Tables 3.1 to 3.6). Crystal form, exsolution, oxidation and alteration of the opaque minerals were noted. The classification system of Haggerty (1976) was used to describe the extent of oxidation exsolution of magnetite-ülvospinel solid solution members. Some petrographic features of the four dike systems to be visited on this field trip are given below

### Holden System

The seventeen samples studied petrographically from the Holden system included eleven thin sections, two from chilled margins and nine from well crystallized interiors; and six polished cores, two from chilled margins and six from well crystallized interiors. The average chilled margins contain 5% normally zoned plagioclase phenocrysts $An_{59-64}$, 1.3 to 1.6 mm long; 3% Mg-rich orthopyroxene phenocrysts (2V=-80-85), 0.8 to 1.5 mm long, altered up to 20% along the edges and in cracks; and 8% clinopyroxene phenocrysts, including both augite and pigeonite of 0.8mm average length, commonly occurring in aggregates of crystals. The matrix is composed of 36% plagioclase, 38% clinopyroxene, red-brown biotite, 10% skeletal to euhedral grains of titanomagnetite classified from C1 to C2, and discrete grains of ilmenite. The sulfides are pyrite and chalcopyrite. The pyrite occurs as euhedral grains, filigree, and as an intergrowth with titanomagnetite.

The interiors of the diabases are well crystallized and in some samples it is difficult to distinguish phenocrysts from matrix minerals. The major minerals include 41-50% normally zoned plagioclase, $An_{58-72}$, ranging from 8.0 to 2.0 mm long; 2-10% of Mg-rich orthopyroxene (2V=-80-85), 1.4 to 2.2 mm long; 22-42% clinopyroxene, including augite and pigeonite, 0.8-2.0 mm long and 3% of red-brown biotite rimmed by green biotite, up to 0.7 mm long. The titanomagnetites have abundant "exsolution" lamellae of ilmenite and are classified as C2 to C3 with increasing ilmenite. These show greater alteration than in the chilled samples, to Fe-Ti amorphous oxides, to titanomaghemite, and possibly to sphene in some samples. Discrete grains of ilmenite are common. Pyrite varies in texture from droplets and filigree to euhedral grains. Chalcopyrite occurs in trace amounts, and pentlandite is

present in one section. Micrographic intergrowths of alkali feldspar and quartz are common and trace amounts of apatite and calcite are present in most samples. Chlorite is present as an alteration product in the matrix of 80% of the sections. Orthopyroxene is altered around the edges and about 15% of the plagioclase is sericitized.

The dikes of the Holden system contain certain petrographic features of particular interest to petrogenetic interpretations. In the interiors there are grains of pigeonite rimmed by augite, probably indicating a peritectic relationship. These rims contain exsolution lamellae, that are probably pigeonite, and are indicative of relatively slow cooling, also suggested by local augite lamellae in orthopyroxenes. The orthopyroxenes are rimmed by a more Fe-rich mineral, in most cases replaced by a sheet silicate, that.could have been olivine produced on orthopyroxene as it reacted during decompression as described by Philpotts and Reichenbach (1985). The clinopyroxene phenocrysts in the chilled margins have ragged edges which may indicate that they were also out of equilibrium with the liquid at the time of intrusion. There are late-stage pyrites that surround the titanomagnetites, and pyrite and titanomagnetite intergrowths. The local micrographic intergrowths of quartz and alkali feldspar, common to all the interior samples, probably represent interstitial residual liquid of granitic composition.

**Pelham-Loudville System**

Eight thin sections from chilled margins and twenty-two well crystallized interior samples were examined in transmitted light. Eighteen specimens were examined in reflected light. The chilled margins contain 5-10% phenocrysts of plagioclase, $An_{55-70}$, 0.8 to 2.0 mm long; 3-5% microphenocrysts of olivine (2V=-80-85), $Fo_{75-85}$, 0.16 to 0.25mm in diameter; and 2-7% clinopyroxene phenocrysts including augite and pigeonite that range in length from 0.6 to 4mm. The plagioclase phenocrysts are commonly normally zoned and approximately 3% of them have interiors that have a coalescing finger-like texture that evolves to a single grain, suggesting crystallization under supersaturated conditions according to Lofgren (1974). Dungan and Rhodes (1978), alternatively, suggest that this plagioclase texture may be produced by resorption during magma mixing. Inclusions of pyroxenes (0.04 mm) are common in the large plagioclase grains. The microphenocrysts of olivine are rimmed by augite and pigeonite. The widths of the reaction rims on the olivine crystals increase from the contact inward toward the interiors of the diabases until all the olivine is reacted out.

The matrix is composed of 39-47% plagioclase (0.1-0.2mm); 40-45% clinopyroxene, augite (2V=+30-50) and pigeonite (2V=+0-10) average size 0.12mm; 10-20% titanomagnetite (C1-C3), ranging from skeletal arrangements with herring-bone texture to euhedral grains; 3% fine-grained titanomagnetite dust; 1% sulfides including pyrite as sulfide droplets and euhedral crystals, and chalcopyrite; less than 1% biotite; and discrete grains of ilmenite. Trace amounts of apatite, and zircon are also present. Quartz xenoliths with reaction rims of pyroxene, ilmenite, and micrographic K-feldspar and quartz are common in most thin sections. Secondary alteration is less than 10%, including sericitized plagioclase that is more abundant in this system than in any of the others. 1-3% Fe-chlorite, possibly replacing olivine is common. In some sections the titanomagnetites have been altered to an amorphous Fe-Ti oxide. Trace amounts of secondary calcite, Fe and Mg-chlorite, biotite and hematite occur in the matrix. In some sections there are thin veins or dikelets containing red-brown high-temperature biotite and olive-green hornblende.

The diabases of the Pelham-Loudville system fall into two geochemical classifications, olivine and quartz tholeiites, though only only small amounts of either olivine or quartz, respectively, are present in the norms. This is confirmed petrographically in that microphenocrysts of olivine, where present, are out of equilibrium with the liquid, as demonstrated by the pyroxene rims. The Pelham-Loudville diabases differ from the other central Massachusetts diabases in that augite is the abundant clinopyroxene.

**Ware System**

Three thin sections from chilled margins, and eight from well crystallized interiors were studied. They are relatively fresh with less than 5% alteration. Two chilled margins, and one interior sample were examined in reflected light. The chilled margins contain 5% phenocrysts of plagioclase, $An_{70}$, 1.2 to 2.2mm long, and 2% microphenocrysts of olivine (2V=-80-85), $Fo_{65-75}$, up to 0.16mm in diameter. The plagioclase phenocrysts are commonly normally zoned. Olivines are rimmed by clinopyroxene, indicating a reaction relationship between olivine and liquid. About 3% of the clinopyroxene phenocrysts, including both augite and pigeonite, are present, with an average length of 0.6 mm. The matrix is composed of 35% plagioclase, 40% clinopyroxene, including both augite and pigeonite, 10% devitrified glass, 2% straw-colored oxidized glass and 10% fine-grained herring-bone skeletal titanomagnetite. Because the titanomagnetite lacks exsolution lamellae, it is classified as Cl.

Due to the interlocking texture in the interior diabase specimens, it is impossible to determine whether the clinopyroxenes and plagioclase grains are phenocrysts or matrix crystals. These rocks consist of 45 to 50% plagioclase, 0.2 to 1.2mm long; 40 to 50% clinopyroxene, including both augite (2V=+40-50) and pigeonite (2V=+0-5), 0.4 to 1.7 mm long; and 3% titanomagnetite, 0.6mm in diameter. Most titanomagnetite grains have abundant lamellae of ilmenite and are classified as C3. Superimposed on the titanomagnetites is a high-temperature deuteric oxidation, in which the grains were modified to an amorphous Fe-Ti oxide. It is considered unlikely that this high-temperature alteration was later than the original cooling of the diabases, because little alteration of the plagioclases and pyroxenes was observed. There are rare discrete grains of ilmenite. Sulfides are common, typically making up 1% of the rock. Pyrite varies in texture from droplets and filigree to euhedral crystals. Chalcopyrite is present in trace amounts. Red-brown biotite, commonly rimmed by green biotite, is present in amounts up to 1% in the matrix. Associated with this are trace amounts of apatite, quartz, K-feldspar and calcite. Ten percent of the plagioclase is sericitized and secondary Fe-chlorite and an oxychlorite are present in amounts less than 1%. Minor cracks are filled with late hematite. Pigeonite is preferentially altered, typically along the circular cracks believed to have been produced by structural contraction during the C to P inversion, whereas augite is less commonly altered. Locally an unidentified greenish-brown sheet silicate has replaced euhedral grains that may have been olivine.

The Ware system diabases are classified as tholeiites (Yoder and Tilley 1962).based on petrographic characteristics, including the reaction relationship between olivine microphenocrysts and liquid, and later interstitial quartz and alkali feldspar, A distinguishing feature of this system, as compared to other systems is the marked abundance of pigeonite relative to augite.

**Cretaceous Quabbin Reservoir Group**

The chilled margins contain 1-2% phenocrysts of plagioclase, $An_{51-62}$, 0.6 to 1.6 mm long; and 2-5% partially altered skeletal olivines, $Fo_{65-75}$ (2V=-80-85), 0.5 to 1.7 mm long. The matrix consists of 40-50% plagioclase laths, 0.04-0.8 mm long; and 22-40% clinopyroxenes, 0.05-0.2 mm long. Magnesium-rich titanomagnetites of several generations make up 7-15% of the sections in morphologies that include euhedral grains, some with blue-gray chrome spinel cores, skeletal coarse herring-bone crystals, and ultra-fine grains. Very few oxidation exsolution lamellae are present and the titanomagnetites are classified C1 to C2. There is up to 6% devitrified glass and trace amounts of pyrite and chalcopyrite. Abundant amygdules, up to 0.8mm in diameter, are filled by devitrified glass, or secondary sheet silicates. Alteration is limited to the olivines, of which 45% are altered to chlorite. There are traces of late hematite and calcite in the sections.

Petrographically these diabases are distinct from the other Massachusetts diabases in the presence of abundant phenocrystic olivine, and the chrome spinel cores in titanomagnetites. The titanomagnetite rims result from the reaction chrome spinel + liquid = titanomagnetite. Based on the classification of Yoder and Tilley (1962) and MacDonald and Katsura (1964) these diabases are olivine tholeiites. These diabases are the first reported tholeiites in the New England - Quebec Cretaceous igneous province of McHone and Butler (1984). The presence of amygdules indicates a shallow level of intrusion, similar to the Bliss Hill diabases.

**Major Element Concentrations in Relation to Petrography and Magma Evolution**

Grove et al. (1982) devised a method for comparing the compositions of basalts with experimental data on liquidus phase relations, particularly based on experiments with mid-ocean-ridge basalts. Both the composition and experimental data are treated in a normative tetrahedron olivine - calcic pyroxene - plagioclase - quartz, and are then plotted on a projection from plagioclase onto the olivine (Oliv) - calcic pyroxene (Cpx) - quartz (Qtz) base as in Figure 5. This shows fields of primary co-saturation of liquids with plagioclase and each of the phases olivine (oliv), augite (aug), pigeonite (pig), and orthopyroxene (opx). The 1 atmosphere field boundaries are the dark solid curves, and the fields of primary crystallization at 1 kbar $pH_2O$ are shown dashed. The plagioclase-orthopyroxene-calcic pyroxene plane that separates olivine-normative from quartz-normative tholeiites appears in the plagioclase projection of Figure 5 as the orthopyroxene - calcic pyroxene line. The approximate compositions of augite (Aug), pigeonite (Pig), and orthopyroxene (Opx) coexisting with plagioclase, olivine, and liquid at 1 atmosphere are also plotted.

All except one analysis (CS) of the central Massachusetts diabases plot within the triangle olivine-clinopyroxene-quartz and hence are tholeiites, but are clearly split into those that are quartz-normative and those that

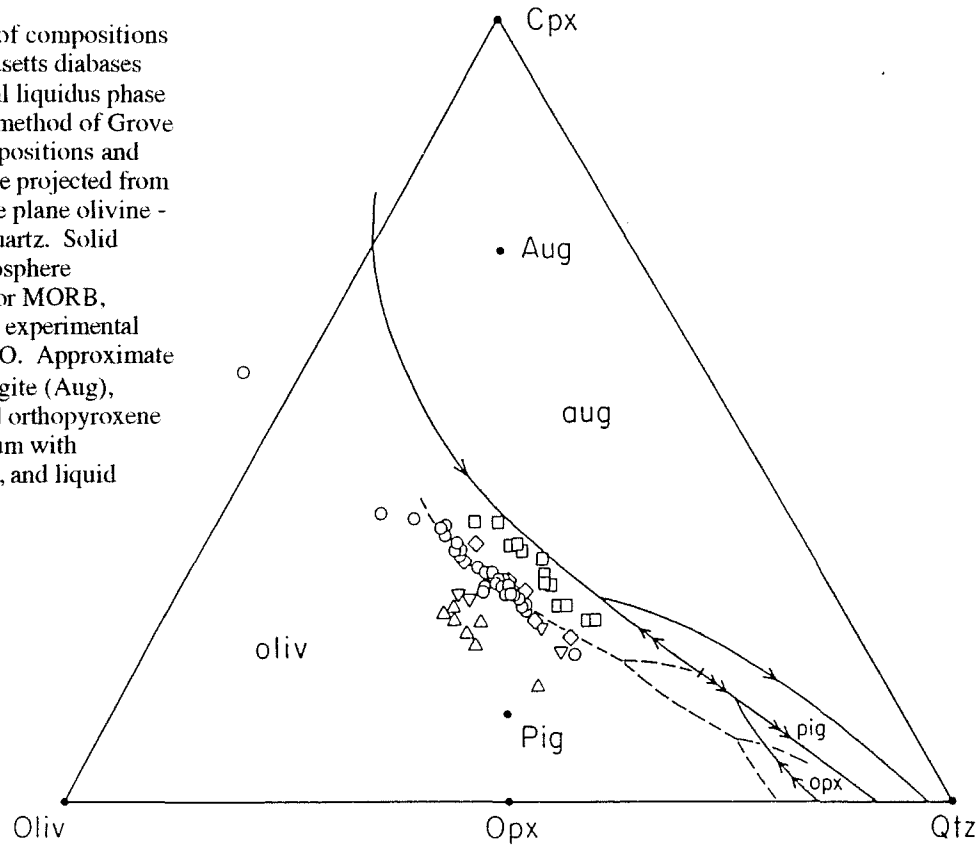



Figure 5. Comparison of compositions of central Massachusetts diabases against experimental liquidus phase relations using the method of Grove et al. (1982). Compositions and liquidus relations are projected from plagioclase onto the plane olivine - calcic pyroxene - quartz. Solid curves show 1 atmosphere experimental data for MORB, dashed curves show experimental data for 1 kbar $pH_2O$. Approximate compositions of augite (Aug), pigeonite (Pig), and orthopyroxene (Opx) in equilibrium with plagioclase, olivine, and liquid indicated.

are olivine-normative. The analyses from the Holden, Pelham-Loudville and Ware systems all plot in the projected olivine field of primary crystallization, yet show little to no olivine in their mineralogy. No olivine was found at all in thin sections in the Holden system. In the Pelham-Loudville and Ware systems olivine was found in a few samples and there is the suggestion of a peritectic reaction relationship in which olivine appears to have reacted with liquid to produce augite and pigeonite. The presence of biotite could explain some normative olivine, because it is a silica undersaturated mineral. The Bliss Hill diabases plot in two distinct areas. The northern diabase (4I4) is olivine-normative and contains relict olivine. The southern diabase (BH) is quartz-normative. The Quabbin Reservoir diabases are the only diabases with abundant phenocrysts of olivine and with one exception all plot in the olivine-normative field. The exception is from a chilled margin, which may not have contained any phenocrysts of olivine. The Quabbin Reservoir diabases also have abundant augite in the matrix.

Curiously, the Holden, Pelham-Loudville, and Ware diabases lie on trends that are parallel to the olivine-clinopyroxene boundary suggesting that the chemistry in these systems was controlled by co-saturation with both these phases. This fact is puzzling for the Holden system in that the dominant primary phenocryst phase is orthopyroxene. The Pelham-Loudville and Ware systems plot nearly perfectly on the olivine-clinopyroxene (+plagioclase) boundary at 1 kbar $pH_2O$. At 1 kbar $pH_2O$ the Holden system plots above the olivine-clinopyroxene boundary and in the field of augite (+plagioclase). Biotite is present in the thin sections from these systems that appears to be magmatic. This would support some $H_2O$ present in the system. A completely different suggestion is presented by the studies of McKenzie and Bickle (1988) on the composition of melt generated by lithospheric extension. The suggestion is that these linear trends are unrelated to crystal saturation during cooling, but represent mixing lines between early liquids of deep-seated origin and later liquids of shallower origin, all produced by continuous decompressive melting of the same mantle source (see especially their Figures 11 and 20). These brief observations plus other features in the geochemistry, particularly in the Holden system, suggest there is a range of intriguing petrogenetic problems yet to be solved in these diabases.

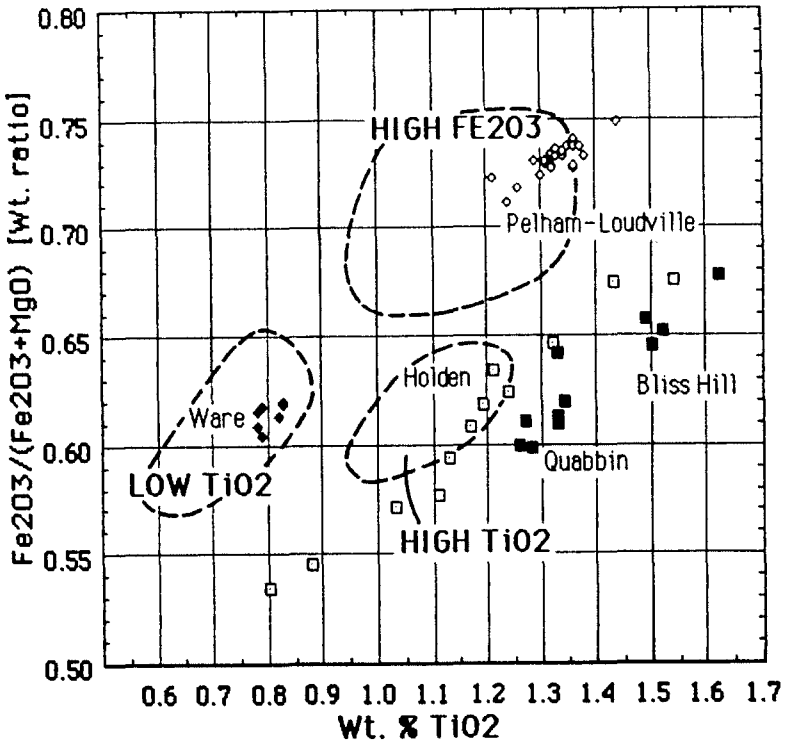



Figure 6. Diagram of weight ratio $Fe_2O_3/(Fe_2O_3+MgO)$ versus wt.% $TiO_2$. Dashed lines indicate fields of low $TiO_2$, high $TiO_2$, and high $Fe_2O_3$ quartz-normative diabases as defined by Weigand and Ragland (1970). All analyses of Massachusetts diabases from this study are plotted. Symbols: Holden - open squares; Pelham-Loudville - open diamonds; Ware - closed diamonds; Bliss Hill and Quabbin Reservoir diabases - closed squares.

## DIABASE GEOCHEMISTRY

### Summary of Chemical Systems

The regional geochemical nature of diabases is shown in Figure 6. The diabases fall into all four of the chemical types for the Eastern North American diabases as defined by Weigand and Ragland (1970). The distinct petrographic nature of the diabases systems is also borne out in the distinct major and trace elements of each system.

The Holden system consistently shows a large range in major elements, MgO, $TiO_2$, $Al_2O_3$, $K_2O$, $P_2O_5$, and in trace elements Zr, Y, Nb, Cr, and Sr. This system is more enriched in Sr, Ce, Zr, Cr, and Ni than the Pelham-Loudville and Ware systems. There is some indication that the variation in elements is a result of crystal fractionation, but other patterns of elements would argue against this inference and support a mixing model or multiple intrusions of slightly different magmatic compositions. There is field evidence of at least two magmatic pulses in a number of the Holden system localities sampled. No significant differences in the phenocrystic assemblages are observed along the dikes, indicating that there is probably is a genetic relationship between the two pulses of intrusion. Although there is a considerable range in the chemistry of this system, there is little overlap between the Holden system and the other systems. The range in elements indicates a more complex history to these diabase intrusions than has previously been postulated (Philpotts and Martello,1986). As shown in the spider diagram of Figure 7, there is a definite chemical pattern that differs markedly from the other Jurassic and Cretaceous systems. Trace elements Y and Sr (Figure 8) show considerable variation. The Zr/Nb ratios fall in the relatively narrow range of 11.7 to 14.6 (Figure 9 ), suggesting a consistent magmatic source for all the intrusions in the Holden system.

The Pelham-Loudville system is characterized by very high iron content (greater than 15 wt.% $Fe_2O_3$), and V. The $TiO_2$, Y, and V contents are the highest of the diabases, and the system is poorest in MgO, $Al_2O_3$, Cr, and Sr. The V and Y contents are particularly striking. All V analyses lie between 340 and 400 ppm, whereas no analyses from other systems exceeds 280 ppm. All Y analyses lie between 32 and 39 ppm (Figure 8) whereas all analyses of other systems lie between 15 and 20 ppm. The negative Sr anomaly in Figure 7 is characteristic and is markedly different from the diagrams of the other systems. The Zr/Nb ratio varies from 16 to 22.

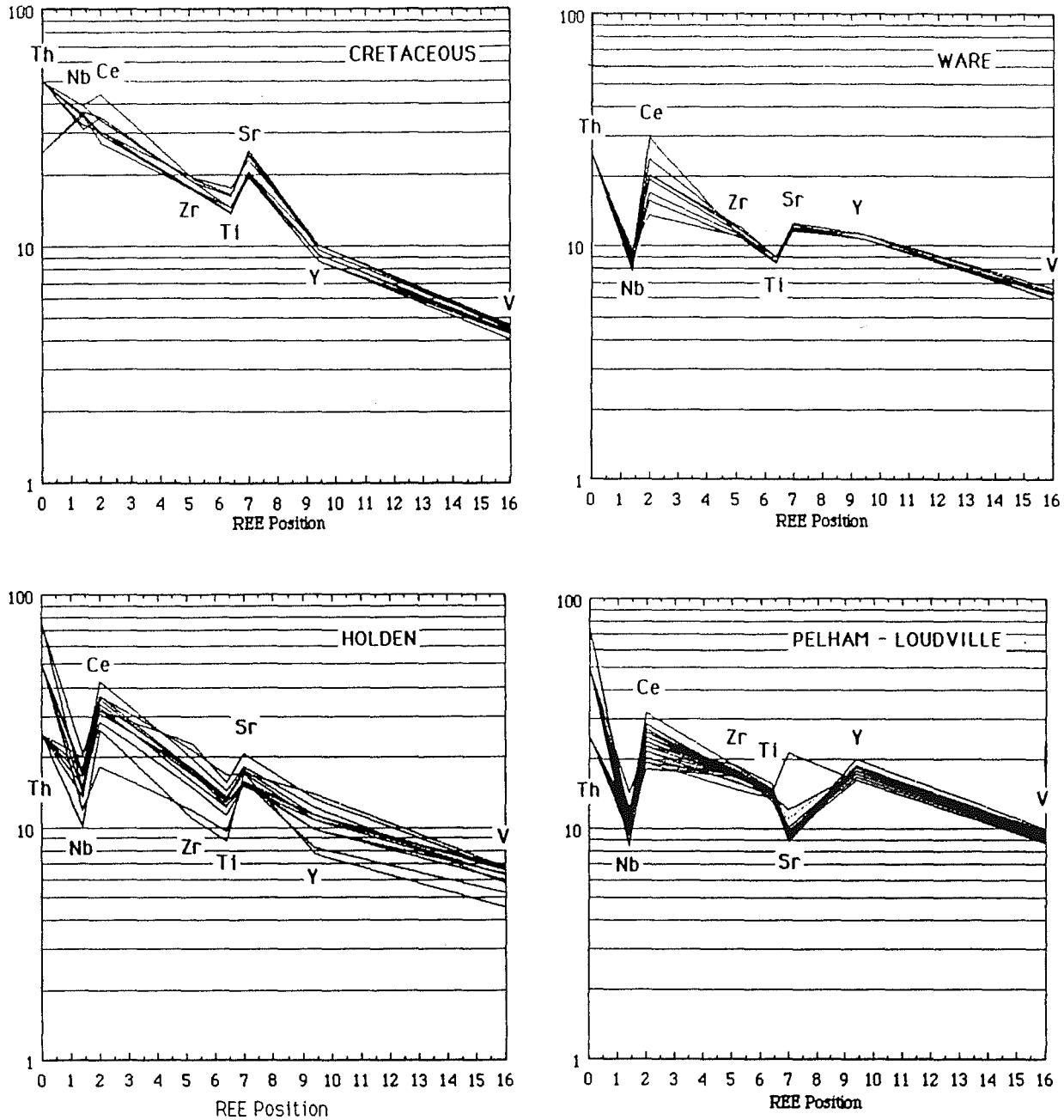



Figure 7. Spider diagram of chondrite-normalized trace element concentrations for all the central Massachusetts diabases. Elements are arranged according to the REE position by the method of Bougault (1980).

The chemistry of the Ware system is tightly clustered. The striking negative $TiO_2$ anomaly in Figure 7 is indicative of the low $TiO_2$ contents of all samples from this system regardless of geographic location (note French King Brige). The average MgO content of this system is the highest of the Jurassic intrusions. The iron content is markedly lower than the Pelham-Loudville system, but is the same as the Holden system. Like the Pelham system, the Ware system is consistently low in Sr, tightly clustered between 128 and 139 ppm (Figures 7 and 9 ), and is slightly higher than typical for the Pelham-Loudville system. The average Zr and Nb contents Figure 9) are lower

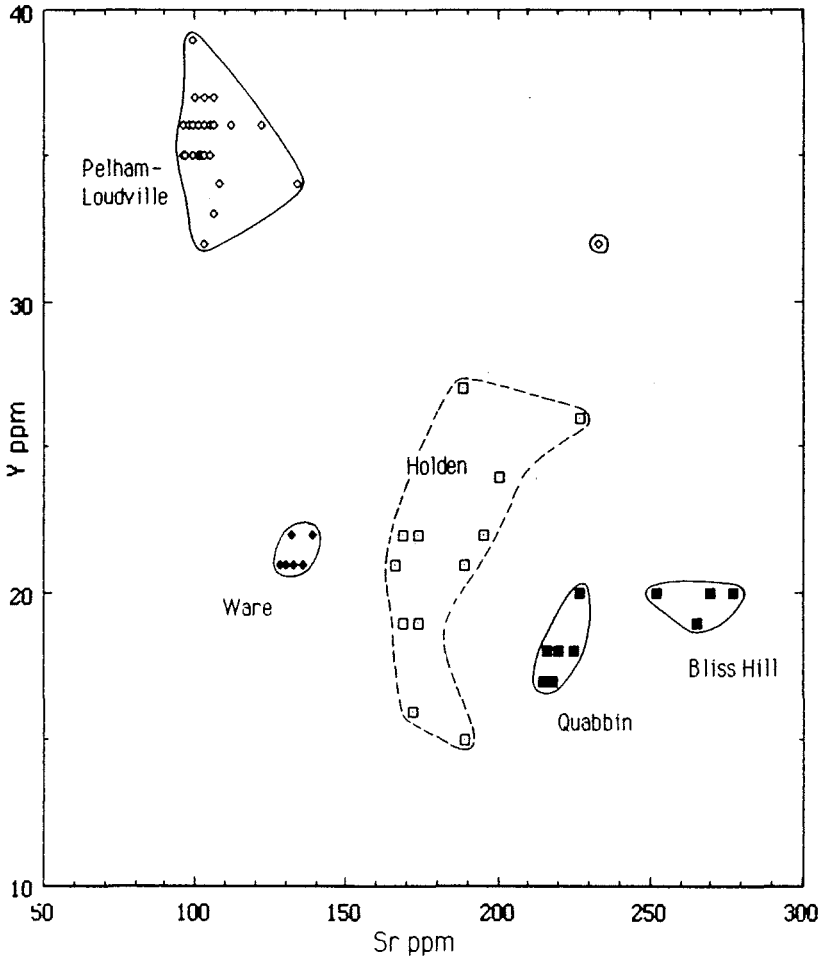



Figure 8. Plot of Y ppm versus Sr ppm. All diabase analyses from this study are plotted. Symbols: Holden - open squares; Pelham-Loudville - open diamonds; Ware - closed diamonds; Bliss Hill and Quabbin Reservoir diabases - closed squares.

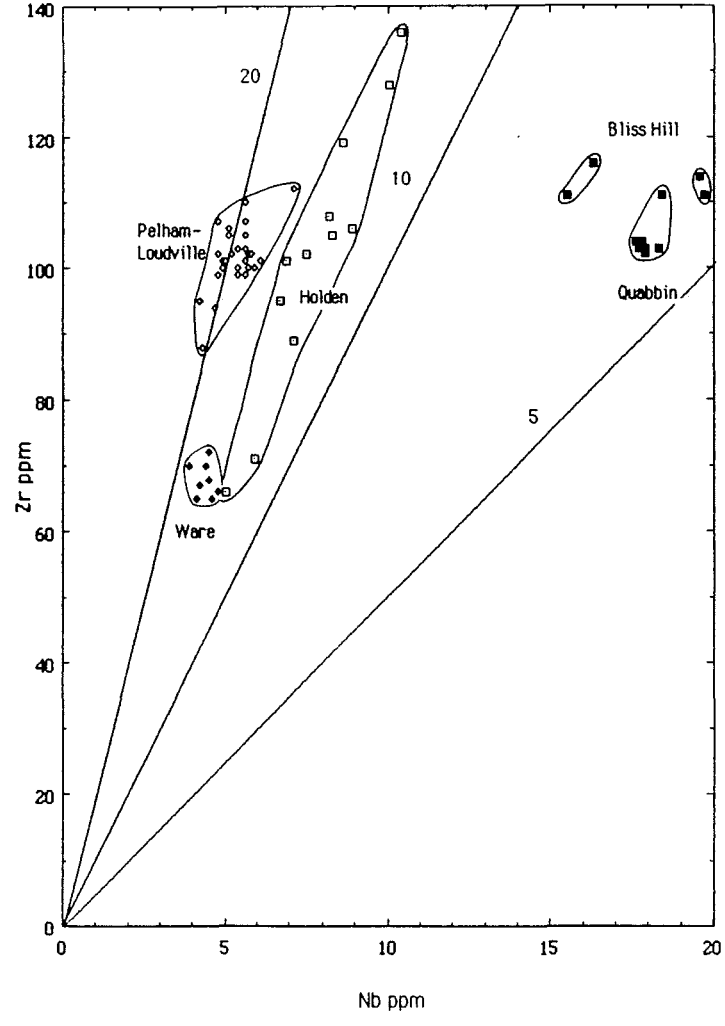



Figure 9. Plot of Zr ppm versus Nb ppm. All diabase analyses from this study are plotted. Symbols: Holden - open squares; Pelham-Loudville - open diamonds; Ware - closed diamonds; Bliss Hill and Quabbin Reservoir diabases - closed squares.

than all the other systems. Zr in the Ware system is consistently between 65 and 72 ppm, whereas only two other analyses fall as low as this and most fall between 90 and 120 ppm. The Zr/Nb ratio ranges from 14.4 to 17.5.

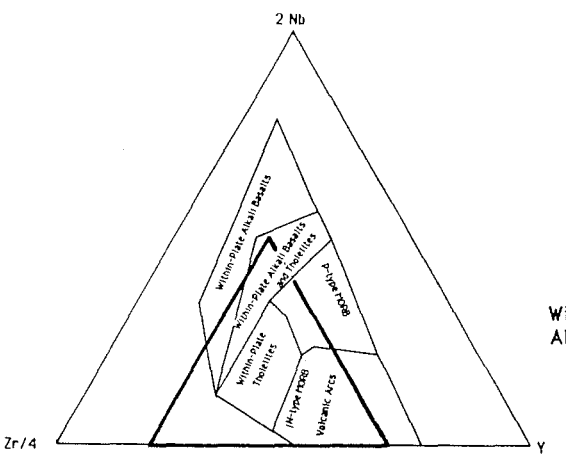
The Bliss Hill and Quabbin Reservoir diabases are distinct chemically from all the other intrusions. The Bliss Hill group is the highest in $TiO_2$ and $Al_2O_3$, whereas the Quabbin diabases are the most enriched in MgO. These diabases are also enriched in Ni, Cr, Sr, Ba, Zr, and Nb and low in V compared to the other diabases. For example, Bliss Hill and Quabbin Reservoir intrusions have less than 200 ppm V, whereas only one analysis from the other diabases falls in this category. Further, all the Bliss Hill and Quabbin Reservoir diabases have between 15 and 20 ppm Nb (Figure 9), whereas all the other diabases have between 4 and 11 ppm. The Zr/Nb ratios in the Bliss Hill and Quabbin Reservoir diabases range from 5.6 to 7.9, considerably different from the Zr/Nb ratios in the other systems which range from 11.7 to 22.

On the basis of the major- and trace-element analyses described above, it is concluded that the Holden, Pelham-Loudville and Ware systems, and the Bliss Hill and Quabbin Reservoir diabases were probably produced from different upper mantle sources, or at least produced from the upper mantle under different conditions. This is consistent with the isotopic differences demonstrated by Pegram (1986) between the correlative Jurassic diabase groups in Connecticut. The geochemical grouping of the diabases demonstrated here is also consistent with their paleomagnetic characteristics

**Discriminant Diagrams**

Meschede (1986) uses the immobile trace elements Nb, Zr, and Y to discriminate between N-type MORBs, P-type MORBs, within-plate tholeiites, and within-plate alkali basalts (Figure 10). Nb was chosen because it is a sensitive indicator of mantle enrichment or depletion processes, and Zr and Y because they show these trends differently from Nb. This discriminant diagram is based on more than 1800 analyses from different geotectonic settings of basaltic rocks. The suggested application is limited to tholeiites and all types of within-plate basalts. Meschede reevaluated data from ancient rocks that were known to be within-plate basalts, but in previous diagrams had plotted in the MORB field. Using this diagram these plot as within-plate basalts. The success of this discriminant diagram for ancient within-plate basalts makes the use of it pertinent. The Massachusetts Mesozoic diabases plot in three fields, and are discussed in order of relative age based on the paleomagnetic data. The oldest, the Holden dikes, all plot in the field of within-plate tholeiites. The Pelham-Loudville intrusions fall in the field of N-type MORB as do the Ware diabases. The youngest, the Bliss Hill and Quabbin Reservoir Cretaceous diabases plot in the field of within-plate alkali basalts and tholeiites. Similar tectonic relationships were found based on a Pearce and Norry (1979) discriminant diagram that uses the immobile trace elements Zr and Y. This diagram suggests a more oceanic-type magmas for the Pelham-Loudville and the Ware systems. The homogeneity in the major- and trace-element analyses from these two systems also suggests a large or a well mixed source for the magmas. This diagram also suggests a within-plate type setting for the Holden system, and for the Bliss Hill and Quabbin Reservoir diabases.

The question arises as to whether the discriminant diagram of Figure 10 fails, or succeeds. The oldest Jurassic diabases of the Holden system, and the Cretaceous groups plot well inside the within-plate fields, consistent with the location where they were intruded. However, the Pelham-Loudville and Ware diabases were also intruded into continental crust even though they have some trace-element characteristics of N-type MORB. Based on thermal modelling of the structural evolution of the Newark basin, Huntoon and Furlong, (1989) suggest that the source of the Jurassic basalts in the Newark basin may not have been beneath the basin, and that the magmas possibly migrated from a region of greater crustal extension offshore, where the birth of the modern Atlantic was about to take place. The geochemical differences may well reflect significant changes in the thermal-tectonic regime over the extended period from earliest Jurassic through the early Cretaceous. In this connection J. A. Philpotts (1985) states that perhaps the best current interpretation for the eastern North American diabases calls for a number of mantle source regions, perhaps previously subduction-related as proposed by Pegram (1986, 1989), to account for his Nd, Sr, and Pb isotopic data. The low Nb contents of the Jurassic diabases from the present study also imply a mantle source that has been involved in subduction-related magmatic processes. Similar unusual isotopic data were also recently reported by Lambert et al. (1988) for Mesozoic volcanic suites in the Canadian segment of the eastern North American margin. Philpotts suggests that the conflicting results of the discriminant diagrams, may indicate a relatively unusual and little sampled thermal-tectonic regime during the generation of some of the Mesozoic diabases.




All diabase analyses are plotted within the area of the enlarged inset.
Symbols: Holden - squares; Pelham-Loudville - circles; Ware - diamonds; Quabbin Reservoir - triangles; and Bliss Hill - inverted triangles.

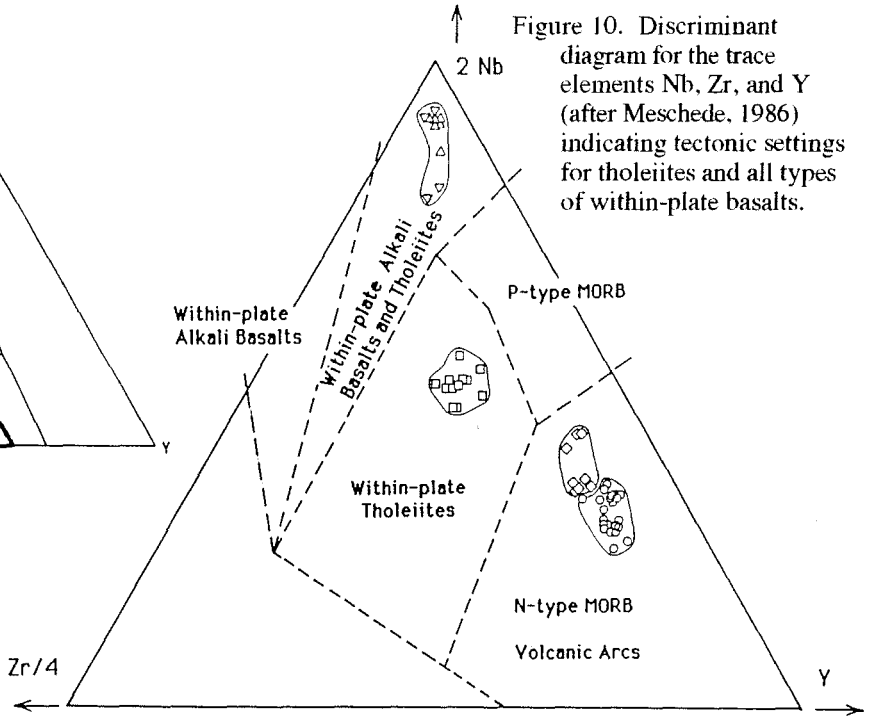



Figure 10. Discriminant diagram for the trace elements Nb, Zr, and Y (after Meschede, 1986) indicating tectonic settings for tholeiites and all types of within-plate basalts.

Based on these discriminant diagrams and the relative age constraints from the paleomagnetic pole positions, a possible scenario for the emplacement of the central Massachusetts diabases and magnetization of the sediments could be as follows:

1. Emplacement of the Holden dike system into a thick continental crust, as characterized by the with-in plate chemical signature, prior to the opening of the Atlantic and the J1 cusp. This magmatism is correlative with the earliest Jurassic Talcott Basalt that was erupted at about the same time as the deposition and primary magnetization of the Sugarloaf mudstones.

2. Thinning of the continental crust, development of a large offshore magma source in the location of the newly forming Atlantic spreading center, and the emplacement of the Pelham-Loudville system after the J1 cusp, possibly by migration from this large magma source. This would account for the striking homogeneity in the chemistry and the oceanic affinities in discriminant diagrams of this system. The migration of magma along fractures from offshore in the Early Jurassic (post-J1 cusp) is supported by thermal and gravity modelling of the Newark basin (Steckler and Karner 1988, Bell et al. 1989; Huntoon and Furlong, 1989). The isotopic data of Pegram (1989) does not support a MORB component for the correlative Jurassic diabases in Connecticut.

3. Change from a rifting to a drifting environment is thought to have occurred around 175 m.y. (Klitgord and Schouten, 1986). The Ware system diabases may be related to this tectonic event. The chemistry of the Ware samples has oceanic affinities and there is a strikingly similarity to the Early Jurassic Atlantic samples in Ti and V contents. The higher latitude pole position for the Ware system would indicate a younger age than the Holden or Pelham-Loudville systems. The possibility of an unusual magnetic signature due to secular variation must be considered, however, because the K-Ar ages are inconclusive. This could lead to an older age assignment for the Ware system than 175 m.y. The secondary magnetization found in the Sugarloaf arkoses is similar in direction to that found in the Ware dikes.

4. By 120 m.y. the thermal regime had changed. There was a well established spreading center at the mid-Atlantic ridge far offshore. There was a return to with-in plate igneous activity. The Bliss Hill and Quabbin Reservoir diabases are tholeiites, but are far more enriched in Ni, Cr, Sr, Ba, Zr, and Nb than the Jurassic diabases. The stress orientations at this time were possibly related to the separation of Newfoundland and Great Britain, but the igneous

activity in Massachusetts may be related to a hot spot, as is postulated for the Cretaceous intrusions in the White Mountains (Foland pers. comm. 1988).

## IMPLICATIONS

Field evidence (Philpotts and Martello, 1986) has invalidated the method used to group the igneous rocks used in the N1 and N2 poles. It is suggested here that there is very strong evidence to completely regroup VGP's used in N1 and N2 poles based geochemistry, which is a function of the magma types. This way of separating the data leads to significantly different pole positions. The paleopole from the first phase of intrusives in the Hartford basin plots near the Triassic-Jurassic track prior to the J1 cusp. These intrusions are all with-in -plate tholeiites (McEnroe, 1991; McEnroe, and Brown 1992). This pole position is in good agreement with poles from the Chinle formation obtained from the Carnian and Norian stages of the late Triassic, all indicating a westward progression of the APW (Bazard and Butler, 1991) during the Late Triassic to very Early Jurassic. A sharp change in the plate motion is indicated by the abrupt change in the APW path. This change is documented in the path and reflected as a cusp and well resolved by the Wingate Formation. Then an eastward progression of the APW is documented by the Moenave Formation (59.4°N, 59.2°E, Ekstrand and Butler, 1989). Paleopoles from the second phase of magmatic activity, Pelham-Loudville intrusions in Massachusetts and the Bridgeport dikes in Connecticut, plot near the J1 - J2 track.

The paleopoles from the Ware - Buttress dikes, interpreted as the youngest set of Jurassic tholeiitic dikes, plot at a high latitude (73°N), north of the J1 - J2 track. This high latitude pole determined for the Ware-Buttress system is too shallow for a present-day overprint. In some of the samples the titanomagnetite grains are altered, yet in many others no alteration is evident and both yield identical results. This suggests that the high latitude position represents the original thermal remanent magnetization obtained during the cooling of the dike. Also, at this time a minor chemical remanent magnetization (CRM) was formed by deuteric alteration of the titanomagnetite. This CRM is not a later remagnetization. This high latitude position is also found as an overprint in the sedimentary rocks of the Hartford and Deerfield basins (Brown, 1988) and Newark Basin (Witte and Kent, 1990, 1991). Hubert et al. (in press) discusses two generations of hematite formation, an early hematite acquired near the time of deposition and a hematite pigment formed by intrastratal solution of iron bearing silicate minerals in the permeable strata of the Hartford Basin. The latter process could continue over tens of millions of years. It is possible that this later hematite in the Deerfield basin is carrying the later overprint found in both the mudstones and the arkoses. Though it is not possible to date the formation of this hematite pigment, it is obviously younger than the authigenic hematite pigment formed by the dehydration of iron-hydroxide surface stains. Further support for this Mid-Jurassic high latitude pole position can be found in the European data set in the Jura Blue Limestones in Switzerland (Johnson et al., 1984), and the Malm Alpha and Beta Limestones in Germany (Heller, 1978). The mid-Jurassic Atlantic Seamounts also document a high latitude pole position for North America at this time (Mayhew, 1986). Limited data for the Mid-Jurassic and erroneous pole positions calculated for the Early Jurassic APW paths, especially those using restricted or synthetic data sets (Gordon et al., 1984; May and Butler, 1986), have lead to an incorrect APW for North America in the mid- to late Mesozoic.

## ACKNOWLEDGMENTS

The authors gratefully acknowledge contributions by Professor Kenneth Foland of Ohio State University for thirteen K-Ar whole rock ages and one $^{40}Ar/^{39}Ar$ release spectrum, Professor J.M. Rhodes for support of thirty XRF analyses, the Massachusetts Water Resources Authority at Quabbin for access to the Quabbin Reservation, and Holden Trap Rock Quarry, Ken Howatt and Bob Smith for permitting and making arrangements for sampling and the quarry visit of the field trip. Early work on the sediments were supported in part by NSF and the Research Corporation. Suzanne McEnroe gratefully acknowledges Sigmi Xi for support of field work, and Kim Mitchell and her family for help in preparation of the road log (Alexandra's first day in the field).

## ROAD LOG

The field trip will assemble in the parking lot north of the U. Mass Football Stadium at 8:00 A. M. An attempt will be made to consolidate into as few vehicles as possible and the trip will end back at the Stadium. Drivers should be prepared with sufficient fuel for 140 miles on the road.

**Mileage**

0.0 Turn left out of east exit of Stadium and proceed north on University Drive.

0.4 T intersection and flashing lights. Turn left (west).

0.8 Turn left on northbound entrance ramp for Route 116. North on Route 116.
1.7 Take first exit (right) off Route 116 for Route 63 North.
2.1 Left turn (north) at traffic lights in center of North Amherst for Route 63 North (this can be a bottleneck). After making turn at light, bear right immediately on Route 63.
9.9 Junction with Route 47, stay on Route 63.
12.2 Left turn (northwest) under powerline on County Road toward Lake Pleasant.
14.4 T junction. Turn left (west) toward Turners Falls.
17.7 Traffic lights in center of Turners Falls. Turn right (north) on Avenue A.
17.9 Begin crossing bridge over Connecticut River.
18.3 Traffic lights at junction with Route 2. Turn left (west) onto Route 2.
18.4 Sharp right bend where oncoming traffic is visible. Turn left across traffic into west entrance of parking area on eastbound (south) side of road and park overlooking the Connecticut River.

**STOP 1. RED ARKOSES AND GRAY MUDSTONES OF THE EARLIEST JURASSIC PART OF THE SUGARLOAF FORMATION** (40 MINUTES) Sandstones and mudstones of the earliest Jurassic upper part of the Sugarloaf Formation crop out to the west of the parking area on the south side of Route 2 across the Fall River. Here 15 meters of lacustrine and fluvial facies are found directly beneath the pillowed base of the Deerfield Basalt. Paleomagnetic samples were collected throughout the sequence at approximately 1 meter intervals. A similar section of the Sugarloaf Formation was sampled at Poet's Seat, two miles southwest of this stop. Mudstones from both sections have three components of magnetization (present day overprint, secondary steep component post-tilting and a primary magnetization pre-tilting). The sandstones exhibit only the first two components; all their magnetization is secondary and post-tilting. From the parking area there is an excellent vista of the flow top of the Deerfield Basalt (south and west) and the overlying Turners Falls Sandstone in the bed of the Connecticut River.

Following Stop 1 proceed east on Route 2.
18.6 Traffic lights at end of Connecticut River bridge. Stay straight on Route 2. Do not cross river.
21.7 Begin crossing of French King Bridge over Connecticut River. This is the location of the Connecticut Valley eastern border fault.
22.4 Turn right on River Road at sign for Route 63 South.
23.1 Stop sign. Make sharp right turn (south) onto Route 63 and travel for one block.
23.2 Turn right (west) into factory parking lot just before bridge over Millers River. Park in first available space on left hand side adjacent to river. Walk east a short distance to outcrop near north abutments of bridge.

**STOP 2. THE MILLERS FALLS SILL OF THE PELHAM-LOUDVILLE SYSTEM** (15 MINUTES) The Millers Falls sill is at the northern end of a group of dikes and sills that intrude the strata of the Pelham gneiss dome a short distance east of the Connecticut Valley border fault, and was mapped in detail by Laird (1974). This outcrop is in the Village of Millers Falls at the north end of the bridge over the Millers River. The sill is subparallel to foliation of the Hornblende Member of the Dry Hill Gneiss, striking N56E and dipping 20NW. The thickness of the sill is approximately 30 m. The chilled contact at the top of the sill is perfectly exposed west of the northern bridge abutment. The petrography and geochemistry are similar to most other members of the system (see text).

Progressive AF demagnetization studies on selected samples in the Pelham-Loudville system in general, show the presence of two components of magnetization. A minor viscous secondary component due to post-solidification alteration and/or a low-coercivity present field overprint is removed at low fields followed by an orderly decay of the stable remanent component. Based on petrographic work, the main magnetic carrier is titanomagnetite. There is fine-grained pseudo-single-domain titanomagnetite in the contact samples and multidomain titanomagnetites in the interiors that are less altered than in the Holden or Ware systems. Note that the intensities in this system from 0.8 to 4.0 A/M, are an order of magnitude greater than in the Holden or Ware systems. The magnetic decay patterns from the twenty-four sites in the Pelham-Loudville system show the largest variety in overprint directions. These overprints are soft and are easily removed. The majority of the sites yielded stable remanent directions with good in-site, and site-to-site agreement after AF demagnetization.

Following stop proceed back to Route 63.
23.3 Stop sign at north end of bridge. Turn left (north) on Route 63.
23.5 Turn right just beyond railroad crossing off Route 63 onto road leading to Route 2 East.
24.0 Oblique entrance onto Route 2 East.
29.4 Country Store in center of Erving.
31.0 Road forks. Stay left on Route 2.

35.6 Take exit ramp for Exit 15 to Route 122.
35.9 Junction with Route 122. Turn right (south).
36.5 Stop sign at junction with Route 202. Stay on combined Routes 122 and 202.
37.4 Junction. Take Route 122 left (southeast) off of Route 202.
44.7 Junction with Route 32A. Stay on Route 122.
45.2 Junction with Route 32. Continue straight (east) on combined Routes 122 and 32.
52.3 Center of Barre. Stay on Route 122.
56.1 On right is Ware River Intake Works for Quabbin Tunnel.
56.6 Bridge over Ware River.
59.5 Rutland Town Line.
61.1 Junction with Route 122A. Turn left (northeast) on Route 122A.
63.4 Junction with Route 56 in center of Rutland. Stay on Route 122A.
65.2 Holden Town Line
65.3 Turn right into entrance for Holden Trap Rock Quarry.

**STOP 3. HOLDEN TRAP ROCK QUARRY** ( 60 MINUTES) After stopping at the quarry office for hard hats and safety instructions, drive to high point at north edge of pit and park. Walk south a short distance down the main ramp into the pit and turn sharp left (northeast) on a smaller ramp where a dikelet is beautifully exposed. Continue down this small ramp to a large exposure of the main dike along the north wall of the quarry. If the pit is sufficiently pumped out it may be possible to visit other localities including the contacts at the south wall.

Based on reconnaissance mapping and interpretation of the aeromagnetic map by Peter Robinson and R. D. Tucker for the State bedrock map (Zen et al., 1983), the Holden Trap Rock Quarry is in the southeastern of three en echelon segments of the Holden dike. Sample locations were based on a pit map prepared by Robinson and Tucker on December 15, 1976. The main dike (samples numbered 1681 and 1683) and a dikelet (sample 1688) were sampled.

According to the pit map the dike trends approximately north at the northeast end of the pit where it has a measured thickness of 39 m. In the central part of the pit the dike turns to a trend of about N45E and reaches a maximum thickness of about 55 m. In the southern part of the pit the dike bends back to a more northerly trend and thins to about 25 m. The overall exposure along the length of the dike is 180 m and the contact attitudes are highly irregular. Individual attitudes measured along the northwest contact include N1E 85E, N67E 86E, N10E 87W and N5E 73E; along the southeast contact N15E 60W, N6E 50W, N47E 79W, and N62E 79SE. The country rocks on the northwest side of the main dike include the Granulite Member of the Paxton Formation (Sp) containing a separately mapped layer of rusty mica schist and an overlying sill of foliated biotite granitic gneiss. Country rock contacts dip gently to the northeast. The country rock on the southeast side of the main dike is the quartzite and rusty schist member of the Paxton Formation (Spqr) also dipping gently. These relations combined with regional geology suggest there is a normal fault, down on the northwest, along the dike. This is confirmed in the northeast corner of the quarry where slickensided fault surfaces in the diabase itself close to the southeast contact show attitudes of N9E 60W with slickensides plunging N80W, and N18E 57W with deep slickensides plunging N67W and later fine slickensides plunging 9S.

Within the main dike there appear to be multiple internal chilled contacts, away from which the diabase texture coarsens to that of an ophitic gabbro. All samples were taken from oriented blocks. Samples 1681 were all collected near the northwest contact of the main dike near the north end of the quarry. 1681B was at the contact, 1681A was 0.2-0.5 m east of the contact, and 1681C was 6 m east of the contact. An internal chill was observed in this vicinity 10 meters from the dike contact but no sample was obtained east of the chill. Sample 1683 consisted of coarse diabase collected 23 m east from the west contact at the south end of the pit, and probably lies east of the internal chilled contact described above. Site 1688 is from a newly located fine-grained dikelet, 0.41 m thick, with an attitude N 29W 70SW, that has intruded the biotite granitic gneiss in the northwestern part of the quarry. This is probably the same dikelet now exposed on the small entrance ramp described above. The petrography and geochemistry of these samples is typical of the Holden system as described above or in McEnroe (1989).

Magnetic intensities in the Holden system range from 0.4 to 2.4 A/M. Progressive AF demagnetization studies on selected samples indicate the presence of several components of magnetization including a low-coercivity present-field overprint. These soft magnetic components are considered to be a viscous remanence gained in the present-day field. By 40 mT the stable characteristic remanence was obtained in most samples.

After exiting quarry, turn right (east) on Route 122A and resume mileage.

66.8 Junction with Route 68. Go straight on Route 122A.

67.3 Turn left (northeast) on Mount Pleasant Avenue.

67.6 Four-way junction at Jefferson Post Office. Turn left (north) on Princeton Road.

68.3 Fork at Whitney Street. Stay right on Princeton Road.

68.8 Grade crossing of Providence and Worcester Railroad.

69.4 Bridge over spillway from Quinapoxet Reservoir. Park on either side. Climb around fence at bridge to outcrop immediately west of bridge. There is poison ivy close to the bridge but not at the outcrop.

**STOP 4. SPILLWAY EXPOSURE OF HOLDEN COMPOSITE DIKE AT QUINAPOXET RESERVOIR** (30 MINUTES) The exposure of the Holden dike exposed at Quinapoxet Reservoir is the northeastern extension of the northernmost of the three en echelon segments identified in the vicinity of the Holden Trap Rock Quarry. The dike intrudes vertically into the granulite member of the Paxton Formation and trends N49E. Only the southeast contact is exposed and there is a minimum thickness of 24 m of diabase. There is a 0.5 m screen of country rock separating the main dike from a 1.2 m dike to the southeast. All the samples were obtained by drilling the outcrop. Samples QX-1 and QX-2 are from the thin southeast dike, QX-1 at 0.28 m and QX-2 at 0.76 m from the southeast contact. Sample QX-3 was taken at the chilled contact of the main dike, and samples QX-4, 5, 6, 7, and 8 were taken at distances of 0.9 m, 1.8 m, 6.1 m, 9.7 m, and 13.1 m northwest of the contact of the main dike. The middle part of the exposure consists of medium-grained ophitic gabbro, but the northern part is finer-grained, suggesting a contact may be a short distance beyond the edge of the outcrop.

Following this stop return south along Princeton Road.

71.1 Jefferson Post Office. Turn right (west) on Mount Pleasant Avenue.

71.4 Stop sign. Bear right (west) on Route 122, pass entrance to Holden Trap Rock Quarry, and return through Rutland.

77.0 Sharp right turn for Rutland State Forest. Stay straight at next junction and pass over scenic causeway between lakes.

77.9 Junction Route 122. Bear right (west) on Route 122.

79.0 Junction Route 148. Stay on Route 122.

79.4 Left turn (west) off of Route 122 onto Old Turnpike Road toward Oakham.

84.1 Stop light at Route 67. Go straight across and continue west.

86.6 Bridge over Ware River and junction with Route 32. Go straight across through Village of Furnace toward Hardwick Center.

89.3 Hardwick Center and junction with Route 32A. Go straight through middle of common to stop sign and continue straight on Greenwich Road.

91.8 Gate 43, Quabbin Reservation. Proceed straight ahead through locked steel barway, and west on road to Intake Works.

93.9 Turn right (north) onto Baffle Dam Road and drive across South Baffle Dam.

94.1 North end of South Baffle Dam. Continue north on Baffle Dam road. As you drive northward along the road you will cross over a dike locality originally mapped by Robert Balk (1940) during construction of Quabbin Reservoir. The exposure was subsequently destroyed during road construction. It lies near a small road cut where the road from South Baffle Dam follows the southeast shore of Baffle Dam Island. Ground-magnetic profiles indicate a sill 1 m thick with strong reversed polarity. A sample from Balk's collection housed at the University of Massachusetts was analyzed geochemically and for K-Ar. Additional samples of a thinner chilled dikelet were obtained from rubble adjacent to the road.

94.3 Park at high point of road on Baffle Dam Island for Stop 5.

**STOP 5. THE BAFFLE DAM ISLAND SILL OF THE QUABBIN CRETACEOUS DIABASES** (30 MINUTES) From high point of road climb west up hill beginning to left (south) of large outcrop on road. A short distance above there is a sloping terrace below a small cliff with overhangs. Walk southwest along the terrace for 100-150 feet and then find the diabase sill about half way up the small cliff at elevation 655'. The country rock of the sill is Late Ordovician Monson Gneiss, about 200 feet east of the basal contact of the Partridge Formation, which crosses the top of the island. The Monson Gneiss consists of coarse-grained biotite-quartz-plagioclase gneiss, microcline augen gneiss, and large boudins of coarse-grained hornblende amphibolite that are probably metamorphosed gabbro. R. D. Tucker (personal communication, 1990) has obtained an Acadian age on metamorphic zircon from the microcline gneiss in a small road cut on the Baffle Dam road.

The diabase sill is about 0.3 m thick. It is irregular in orientation and shows excellent chills on both contacts It was discovered in the present study during field search for Balk's South Baffle Dam locality. The sill strikes N27E and dips 55SE. It has been traced along strike for a distance of 100 m, and where last exposed at the north end, turns into a dike striking N62W and dipping 63SW. The dike contains 5% of fresh and altered olivine phenocrysts up to 1.7 mm in diameter, 1% of plagioclase phenocrysts up to 1.6 mm in diameter, and 2-3% of amygdules up to 0.4 mm in diameter. In keeping with the narrow width of the dike, the matrix consists of very fine plagioclase (40-45%), augite ± pigeonite 22-29%, titanomagnetite C1-2 (14-16%) devitrified glass (5-6%) and chlorite 2-5%.

The Cretaceous diabases in Massachusetts record both normal and reversed polarity, at Bliss Hill and Quabbin Reservoir, respectively. These magnetic directions were compared to and agree well with other New England Cretaceous intrusions of approximately the same age. Three small sills are exposed in the Quabbin Reservoir area. The largest, 1m thick is beautifully exposed on a shore exposure at Chapman Island, inaccessible to this field trip. Its reversed remanent magnetization is so intense that it gives a negative ground-magnetic anomaly. (McEnroe, 1989, Figure 1.6). Samples from the Chapman Island and the Baffle Dam island sills show a small normal overprint that is removed by 10 mT, then a stable direction that is southerly and with a steep negative inclination. It is believed that the stable remanent direction obtained in these studies is the primary remanence obtained by the intrusions upon cooling. A thermal demagnetization study confirms the direction obtained in the AF demagnetization.

| | |
|---|---|
| | Continue north on Baffle Dam road. |
| 94.4 | South end of north Baffle Dam. Proceed across north Baffle Dam for scenic outlook. |
| 94.8 | North end of north Baffle Dam. Reverse direction with care and return across north Baffle Dam. |
| 95.5 | North end of south Baffle Dam. Continue south across south Baffle Dam. |
| 95.7 | Return to paved road. Turn left (east). |
| 97.7 | Gate 43 barway. Pass through locked barway and continue straight on Greenwich Road. |
| 100.3 | Center of Hardwick. Turn right (south) on Route 32A. |
| 102.8 | Junction with Route 32. Bear right (southwest) on Route 32. |
| 106.1 | Junction with Route 9. Turn sharp left (northeast) on Route 9. |
| 106.2 | Cross traffic to sandy area at ancient ruins of gas station which is just in sight of powerline. Turn around park here for Stop 6 with vehicles pointed west.. |

**STOP 6. THE WARE DIKE** (60 MINUTES) From sandy area cross Route 9 diagonally toward the east and find beginning of rocky trail which leads to powerline a short distance south of the Route 9. Continue south up steep trail along powerline with thick blackberry bushes to flatter area within sight of the summit of the hill. Just beyond the steepest section, the trail will pass over several diabase outcrops. Watch out for poison ivy in this area. The dike trends more easterly than the powerline and crosses from the east side to the west side as we walk south. Several hundred feet farther, but still well short of the hill summit, the trail passes near a country rock exposure with a diabase dikelet that lies east of the main dike. At this point walk a few meters west to a large outcrop of the main dike showing the chilled eastern contact.

In the exposures of the Ware System near Ware the dike intrudes the Rangeley Formation, the Gneiss of Ragged Hill, the Fitch (=Francestown) Formation, and the Coys Hill Granite. The strike of the dike is N25E, the dip is vertical, and the width is 30 m. This dike has been traced in the field for a distance of 5.4 km (Field, 1975). This dike has been sampled at two localities 1 km - 2 km south on the powerline from Route 9 and the small dikelet, less than 0.3 m thick that strikes N6E, and dips 74E near site WA. In the exposures to be seen on this field trip the dike cuts the Gneiss of Ragged Hill and Rangeley Formation. The Francestown Formation is exposed to the south at the high point of the powerline. The petrography and geochemistry of the samples from this location are typical of the Ware system (see text above) in which olivine is preserved in chilled margins, whereas interior samples consist of interlocking plagioclase, augite, pigeonite and titanomagnetite among which phenocrysts cannot be identified.

Progressive AF demagnetization studies on selected samples in the Ware system show the presence of two components of magnetization. A small component is removed at low fields. Based on petrographic work the main magnetic carrier is titanomagnetite. There is fine-grained unaltered pseudo-single-domain titanomagnetite in the contact samples. The euhedral multidomain titanomagnetite grains in the interior samples are altered to an Fe-Ti amorphous oxide. This alteration was probably deuteric because the same remanent direction is obtained from the contact samples as the interior samples. Intensities from the Ware system ranged from 0.3 to 2.4 A/M. Vector end-point diagrams of interior sites from the Ware system diabases show similar overprintings. There is a soft secondary component that is removed by 20 mT with an orderly linear decay following to the origin. The direction obtained

from the French King Bridge site is the same as the directions from the other diabases in the Ware system exposed 35 km to the southeast.

After return to vehicles enter Route 9 with care and proceed west.
106.3 Junction of Route 9 with Route 32. Continue west on Route 9.
107.5 Bridge over Ware River in Ware (Weir?).
113.4 Right turn off Route 9 toward Winsor Dam for Quabbin view.
114.0 Drive onto Winsor Dam.
114.9 Junction with Route 9. Turn right (west) onto Route 9.
117.9 Junction of Route 9 with Route 202. Turn right (north) on Route 202.
125.0 Yellow flashing light at Amherst Road, Pelham. Continue straight on Route 202.
125.5 Overlook for Quabbin Reservoir and Wachusett Mountain . Late Proterozoic Pelham Quartzite on left.
127.2 Road cut on left at dangerous corner exposes Pelham dike cutting the Mount Mineral Formation (see Robinson et al., Trip B-3, this guidebook). Continue north without pause, crossing culvert over Atherton Brook, and proceed to high point of next hill.
128.0 Pull off to right on ample parking area and walk southeast over summit of hill 1012 to large exposures on south ridge.

**STOP 7. PELHAM DIABASE DIKE** (60 MINUTES) The Pelham dike proper intrudes the strata of the Pelham gneiss dome directly east of the Amherst arch that separates the Deerfield and Hartford Mesozoic basins, It has a total exposed length of 9.5 km. This was sampled at four localities (Figure 1) from northeast to southwest; 3C6, 8C9, 2O2, NV. The exposure at this stop is on the south ridge of hill 1012 northeast of Atherton Brook and is the largest exposure of the Pelham dike. The dike intrudes the Fourmile Gneiss perpendicular to foliation, and is 20 m thick. The strike is N60E, and the dip is vertical. Sample 8C9A is from a 10 cm dikelet adjacent to the northwest contact of the main dike, sample 8C9B is from the northwest contact of the main dike, sample 8C9C is 8 m from the northwest contact, and sample 8C9D is at the southeast contact about 30 m due east from 8C9B.

Chilled margins in the Pelham dike contain microphenocrysts of olivine surrounded by augite. Phenocrysts of plagioclase and pyroxene contain abundant inclusions of trapped liquid that has crystallized glass or fine-grained crystals and oxides. Some plagioclase phenocrysts also contain inclusions of pyroxene. Titanomagnetite grains range in oxidation from C1 to C3. The texture varies from ultra-fine dust, skeletal herring-bone, to coarse-grained. Present in some of the titanomagnetites is an internal lit-par-lit texture. Pyrite is abundant, and texturally ranges from droplets to euhedral grains to titanomagnetite and pyrite intergrowths. Paleomagnetism of the Pelham-Loudville system is discussed under Stop 2.

END OF TRIP! To reach Amherst return south three miles to flashing yellow light. Turn right (west) on Amherst Road in Pelham, which becomes Pelham Road and then Main Street in Amherst. Distance to the football stadium from Stop 7 is about 10.5 miles.

## REFERENCES CITED

Ashenden, D.D., 1973, Stratigraphy and structure, northern portion of the Pelham Dome, north-central Massachusetts: Contribution No. 16 (M.S. thesis) Dept. of Geology, University of Massachusetts, Amherst, 132p.

Balk, Robert 1940, Preliminary report on the geology of the Quabbin Reservoir area, Massachusetts: U.S. Geological Survey open-file report, 58p.

Bazard, D. R., and Butler, R.F., 1991, Paleomagnetism of the Chinle and Kayenta Formations, New Mexico and Arizona.: Journal of Geophysical Research, v.96, p. 9847 - 9871.

Bell, R.E., Karner, G.D., and Steckler, M.S., 1989, Gravity anomalies, detachments and the source of mafic materials associated with the Early Mesozoic basins of eastern North America: Geological Society of America Abstracts with Programs, v. 21, p. 4.

Bougault, H.,1980, Contribution des elements de transition a la comprehension de la genese des basalts oceaniques: Ph.D. thesis, Universite de Paris VII, 221p.

Bowker, D.E., 1960, Remanent magnetization of eastern United States Triassic rocks: Ph.D. thesis, Massachusetts Institute of Technology, Cambridge Massachusetts, 186 p.

Brown, L.L., 1979, Magnetic observations of the red beds of the northern Connecticut Valley-DRM or CRM? (abstract): EOS, v. 60, p. 815.

Brown, L.L., 1988, Multicomponent paleomagnetic directions from the Sugarloaf Arkose, Deerfield basin, Massachusetts: Geological Society of America Abstracts with Programs, v. 20, p. 10.

De Boer, J. Z., 1967, Paleomagnetic-tectonic study for Mesozoic dike swarms in the Appalachians: Journal of Geophysical Research, 72, p. 2237-2250.

De Boer, J. Z., 1968, Paleomagnetic differentiation and correlation of the Late Triassic volcanic rocks of the central Appalachians (with special reference to the Connecticut Valley): Geological Society of America Bulletin, v. 79, p. 609-626.

DuBois, P. M., Irving, E., Opdyke, N.D., Runcorn, S.K, and Banks, M.R.,1957, The geomagnetic field in the Upper Triassic times in the United States: Nature, v.180, p. 1186-1187.

Dungan, M.A., and Rhodes, J.M., 1978, Residual glasses and melt inclusions in basalts from DSDP leg 45 and 46: evidence for magma mixing: Contributions to Mineralogy and Petrology, v.67, p. 417-431.

Elbert, D. C., 1986, Recognition and implications of a structurally inverted Monadnock- Western Maine stratigraphy directly above the core of the Bernardston Nappe, Hinsdale, New Hampshire: Geological Society of America Abstracts with Programs, v. 18, p.15.

Ekstrand, E. J., and Butler, R. F., 1989, Paleomagnetism of the Moenave Formation: Implications for the Mesozoic North American apparent polar wander path, Geology, 17, 245-248.

Field, M.T., 1975, Bedrock geology of the Ware area, central Massachusetts: .Contribution No. 22, (Ph.D. thesis), Department of Geology, University of Massachusetts, Amherst, 186p., 6 plates.

Gordon, R.G., Cox, A., and O'Hare, S., 1984, Paleomagnetic Euler poles and apparent polar wander and absolute motion of North America since the Carboniferous: Tectonics, v. 3, 499-537.

Grove, T.L., Gerlach, D.C., and Sando, T.W., 1982, Origin of calc-alkaline series lavas at Medicine Lake volcano by fractionation, assimilation and mixing: Contributions to Mineralogy and Petrology, v.80, p. 160-182.

Heller, F., (1978), Rock magnetic properties of Upper Jurassic limestones from southern Germany, J. Geophys., 44, 525-543.

Haggerty, S.E., 1976, Oxidation of opaque mineral oxides in basalts: in Rumble, D. III, editor, Reviews in Mineralogy, V. 3, Oxide Minerals: Mineralogical Society of America, Blacksburg, VA., p. Hg.1- Hg.100.

Huntoon, J.E, and Furlong K.P., 1989, Thermal evolution of the Mesozoic Newark Basin and implications for tectonic processes: (abstract): EOS, v. 70, p. 1344.

Hubert, J.F., Feshbach-Meriney, P.E., and Smith, M.A., (1992), The Triassic-Jurassic Hartford rift basin, Connecticut and Massachusetts: Evolution, sandstone diagenesis, and hydrocarbon history. AAPG Bull., in press.

Irving, E., and Irving G.A., 1982, Apparent polar wander paths: Carboniferous through Cenozoic and the assembly of Gondwana: Geophysical Survey, v. 5, p.141-188.

Johnson, R.J., Van der Voo, R.E., and Lowrie, W., (1984), Paleomagnetism and late diagenesis of Jurassic carbonates from the Jura Mountains, Switzerland and France, Geol. Soc. Am. Bull., 95, 467-488.

Klitgord, K.D., and Schouten, H., 1986, Plate kinematics of the central Atlantic: in Vogt, P.R., and Tucholke, B.E., Editors: The Geology of North America, Volume M,The Western North Atlantic Region, Geological Society America, Boulder, Colorado, p. 351-378.

Laird, S.H., 1974, Geology of the Pelham dome near Montague, west-central Massachusetts: Contribution No. 14 (M. S. thesis), Department of Geology and Geography, University of Massachusetts, Amherst, 84p.

Lambert, R. St. J., Pe-piper, G., and Jansa, L.F., 1988, Sr, Nd, and Pb isotopic data from Mesozoic volcanic suites on the eastern North American continental margin: Geological Association of Canada Programs with Abstracts, v. 13, p. A71.

Lofgren, G., 1974, An experimental study of plagioclase crystal morphology: Isothermal crystallization: American Journal of Science, v. 274, p. 243-273.

MacDonald, G.A., and Katsura, T., 1964, Chemical composition of Hawaiian lava: Journal of Petrology, v.5, p. 82-133.

Mayhew, M.A., 1986, Approximate paleomagnetic poles for some New England seamounts: Earth and Planetary Science Letters, 79 185-194.

May, S. R., and Butler, R.F., 1986, North American Jurassic apparent polar wander: implications for plate motion, paleogeography and Cordilleran tectonics: Journal of Geophysical Research, v. 91, p. 11519-11544.

McEnroe, S. A., 1988, Diversity in paleomagnetism, geochemistry and age in Mesozoic diabase intrusions, central Massachusetts: Geological Society of America Abstracts with Programs, v. 20, p. 55.

McEnroe, S. A., 1989, Paleomagnetism and geochemistry of Mesozoic diabase dikes and sills in west-central Massachusetts Contribution No. 64 (M. S. thesis), Department of Geology and Geography, University of Massachusetts, Amherst, 136p.

McEnroe, S. A., 1991, The Newark Trend I and II pole positions: Are they valid Early Jurassic pole positions for North America? Geological Society of America Abstracts with Programs, v. 24, p. A317.

McEnroe, S. A., 1991, Constructing a Mesozoic Apparent Polar Wander Path for North America Based on Paleomagnetic data from intrusions in southern New England (extended abstract): Northeast Paleomagnetic Workshop, 5th Annual Meeting, Toronto, Canada.

McEnroe, S.A. and L.L. Brown, 1987, Paleomagnetism of diabase dikes and sills intruding pre-Mesozoic basement rocks near and east of the Connecticut Valley Mesozoic basins, west-central Massachusetts (abstract): EOS, v. 68, p. 295.

McEnroe, S.A. and L.L. Brown, 1988, Paleomagnetism of Mesozoic diabase intrusions and redbeds in central Massachusetts: Relations to the APWP for North America: Geological Society of America Abstracts with Programs, v. 20, p. A351.

McEnroe, S.A. and L.L. Brown, 1992, Implications for the Mesozoic APW for North America from intrusions and sedimentary rocks in New England: EOS, v. 73, p. 95.

McEnroe, S.A., Robinson, Peter., Brown, L.L., and Foland, K.A., 1987, Paleomagnetic discovery of Cretaceous diabase dikes in central Massachusetts: Geological Society of America Abstracts with Programs, v. 19, p.765.

McHone, J.G, and Butler, J.R., 1984, Mesozoic igneous province of New England and the opening of the Atlantic Ocean: Geological Society of America Bulletin, v. 94, p. 757-765.

McKenzie, D. and Bickle, M.J., 1988, Melt generated by lithospheric extension: Journal of Petrology, v.29, p. 625-679.

Meschede, M., 1986, A method of discriminating between different types of mid-Ocean ridge basalts and Continental tholeiites with the Nb-Zr-Y diagram: Chemical Geology, v. 56, 207-218.

Pearce, J.A., and Norry, M.J., 1979, Petrogenetic implications of Ti, Zr, Y, and Nb variations in volcanic rocks: Contributions to Mineralogy and Petrology, v. 69, p. 33-47.

Pegram, W. J., 1986, The isotope, trace element, and major element geochemistry of the Mesozoic Appalachian tholeiite province: Ph.D. thesis, Massachusetts Institute of Technology,Cambridge, Massachusetts, 622 pp.

Philpotts, A.R., and Martello, A., 1986, Diabase feeder dikes for the Mesozoic basalts in southern New England: American Journal of Science, v. 286, p. 105-126.

Philpotts A.R., and Reichenbach, Ingrid, 1985, Differentiation of Mesozoic basalts of the Hartford basin,Connecticut: Geological Society of America Bulletin, v.96, p. 1131-1139.

Philpotts, J.A., 1985, Pearce-Cann discriminant diagrams applied to the North American Mesozoic diabases: in Robinson, G.R., and Froelich, A.J., Editors, Proceedings of the Second U.S. Geological Survey Workshop on the Early Mesozoic Basins of the Eastern United States: U. S. Geological Survey Circular 946, p.114-117.

Robinson, Peter, 1986, Introduction: in Robinson and Elbert, Editors, Field trip guidebook: regional metamorphism and metamorphic phase relations in northwestern and central New England: Field Trip B-5, International Mineralogical Association, 14th General Meeting at Stanford University, Contribution No. 59, Department of Geology and Geography, University of Massachusetts, Amherst, Massachusetts, p. 1-10.

Robinson, Peter, Tracy, R.J., Hollocher, K.T., Schumacher, J.C., and Berry, B.N. IV, 1986, The central Massachusetts metamorphic high: in Robinson and Elbert, Editors, Field trip guidebook: regional metamorphism and metamorphic phase relations in northwestern and central New England: Field Trip B-5, International Mineralogical Association, 14th General Meeting at Stanford University, Contribution No. 59, Department of Geology and Geography, University of Massachusetts, Amherst, Massachusetts, p. 195-283.

Smith W.A., 1987, Paleomagnetic results from a crosscutting system of northwest and north-south trending diabase dikes in the North Carolina Piedmont: Tectonophysics, 136, p. 137-150.

Smith, T.E., 1976, Paleomagnetic study of lower Mesozoic diabase dikes and sills of Connecticut and Maryland: Canadian Journal of Earth Sciences, v. 13, 567-609.

Smith, T.E., and Noltimier, H.C., 1979, Paleomagnetism of the Newark trend igneous rocks of the north-central Appalachians and the opening of the central Atlantic Ocean: American Journal of Science, v. 279, p. 778-807.

Steckler, G.R., and Karner G,D., 1988, The distribution of extension with depth in the Newark basin: Geological Association of Canada Programs with Abstracts, v.13, p. A117.

Sutter, J.F. and Smith, T.E., 1979, $^{40}Ar/^{39}Ar$ ages of diabase intrusions from the Newark basins in Connecticut and Maryland: initiation of central Atlantic rifting: American Journal of Science, v. 279, p. 808-831.

Weigand, P. W., and Ragland P. C., 1970, Geochemistry of Mesozoic dolerite dikes from eastern North America: Contributions to Mineralogy and Petrology, v. 29, p. 195-214.

Witte, W.K., and Kent , D.V., 1990, The paleomagnetism of red beds and basalts of the Hettangian extrusive zone, Newark Basin, New Jersey, J. Geophys. Res., 95, 17533-17545.

Witte, W.K., and Kent , D.V., 1991, Tectonic Implications of a remagnetization event in the Newark Basin J. Geophys. Res., 96, 19,569-19,582.

Yoder, H.S., and Tilley, C.E., 1962, Origin of basaltic magmas: An experimental study of natural and synthetic rock systems: Journal of Petrology, v. 3, p. 342-532.

Zen, E-an, Goldsmith, Richard. Ratcliffe, N.M., Robinson, Peter, and Stanley, R.S., 1983, Bedrock geological map of Massachusetts, U.S. Geological Survey, Washington, D. C.

# GLACIAL LANDFORMS AND MORPHOSEQUENCES IN THE ASHUELOT RIVER VALLEY, WINCHESTER TO KEENE, NEW HAMPSHIRE

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## INTRODUCTION

The main purpose of this field trip is to study constructional ice-contact landforms (deltas, eskers and kame terraces) and erosional ice-marginal channels in the lower Ashuelot River valley between Winchester and Keene (Fig. 1), in order to interpret the late-glacial history of the area. Several interrelated subjects that will be addressed include morphosequences, the mode of ice retreat, glacial Lake Hitchcock, glacial Lake Ashuelot and crustal rebound. Field trip stops will be made on the Keene, VT-NH, and Winchester, VT-NH, U. S. Geological Survey 7.5' x 15' quadrangles with a metric scale of 1:25,000.

The Ashuelot River drains the southwest corner of New Hampshire and is a major tributary of the Connecticut River. The Ashuelot River flows southwest and south about 58 km (36 mi) from its headwaters near Washington through Surry and Keene to Winchester where it turns sharply to the west, continues for 8 km (5 mi) and enters the Connecticut River 0.8 km (0.5 mi) south of Hinsdale. From Surry to Ashuelot, the Ashuelot flows mainly on non-resistant gneissic rocks with a gradient of 0.71 m (3.75 ft/mi). Between Ashuelot and Hinsdale the river steepens to 12.0 m/km (63.5 ft/mi) as it cuts across the regional structure.

During deglaciation in the Connecticut valley, the margin of the retreating Laurentide ice sheet was accompanied by a northward-expanding glacial Lake Hitchcock (Lougee, 1939, 1957). Lake Hitchcock developed a stable outlet over a bedrock threshold at New Britain, Connecticut, and drainage down the present-day course of the Connecticut River was blocked by a large ice-contact delta at Rocky Hill, Connecticut. The lake formed during ice retreat when the land was still depressed by the weight of the ice, and it extended 320 km (200 mi) from its spillway to West Burke, Vermont, before uplift due to the removal of the ice sheet commenced.

When the ice margin retreated past the mouth of the Ashuelot River at Hinsdale, Lake Hitchcock extended only 2.4 km (1.5 mi) east of Hinsdale, where a large delta was deposited by drainage from the Ashuelot valley. The projected water plane of Lake Hitchcock intersects the present Ashuelot River only 3.6 km (2.2 mi) east of Hinsdale, therefore, there can be no evidence for the existence of Lake Hitchcock east of that point. Instead, there is evidence for a higher lake known as glacial Lake Ashuelot, which stood about 30 m (98 ft) above Lake Hitchcock and extended 36 km (22 mi) from its outlet near Ashuelot to the vicinity of Surry.

Identification of the controlling threshold or outlet for Lake Ashuelot has been elusive because of lack of knowledge about the precise amount and direction of isostatic rebound following deglaciation and an apparent topographic error on the Keene 15-minute quadrangle at the location of the outlet. Work by Koteff and Larsen (1989) indicates that the former shoreline of Lake Hitchcock now rises toward N21.5°W with a gradient of 0.90 m/km (4.74 ft/mi). When that gradient was applied to deltaic topset-foreset contacts in the Ashuelot basin, the only apparent outlet as shown on the Keene 15-minute quadrangle was too high to control the level of Lake Ashuelot. When the new Winchester, VT-NH, metric quadrangle became available, the same outlet was shown as being 6.1 m (20 ft) lower, falling at the right vertical position to be the controlling threshold for Lake Ashuelot. The threshold, named here the "Ashuelot channel", is located 1.9 km (1.2 mi) N65°W of the village of Ashuelot.

## MORPHOSEQUENCES

A morphosequence consists of all the stratified deposits formed by one meltwater stream system and deposited in a given depositional basin defined by a particular ice margin and outlet from that basin (Koteff, 1974). A typical morphosequence is an ice-contact delta with coarse sediment at its proximal (ice-contact) side that grades downstream to fine sediment at a more distal (lacustrine) side. Retreat of the ice margin from the ice-contact slope marks the end of construction of a given morphosequence. When the ice margin again becomes stationary, or if there is a sudden outburst of meltwater, construction of a new morphosequence may start.

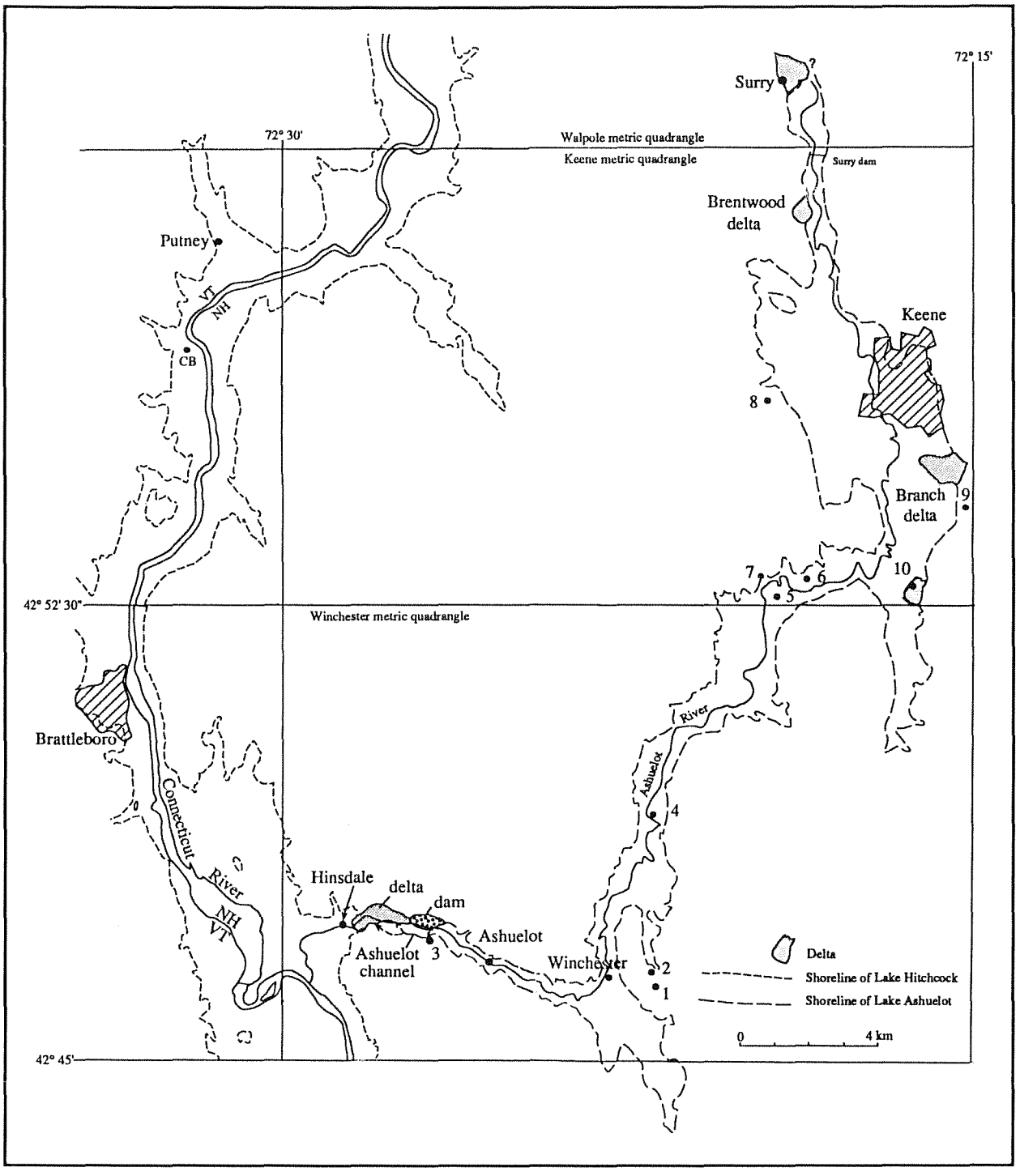


Figure 1. Location of study area in southwest New Hampshire showing relationship of Lake Ashuelot and Lake Hitchcock. Field trip stops are numbered. CB, Canoe Brook section of Ridge and Larsen (1990).

The mapping of surficial deposits by many geologists using the morphosequence concept has documented an orderly generally northward retreat of the margin of the last ice sheet throughout New England (Koteff and Pessl, 1981). As the ice margin retreated northward up the Connecticut valley at Brattleboro, at least 10 ice-contact deltas (morphosequences) were formed sequentially from south to north (Larsen and Koteff, 1988).

For the purposes of this field trip, we consider three general types of ice-contact landforms, ice-contact deltas, kame terraces and eskers, each of which may constitute a single morphosequence or part of a larger morphosequence. These landforms commonly grade laterally into each other and a single morphosequence may contain two or all three of these general types in addition to non-descript stratified deposits. All ice-contact landforms will contain, in some part of their volume, ice-contact stratified drift, which consists of sediment deposited by meltwater streams or currents on, in, under or adjacent to stagnant ice. The criteria used to recognize such deposits are the presence of flowtill interbedded with stratified sediments, collapse structure due to melting of ice below or adjacent to stratified sediments, abrupt change in grain size, large boulders in fine-grained sediment, and layers containing many angular clasts.

An ice-contact delta consists of outwash deposits carried by high-energy meltwater streams that deposit their sediment load in a lake or marine environment adjacent to the ice sheet (Fig. 2). The outwash is mainly sand and gravel deposited in topset, foreset and bottomset beds. The coarsest sediment is usually deposited close to the ice margin while other sediments become finer grained downstream from the ice margin. The coarsest sediment is found in topset beds that are formed by streams and are characterized by fluvial crossbeds and plane beds. Foreset beds are comprised of layers of interbedded pebble gravel, pebbly sand and sand that dip at angles up to 32 degrees. Bottomset beds make up the toe of the delta and constitute the finest deltaic sediment. These beds often grade laterally into fine-grained lake-bottom deposits of silt and clay. The elevation of the highest topset-foreset contact in any delta is significant in that it represents the minimum elevation of the lake into which the delta was deposited.

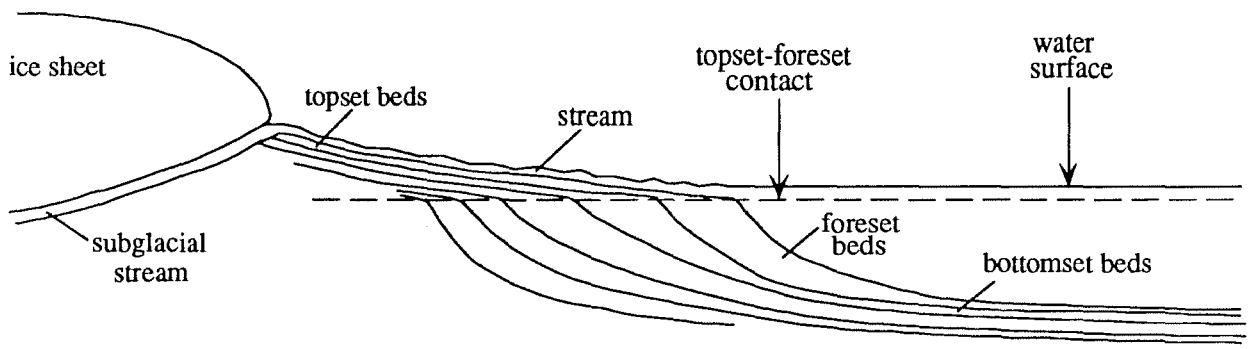


Figure 2. Diagrammatic cross section of ice-contact delta showing topset, foreset and bottomset beds. Note that the topset-foreset contact lies below the level of the lake. Diagram has high vertical exaggeration.

Three ice-contact deltas near West Swanzey were formed sequentially from southwest to northeast near the axis of the valley. They were deposited by meltwater flowing directly from the ice and were graded to a single lake, Lake Ashuelot, whose outlet was controlled by the Ashuelot channel. One of these, the West Swanzey delta at Stop 5, has a topset-foreset contact, which was measured to be 149.5 m (490.4 ft) ASL (above sea level).

A second type of ice-contact landform that is well represented in this area is the kame terrace. These are flat-topped landforms composed of ice-contact stratified drift that was deposited in a depression or valley formed between a mass of stagnant ice and a till or bedrock slope. The mass of ice commonly is a tongue or lobe that occupies the valley floor, and the till or bedrock slope commonly is the side of a valley. A kame terrace is often linear, and its upper surface slopes in the direction of meltwater drainage at the time of formation. Sediments deposited in contact with the ice subside and collapse downward toward the valley as the ice melts back. If there are stillstands as the ice margin retreats down the slope, a series of kame terraces can be formed with the older terrace standing highest on the slope and the youngest at the bottom.

Figure 3 illustrates just such a series of ice-contact landforms located on the east side of the Ashuelot valley south of Keene. A vertical line drawn anywhere on the diagram will intersect oldest deposits at the top and youngest at the bottom. For example, a vertical line drawn above the middle of Wilson Pond intersects wpg-1 (oldest) at the top, wpg-3 in the middle and ag-4 (youngest) at the bottom. This represents the order of formation of the landforms (morphosequences, in this case) as the ice margin retreated westward down the east slope of the Ashuelot valley. Similarly, a horizontal line drawn on the lower part of the diagram will intersect oldest deposits at the right (south) and the youngest at the left (north), which represents northward retreat of the ice margin in the Ashuelot valley.

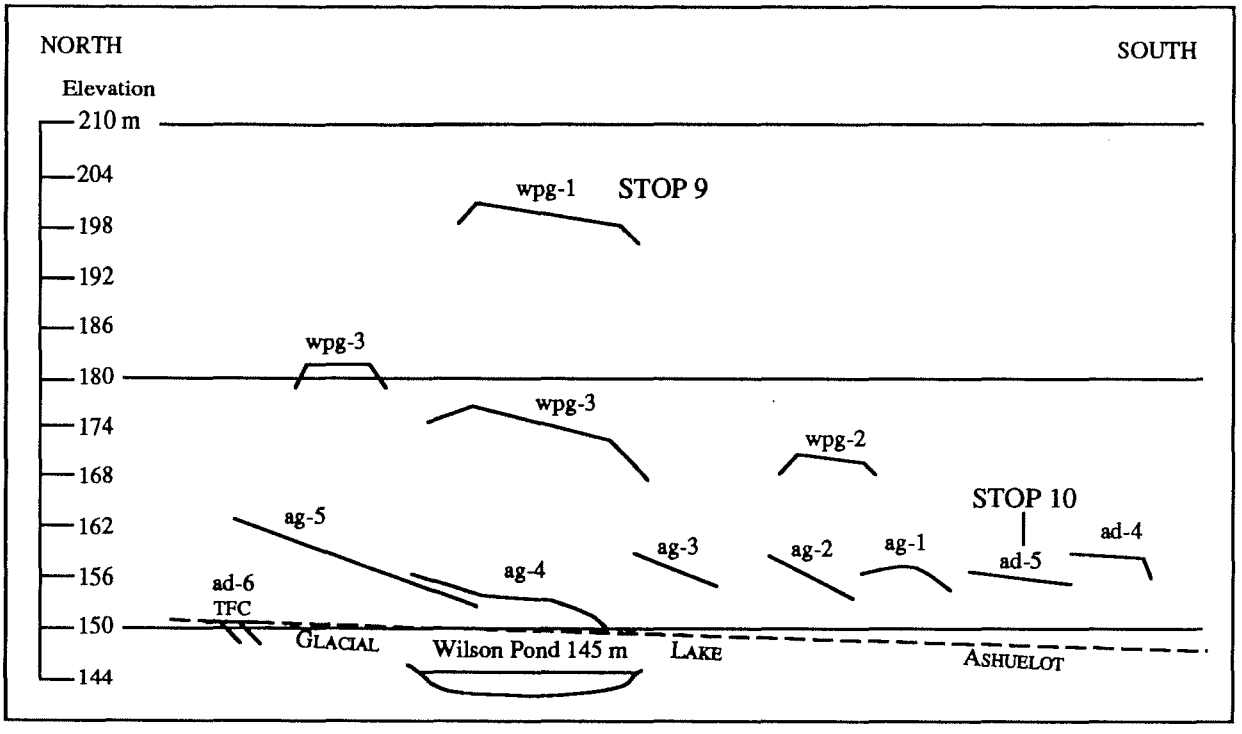


Figure 3. Vertical north-south projected profile of kame terraces and ice-contact deltas located on the east side of the Ashuelot valley south of Keene. Deposits occur in four groups: (A) wpg-1 to wpg-3, Wilson Pond glacio-fluvial deposits (kame terraces) on fluvial grade to control points above the level of Lake Ashuelot, (B) ad-4 to ad-5, ice-contact deltaic deposits graded to Lake Ashuelot and formed sequentially from south to north, (C) ag-1 to ag-5, glaciofluvial deposits (kame terraces) graded to Lake Ashuelot and formed sequentially from south to north, (D) ad-6 TFC, delta with topset-foreset contact at 150.3 m (493.1 ft) ASL that represents a fluvial (non-ice-contact) deposit graded to Lake Ashuelot.

Eskers, a type of ice-channel filling, constitute a third type of ice-contact landform that occurs in the Ashuelot valley. Ice-channel fillings are narrow elongate ridges composed of ice-contact stratified drift deposited by meltwater streams. They are divided into eskers, which are formed in subglacial and englacial streams, and crevasse fillings, which are formed in open air channels between and around stagnant blocks of ice (Jahns, 1953). An esker at Stop 7 is 212 m (700 ft) long, 15 to 23 m (50 to 75 ft) wide and up to 4.5 m (15 ft) high. It is a sinuous ridge of pebbly sand that obliquely climbs the slope from northeast to southwest and trends toward a low divide on the valley wall. The meltwater stream that deposited the esker flowed uphill under hydrostatic pressure in a subglacial tunnel. At the time of esker formation, the ice margin must have been located at least as far upslope as the furthest extent of the esker, and probably had a north-south trend on the west side of the valley. A small esker that was formed under similar conditions can be seen at Stop 8.

The above-described ice-contact landforms are all constructional in nature, that is, they represent positive depositional elements in the landscape. Next, we consider ice-marginal channels that represent negative erosional elements in the landscape. Ice-marginal channels are carved mainly in till or bedrock slopes by meltwater streams flowing along or just under the edge of the stagnant ice sheet. As the ice margin retreats down a slope, a succession of stream channels may be formed when a meltwater stream held against the valley wall is permitted to drop to lower

levels. By definition, an erosional feature cannot be a morphosequence, but a channel can often be correlated with a specific morphosequence. A series of three ice-marginal channels can be observed at Stop 8, where the ice margin retreated to the north and down the slope (Fig. 4). Small morphosequences located to the west can be correlated with each channel, but are not shown on Figure 4.

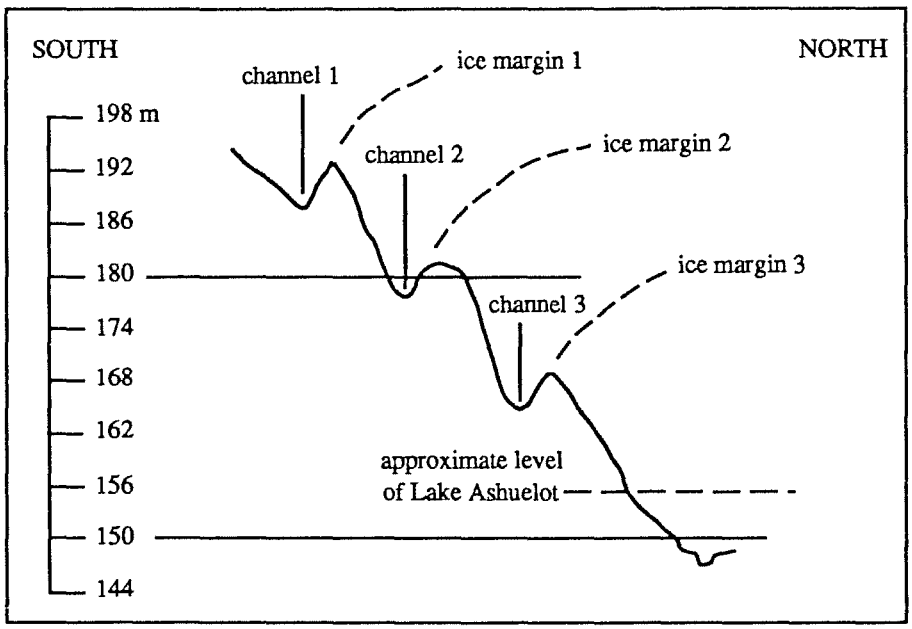


Figure 4. Diagrammatic north-south projected profile of ice-marginal channels located at Stop 8 just west of Keene Country Club. Successive channels were carved by marginal or subglacial streams as the ice margin retreated northward down the slope. Stream flow was toward observer.

## MODE OF ICE RETREAT

Deglaciation of the Connecticut Valley of Massachusetts was by an active lobe of ice that readvanced several times leaving till over deformed and thrust-faulted outwash deposits (Larsen and Hartshorn, 1982). The activity of the former lobe is also indicated by a nearly 180-degree radial pattern of striations stretching across the valley and by the distribution of erratics of Jurassic-Triassic rocks transported both southeast and southwest of their source area in the Connecticut Valley. A compilation of striations in New England by James Goldthwait (Flint, 1957, p. 60) shows southwest and west-southwest striations located west of the Connecticut River in Massachuetts, but on the west side of the Connecticut valley north of Putney, Vermont, no strongly west-trending striations are shown. The evidence from the Goldthwait map suggests to me that, when the ice sheet retreated north of Putney, it lost its ability to spread out as an active lobe, as it had done further to the south.

In the Brattleboro area, striations occur in two modal groups (Larsen and Koteff, 1988). One is S5°E to S5°W and the other is S20°W to S35°W. Striations that trend due south are cut by a younger set trending S40°W just east of the Route 9 bridge over the Connecticut River. In addition, granitic erratics from the Black Mountain pluton, 8.8 km (5.5 mi) north-northwest of Brattleboro, have been transported both south and southwest. The data from striations and indicator clasts can be interpreted to indicate two separate phases of glacial movement. The first phase was characterized by essentially due south movement during glacial maximum. The second phase occurred in late-glacial time when an active, but waning, ice lobe retreated northward in the Brattleboro area.

Striations mapped by Ridge (1988, Fig. 3) in the Walpole and southern Bellows Falls metric quadrangles trend S5°E to S15°E in uplands east of the Connecticut River, and S10°W to S25°W near the Connecticut River. Ridge concluded that ice flow was parallel to the trend of the Connecticut valley and structurally-controlled bedrock ridges in the western part of the Walpole quadrangle. It should be noted that after flowing due south in the western part of the Walpole quadrangle, the Connecticut River turns to S50°W in the Keene metric quadrangle and flows for 8 km

(5 mi) before turning again to the south (Fig. 1). In that area, the northwest corner of the Keene quadrangle, striations trend between S5°W and S30°W. It appears that a well developed radial pattern of striations due to retreat of an active lobe of ice is lacking in both the Walpole quadrangle and the northwest corner of the Keene quadrangle. In the eastern half of the Keene metric quadrangle, striations have not been well mapped and are not well preserved on granitic gneiss underlying the Ashuelot basin. The three striations that have been mapped in the eastern half of the Keene quadrangle trend between S12°E and S13°W, and hardly constitute a sample large enough to draw any conclusion about the mode of ice retreat in the Ashuelot basin.

A change in the overall budget of the retreating ice sheet when its margin was near Putney, as suggested by the ice-directional features, can also be seen in the evidence from morphosequences (Larsen and Koteff, 1988). In addition to the evidence from striations and indicator clasts for an active lobe in the Brattleboro area, at least ten ice-contact deltas (morphosequences) were formed sequentially from south to north as the ice margin retreated in the same area. What is unusual is the fact that near the axis of the Connecticut valley north of Putney there are very few morphosequences built directly from the ice. North of Putney, there is no place that matches the number and close spacing of morphosequences as can be observed at Brattleboro. Numerous small morphosequences in the Ashuelot valley at Keene suggest that ice activity was moderate and similar to that during retreat at Brattleboro.

## GLACIAL LAKE HITCHCOCK

During retreat of the ice sheet in the Connecticut valley, the ice margin was accompanied by a northward-expanding glacial Lake Hitchcock (Fig. 1) (Lougee, 1939, 1957). Lake Hitchcock developed a stable outlet over a bedrock threshold at New Britain, Connecticut, and drainage down the present-day course of the Connecticut River was blocked by a large ice-contact delta at Rocky Hill, Connecticut (Stone and others, 1982). The lake formed during ice retreat when the land was still depressed by the weight of the ice, and it extended 320 km (200 mi) from its spillway to West Burke, Vermont, before uplift due to the removal of the weight of the ice sheet commenced.

Lougee's (1957) projected profile of the levels of Lake Hitchcock and "Lake Upham" shows uplift toward N15°W, very close to the probable true value. When I measured the elevation of 11 topset-foreset contacts in ice-contact deltas between Brattleboro and Lisbon, NH, it became apparent that Lake Hitchcock extended to West Burke and Gilman, VT, and Littleton, NH, and was similar in those areas to the map of Lake Upham by Lougee (Larsen, 1984). Based on the 11 data points in Vermont and New Hampshire plus 4 others in Massachusetts, it was noted that the best-fit plane for the uplifted shoreline rose toward N20°W with a gradient of 0.86 m/km (4.54 ft/mi). That study has since been extended and now includes 28 data points on or near the Lake Hitchcock water plane, which is now thought to rise toward N21.5°W with a gradient of 0.90 m/km (4.74 ft/mi) (Koteff and Larsen, 1989). This represents the amount of crustal rebound that occurred in western New England since the retreat of the last ice sheet.

Having established the position of the Lake Hitchcock water plane in space by this latter study, it became clear that Lake Hitchcock theoretically could have extended only 3.6 km (2.2 mi) east of Hinsdale in the Ashuelot valley. Since portions of a large delta built directly into Lake Hitchcock by drainage from the Ashuelot basin extend from 0.6 to 2.1 km (0.4 to 1.3 mi) east of Hinsdale, from a practical point of view Lake Hitchcock only extended 2.1 km (1.3 mi) east of Hinsdale before it was driven back by the construction of the delta.

## GLACIAL LAKE ASHUELOT

Lake-bottom deposits of Lake Ashuelot have been excavated in Keene for the purpose of making bricks since 1750 A.D. or before. On a "Sketch of Keene-1750" in the Annals of Keene, "clay pits" are shown just east of the north end of Main Street. On maps of Keene made during the 1800's, brickyards are shown in various places along Beaver Brook between Roxbury and Marlboro Streets. In the summer of 1986, varved silt and clay was exposed in a shallow excavation for a flood-control project adjacent to Beaver Brook just south of Water Street. Of the few varves measured, total thickness of individual varves was 12 to 15 mm (0.5 to 0.6 in) of which clay comprised 2 to 3 mm (0.09 to 0.13 in). Today, a large abandoned clay pit exists at the south end of the Keene State College campus as Brick Yard Pond, where Antevs (1922) measured a short section of 76 varves that he correlated with his New England varve chronology (NE varve 5804-5879). Antevs' varve chronology arbitrarily begins with NE varve 3001 and ends with NE varve 7400.

In the summer of 1987, Jack Ridge and I measured the thickness of 645 individual varves or couplets at Canoe Brook 3.6 km (2.2 mi) south of Putney (Fig. 1) (Ridge and Larsen, 1990). Of the 645 varves measured, the lowest

545 had been deposited in Lake Hitchcock and correlate very well with NE varves 5687-6231 of the New England varve chronology (Antevs, 1922). In addition to proving that Antevs' technique of varve correlation was valid at least for Lake Hitchcock, we found organic material that has been radiocarbon dated at between 12,300 and 12,950 B.P. thereby tying radiocarbon dates to the New England varve chronology for the first time. By chance, the Canoe Brook section also includes that portion that Antevs correlated with the 76-varve section at Keene. By extrapolation from the layers containing the dated organic material, we can estimate that the varves that Antevs measured at Keene were deposited in Lake Ashuelot approximately 12,600 to 12,700 B.P..

Thick sequences of lake-bottom sediments have been recorded in numerous borings made in the Keene area. The thickness of clay in these borings ranges from 9 to 38.7 m (30 to 127 ft) in the area just west and south of Keene. The figure of 38.7 m (127 ft) is suspect because a nearby well of similar depth has only 16.8 m (55 ft) of clay. Some well records are better than others, but values up to 18.3 m (60 ft) are commonly recorded. In six well-documented borings by Haley and Aldrich (1976) at Brick Yard Pond, Keene State College, "varved clay, believed to have been deposited in an ancient glacial lake bed", averages 10.4 m (34 ft) in thickness. The point to be made here is that there are more than 9 m (30 ft) of varved silt and clay below the short section of 76-varves that Antevs measured at the surface in 1922.

In 1986, Carl Koteff and I measured the topset-foreset contact in three deltas that were deposited in Lake Ashuelot. The three deltas and the elevation of their topset-foreset contacts from north to south are: (A) Brentwood ice-contact delta, 529.8 ft (161.5 m) ASL, (B) Branch non-ice-contact delta, 493.1 ft (150.3 m) ASL, and (C) West Swanzey ice-contact delta (Stop 5), 490.4 ft (149.5 m) ASL (Fig. 1). The three elevations define a plane that rises 1.39 m/km (7.35 ft/mi) toward N37°W and is not parallel to the Lake Hitchcock water plane. It should be parallel, given the fact that the latter was established with many more data points and over a much greater area. Any measurement of crustal rebound in the Ashuelot basin should be the same as that for the Connecticut basin as a whole unless neotectonics is at work. The problem lies in the lack of data points and the fact that the three data points are close together. A fluctuation in lake level of one or two meters can change the elevation where a topset-foreset contact forms thus drastically altering the orientation of a water plane based on just three data points. In addition, the Brentwood and West Swanzey deltas are ice-contact deltas formed directly at the ice margin, whereas the Branch delta is a non-ice-contact delta fed by a permanent stream and could have been built over a long period of time while erosion occurred at the outlet.

Using the parameters of the Lake Hitchcock water plane, a projected profile similar to Figure 5 was drawn using the Keene 15-minute quadrangle (scale 1:62,500) with a 20-foot contour as a base. When shorelines were extended to

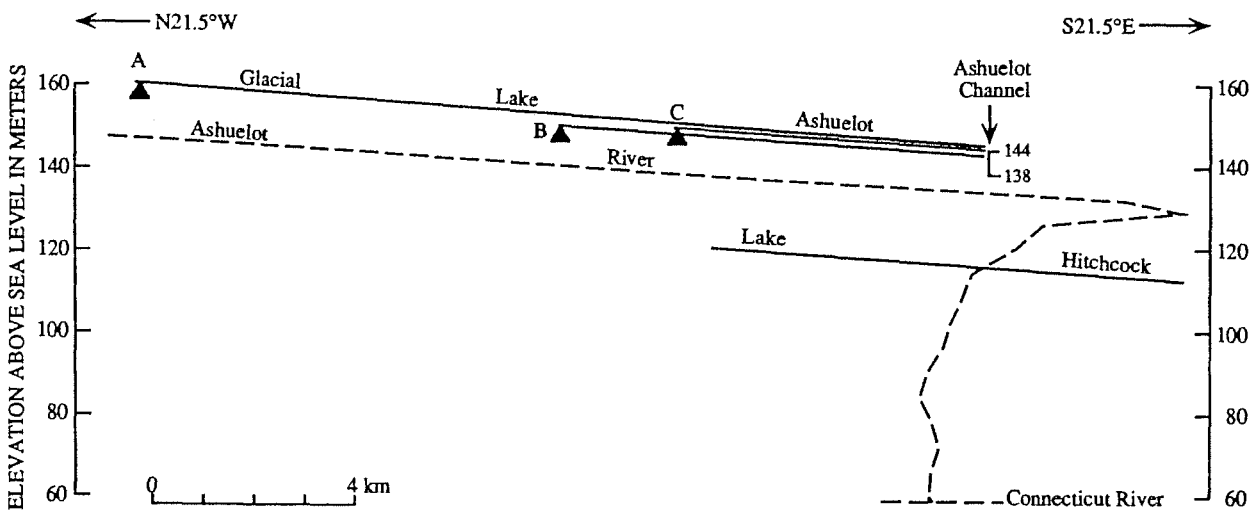


Figure 5. Vertical projection looking N68.5°E perpendicular to the direction of uplift and showing spacial relations between tilted water planes of Lake Ashuelot and Lake Hitchcock (solid lines) and the longitudinal profile of the Ashuelot River (dashed line). Solid triangles represent measured topset-foreset contacts at (A) Brentwood delta, (B) Branch delta, and (C) West Swanzey delta (Stop 5). Plot made from the Keene and Winchester metric quadrangles.

the south and parallel to that of Lake Hitchcock from each of the three topset-foreset contacts, the only obvious possible control point, the Ashuelot channel, was too high falling between 146.3 and 152.4 m ASL (480 and 500 ft). Later, when the new Winchester metric quadrangle became available, the Ashuelot channel was between 138 and 144 m ASL (452.6 and 472.3 ft) and at the right elevation to serve as a possible stable threshold at least for an early phase of Lake Ashuelot, that is the lake into which the three deltas were deposited. Figure 5 shows that when shorelines are projected south parallel to the gradient of Lake Hitchcock from each of the three topset-foreset contacts, the lake that they represent, Lake Ashuelot, could have drained out through the Ashuelot channel at an approximate elevation of 144 m (472 ft).

As seen on the Winchester metric quadrangle, the Ashuelot channel is sinuous and is about 1.0 km (0.62 mi) in length (Fig. 6). The Ashuelot channel is the only apparent marginal channel on the south side of the Ashuelot valley between Hinsdale and Winchester. The channel drops in elevation from its threshold at the eastern end of the channel and ends at the west on a terrace underlain by pebble-cobble gravel that I interpret to be part of the delta built into Lake Hitchcock (Fig. 1). Coincidently, it falls at the right elevation, about 144 m (472.3 ft), to have served as the controlling threshold for Lake Ashuelot. If the Ashuelot channel did serve as the threshold for Lake Ashuelot, then the Ashuelot valley immediately north of the channel had to be dammed with glacial drift as shown in Figure 1. Also, the valley to the east of the dam probably would have been at least partially filled with glacial drift. If the Ashuelot channel did not serve as the control for Lake Ashuelot, then location of the former threshold is not obvious, but probably existed near the village of Ashuelot and has been removed by erosion.

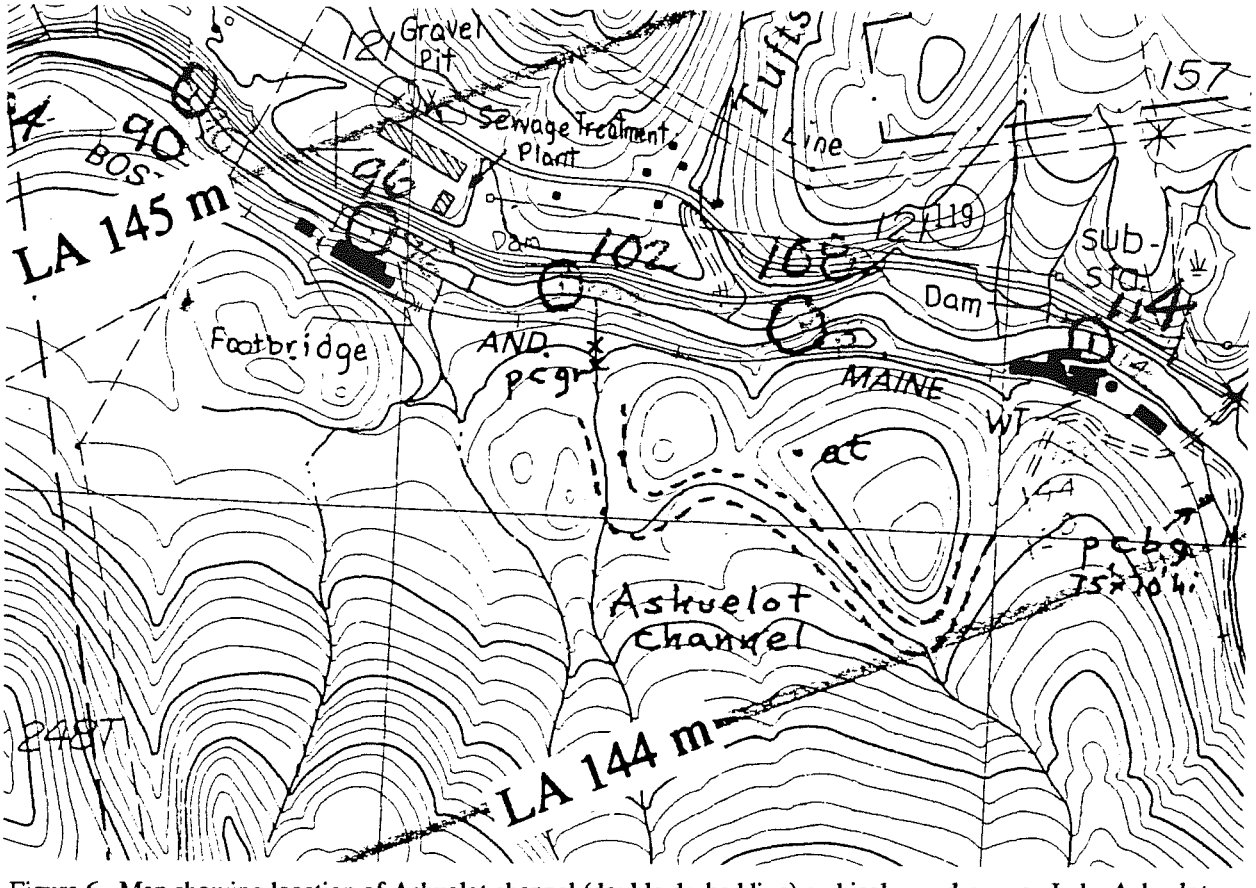


Figure 6. Map showing location of Ashuelot channel (double dashed line) and isobases drawn on Lake Ashuelot water plane (LA 144 m and LA 145 m). Note that the 144 m isobase intersects the 144 m contour at the southeast edge of the channel. at, ablation till; pcgr, pebble-cobble gravel; labeled circles represent elevation in meters of contours crossing the Ashuelot River; scale 1 in = 315 m (1032 ft), contour interval = 6 m (20 ft).

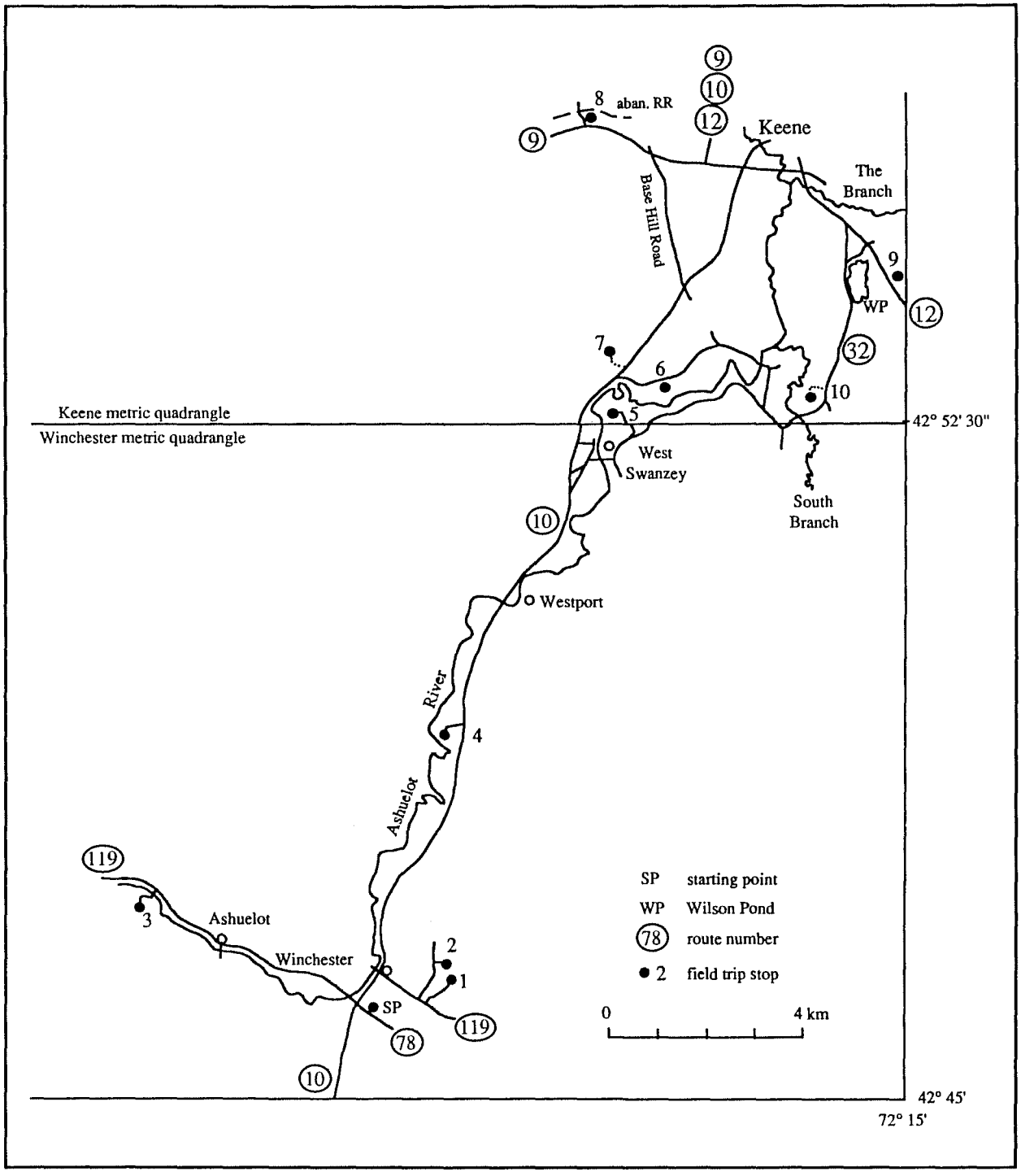


Figure 7. Location of field trip stops

## ACKNOWLEDGEMENTS

Detailed mapping of the Keene metric quadrangle was supported by the New Hampshire Geological Survey directed by Eugene Boudette. Jack Ridge and Carl Koteff supplied information to this paper through past discussions and related studies. Maureen Larsen proofread several drafts of the paper and suggested improvements. David Westerman and Jim Reynolds helped with computer-related problems. Norwich University provided logistical support. I express my sincere thanks to each of the above for their assistance.

## ROAD LOG

The starting point is located on the Winchester metric quadrangle and is the parking lot behind the CITGO gas station on the north side of Route 78, 0.1 of a mile east of a red light at the junction of Routes 10, 119 and 78 (Fig. 7). The red light is 0.55 of a mile southwest of the village of Winchester, NH, and about 32 miles north of Amherst via Routes 63 and 10.

**Mileage**

0.0 Road log begins in parking lot. Leave parking lot, turn right (west) on Route 78
0.15 Turn right at red light, proceed north on Routes 10 and 119
0.7 Turn right at red light in center of Winchester, proceed east on Route 119
1.4 Turn left (north) on dirt road along east side of Evergreen Cemetery, the south slope of the cemetery is the foreset slope of the ice-contact delta to be studied at Stop 1
1.7 Enter pit, kettle pond on right, proceed to north end of pit at 2.0 miles

### STOP 1. KENNETH PERRY PIT IN ICE-CONTACT DELTAIC DEPOSITS

Deltaic deposits in the Perry pit were formed in a temporary proglacial lake that stood approximately 12 m (40 ft) or more above the projected level of Lake Ashuelot. The deposits were formed as a series of ice-contact deltas by meltwater flowing out of the ice when the ice margin was retreating northward through the area of the Perry pit (Fig. 8). A great variety of sedimentary structures are exposed along the 0.25 mile-long west side and the north side of the Perry pit. Foreset beds, dune bedding, ripple-drift cross-lamination and imbricated pebbles in topset beds all indicate southward transport of meltwater during construction of the ice-contact deltas. A 5.5-meter (18 ft) measured section located just northeast of the symbol for Stop 1 on Figure 8A consists of 1.8 m (6 ft) of flat-bedded and crossbedded deltaic topset beds overlying 3.7 m (12 ft) of truncated foreset beds. Grain size of the topset beds ranges from pebbly coarse sand to pebble gravel with cobbles. The foreset layers are fine sand to pebbly coarse sand and dip toward S5°E to S23°E. At the south end of the west face are deformed fine-grained lake-bottom deposits resting on foreset beds.

2.6 Turn right (west) on Route 119
2.7 Turn right (north) on Forest Lake Road and climb foreset slope of Evergreen delta
2.9 Topset surface of Evergreen delta on the right
3.25 Turn right (east) into Winchester "recycling center", bear right and descend to pit floor

### STOP 2. LAKE-BOTTOM DEPOSITS AT WINCHESTER RECYCLING CENTER

Following northward retreat of the ice margin from the head of ice-contact deltas located at the north end of the Perry pit, the site at Stop 2 (Fig. 8) experienced rapid sedimentation on the floor of a glacial lake, or possibly in a tunnel or open channel. Due east of the symbol for Stop 2 (Fig. 8A) are 3.7 m (12 ft) of flat-bedded layers of medium to very fine sand with minor silt. The layers are in depositional sequences similar to those in Lake Hitchcock deltaic deposits described by Gustavson and others (1975). A typical sequence is 0.3 to 0.4 m (1.0 to 1.25 ft) thick and has plane bed at the bottom changing upward into A-type ripple-drift cross-lamination, which in turn grades upward into B type. The angle of climb in the B type increases upward. The layers were deposited as deltaic bottomsets or as prodelta lake-bottom deposits. Southeast of the symbol for Stop 2 (Fig. 8A), a broad channel cut into the flat-bedded layers is filled with south-dipping foreset beds.

Above the middle of the flat beds on the east is a layer 0.3 m (1.0 ft) thick with a series of steeply dipping, imbricate slump folds with sharp crests many of which are offset by thrust faults. Axial planes of the slump folds dip at angles up to 67°N. The folds slid on a detachment surface that is marked by truncated beds of silt.

3.3 Turn left (south) on Forest Lake Road
3.8 Turn right (west) on Route 119

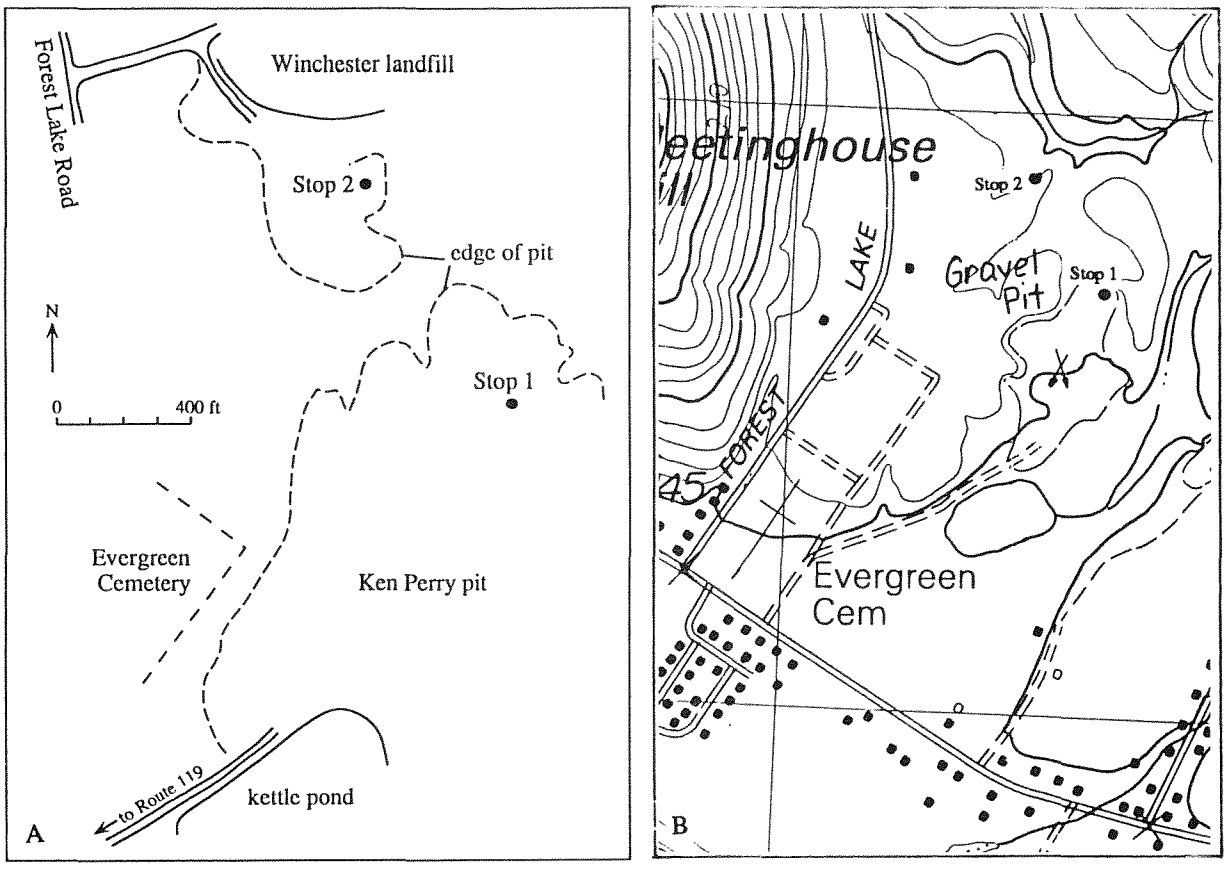


Figure 8. Location maps for Stops 1 and 2, (A) Pace-and-compass map showing approximate position of pit edges as of June, 1992, (B) Portion of Winchester metric quadrangle, scale 1 in = 315 m (1032 ft), contour interval = 6 m, dark contour in Evergreen Cemetery is 150 m ASL.

4.4 Turn left at red light in the center of Winchester, proceed south on Routes 10 and 119
5.0 Turn right at red light, proceed west on Route 119
5.1 Cross Ashuelot River
7.0 Blinking light in Ashuelot village
8.1 Turn left (south) on single lane bridge over Ashuelot River, continue straight ahead and bear slightly to the right. If at 8.25 miles a blue-pipe gate is open, proceed on single lane, dirt road. If not, park and proceed on foot
8.4 Company picnic area on the right
8.45 Woods road on right with yellow "No Trespassing" sign leads to Stop 3 (Fig. 9)
8.5 Park near settling pond and off dirt road. Follow the woods road southwest about 400 feet to the Ashuelot channel. The channel where the road crosses is about 100 ft wide.

**STOP 3. THE "ASHUELOT CHANNEL"**

See Figure 6 for topographic map of Ashuelot channel as described in the text. Southward projection of a water plane parallel to that of glacial Lake Hitchcock from ice-contact deltas in the Keene area led to the identification of the Ashuelot Channel as the only probable stable outlet for glacial Lake Ashuelot.

8.9 Cross bridge and turn right. Proceed east on Route 119
12.0 Turn left (north) at red light
12.5 Proceed north on Route 10 at red light in Winchester
15.6 A deltaic topset-foreset contact formerly was exposed on the west side of the highway in 1982 just south of "Trader John's"
16.0 Gray ablation till on the right

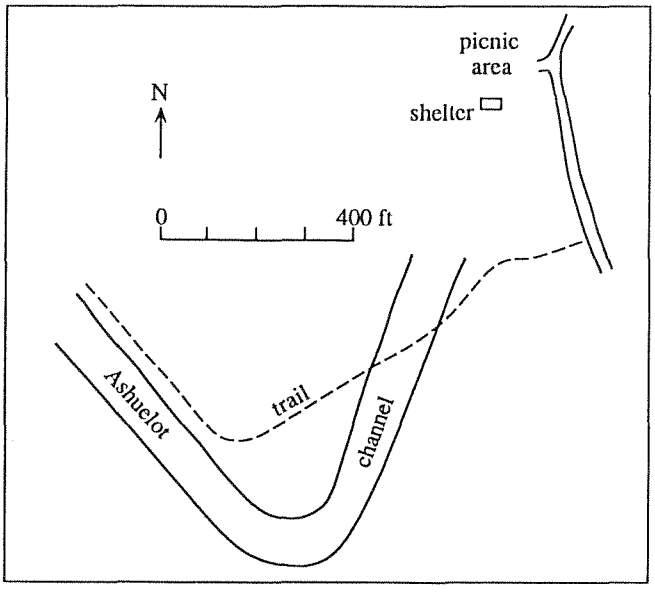


Figure 9. Sketch map of trail leading to the Ashuelot channel. Slabs of rottenstone developed on granitic gneiss of the Ashuelot pluton are common at the sharp bend in the Ashuelot channel.

16.1 Turn left (west) on Kapper Drive at sign "Green Valley Mobile Home Park". Bear left at bend in road and proceed south to large pile of sand

**STOP 4. ESKER BEAD (?)/INCIPIENT DELTA (?)/MORPHOSEQUENCE (?)**

This small hill is shown on the Winchester metric quadrangle as a closed contour with an elevation of 144 m ASL and it lies on the 145-meter isobase of Lake Ashuelot. At one time the hill was over 6 m (20 ft) high. Two basic materials make up the hill, fine to medium sand with ripple-drift cross-lamination makes up the base and pebble gravel is at the top. I infer that the deposit formed where a meltwater stream flowing in a subglacial tunnel emptied into Lake Ashuelot, built up a mound of sand, and just as the mound broke the surface of the lake with the deposition of pebble gravel, the stream was diverted by the melting of ice. Two similar mounds can be seen on the low ground west of Ware Mountain Farm just north of Stop 4.

16.8 Turn left, proceed north on Route 10. Look west just beyond Ware Mountain Farm to view two small mounds similar to the one at Stop 4
17.85 Striated bedrock in bushes at right
18.4 Cross Ashuelot River and rise up on ice-contact delta. Nearly depleted pit on the west formerly had an exposed deltaic topset-foreset contact
18.7 Top of delta, note that the present surface drops on the north, ice-contact, side of the landform
19.1 Sand pit on right
19.35 Bedrock exposed on both sides of Route 10
20.3 Just beyond pond on left, turn right off Route 10 at store "Antique Quest", if you miss the turn take next right turn. Flat surface to next turn is 3 to 5 m below Lake Ashuelot water plane and is a post-lake, possibly post-rebound, terrace of the Ashuelot River
20.75 Turn right (east) on Denman Thompson Avenue
20.95 Cross Ashuelot River
21.1 Turn left (north) on Homestead Avenue
21.25 Straight (north) on Railroad Street at 3-way stop
21.4 Straight ahead (north) where Railroad Street bends to the right
21.6 Sharp bend to the left
21.9 Turn left, proceed west behind Swanzey town garage and enter small pit

**STOP 5. SWANZEY TOWN PIT IN THE WEST SWANZEY ICE-CONTACT DELTA**

The pit is in a small delta complete with topset, foreset and bottomset beds. A 17.5-ft section measured on the south face consists of 2 ft of topsoil and orangish-brown to yellowish-brown fine sand with a few pebbles, 6 ft of fine to medium sand with medium-scale crossbeds and 9.5 ft of flat-bedded and laminated silt, very fine to fine

sand and minor clay. The section is interpreted to consist of stream-terrace deposits overlying bottomset beds. In the past, the 26.5-ft north face has displayed 3.5 ft of flat and crossbedded pebble gravel overlying pebbly fine to coarse sand in southwest-dipping deltaic foresets, which are also exposed in the far northwest corner of the pit. The topset-foreset contact was measured to be 490.4 ft (149.5 m) ASL and is believed to fall on the tilted water plane of glacial Lake Ashuelot. Therefore, the deposits associated with this pit constitute a micromorphosequence that was formed as an ice-contact delta by a meltwater stream flowing from the ice margin directly into Lake Ashuelot.

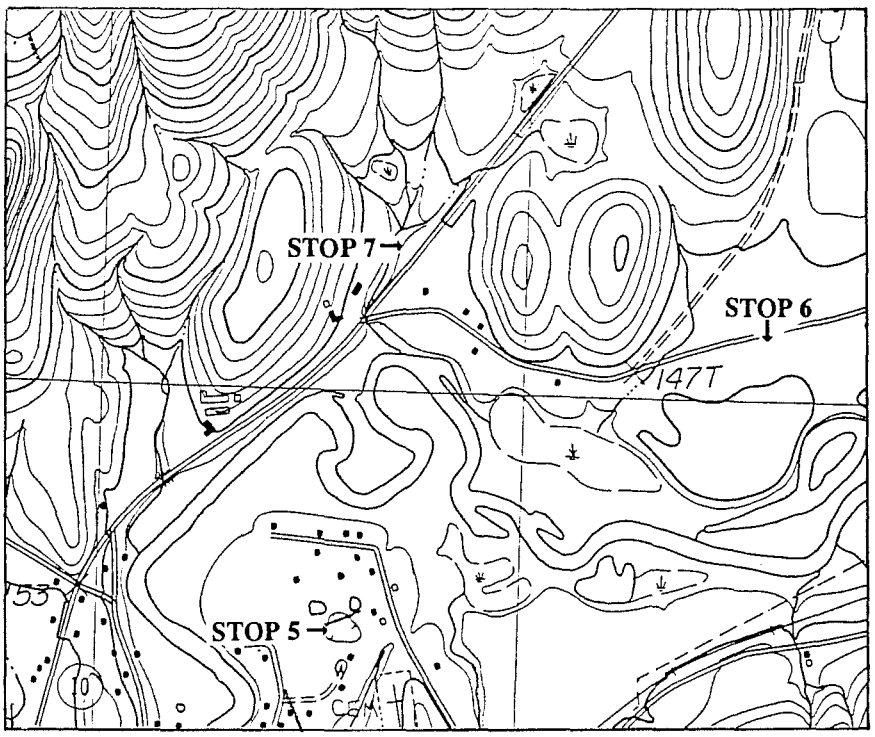


Figure 10. Portion of the Keene metric quadrangle showing the location of Stops 5, 6 and 7. Scale 1 in = 451 m (1480 ft), contour interval = 6 m (20 ft), dark contour south of Stop 6 is 150 m ASL.

22.1 Turn right (south) after leaving pit, retrace route, continue straight ahead (south) at yield sign and at 3-way stop. Turn right (west) on Denman Thompson Avenue, cross Ashuelot River
23.2 Turn right at stop sign. Proceed north, road has right-angle bend to the west
23.7 Turn right, proceed north on Route 10
24.8 Turn right, proceed east on Sawyers Crossing Road
25.5 Park on right side of road at third turnout with cable and logs blocking woods road. Follow flagging on woods road

**STOP 6. DELTA WALK**

There are no exposures on this forested landform on property belonging to Yale University. The landform is an elevated flat-topped surface that slopes gently to the west-southwest and measures about 1,400 ft east-west and 800 ft north-south. A test pit on the north side of the feature has 3 ft of pebble-gravel with cobbles and one on the south side has 2 ft of pebble-gravel overlying medium to coarse sand. This is an opportunity to walk on what is interpreted to be the remains of an ice-contact delta (a morphosequence) built directly into Lake Ashuelot by meltwater streams issuing from an ice margin that wrapped around the eastern end of the feature. The distance between Stop 5 and the northeast ice-contact end of this flat-topped landform is 0.8 of a mile, which I interpret to be the distance that the ice margin melted back in one year. A similar kind of landform lies to the east-northeast with a possible ice-contact slope located 0.6 of a mile from the one at this site.

Return to cars, make a U-turn and proceed west on Sawyers Crossing Road
26.15 Turn right (north) on Route 10

26.35 Turn left (west) into logging area if gate is open, or park on the east side of Route 10 if it is not. Please use extreme caution in crossing Route 10. Follow the trail marked with flagging that leaves the west side of the clearing (Fig. 11)

**STOP 7. ESKER WALK**

The total walk is about 0.9 mi in length and consists of moderate hiking through woods with some wet spots. This is an opportunity to walk on a small ice-contact landform that is pristine and unaltered by man. The start as shown on Figure 11 is at the contact of a steep bedrock slope partially covered with angular slide blocks (talus) and the head of an elongate mound that may or may not be part of the esker. The esker proper starts 100 ft to the southeast. It trends due south and is essentially flat for 200 ft before turning to the southwest and dropping to a gap beyond which it is flat for 100 ft. It rises steeply to the west, then flattens turning to the southwest, and again steepens and climbs westerly before leveling off to the southwest. I infer that the esker formed in a subglacial tunnel in which meltwater under hydrostatic pressure flowed uphill to the southwest. The ice margin at the time of formation of the esker was located at least as far west as the furthest extent of the esker and probably maintained a north-south position while the esker was formed. Meltwater from the subglacial tunnel probably drained over a low divide located about 400 ft southwest of the esker (Fig. 11). A small kame terrace is perched on the bedrock slope just west of the esker. Lacking good elevation control in this area, it is not clear whether the kame terrace formed before, during or after the esker. It appears that the kame terrace is below the divide and probably was deposited after the esker was formed.

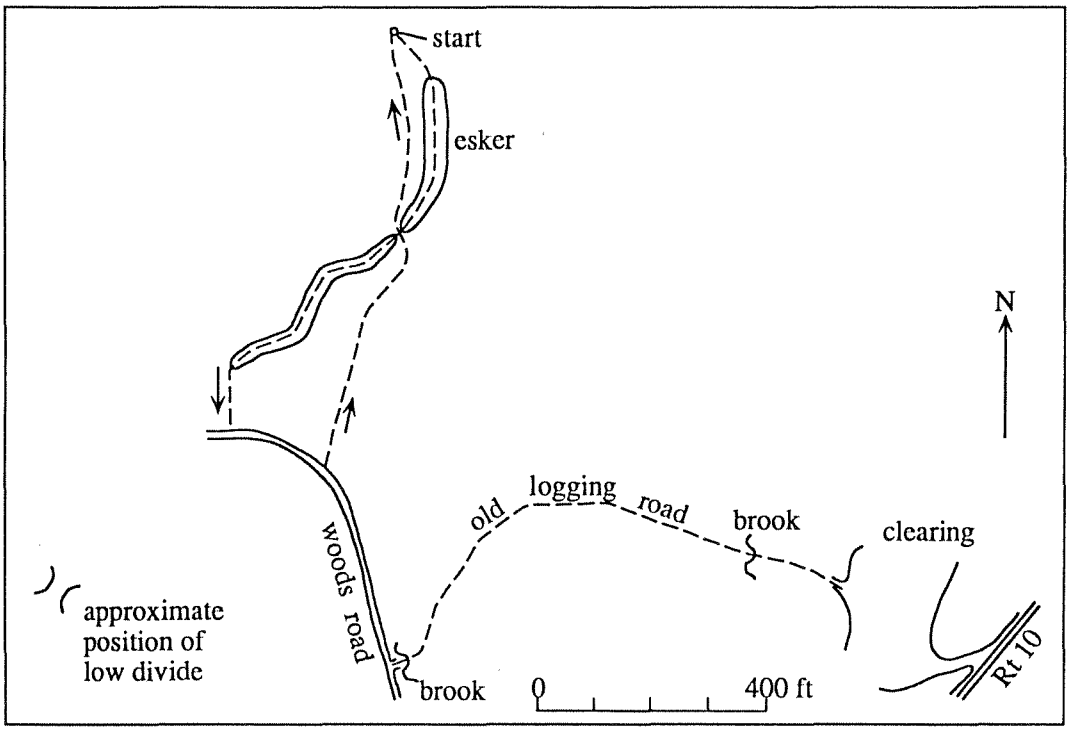


Figure 11. Pace-and-compass map showing location of Stop 7, esker walk, northwest of Route 10

26.45 Turn left (north) on Route 11
26.7 Striations on bedrock at right trend due south to S13°W. Roadcuts for next 0.7 mi mostly are in compact till
27.65 Turn left (northwest) at blinking light on Base Line Road. After descending and passing the highway garage, Base Line Road is within 50 to 100 ft of the shoreline of Lake Ashuelot until we turn left on Route 9
27.8 Pebble gravel in kame terrace at left
29.4 Turn left (west) on Route 9 with caution. Pebble gravel exposed in kame terrace on left after turn
30.3 Turn right (north) on Whitcomb's Mill Road
30.55 Park on right at abandoned railroad. Walk back up road to warehouse (open rectangle, Fig. 12) and

follow flagged trail along south side of warehouse and and into woods along north side of highest channel.

**STOP 8. ICE-MARGINAL CHANNELS**

The total walk is about 1.0 mi in length and consists of moderate hiking through woods with some wet spots. See Figure 4 for projected profile of ice-marginal channels at this stop. This is an excellent locality to document successive ice-margin positions that formed during retreat of the last ice sheet. It always a problem with ice-marginal channels to determine whether they formed at or under the edge of the ice. Morphosequences that lie to the west and north of Stop 8 were formed in open depressions between the ice and the valley wall and can be correlated with each of the three channels. This infers that the channels formed near the edge of the ice and not far under it.

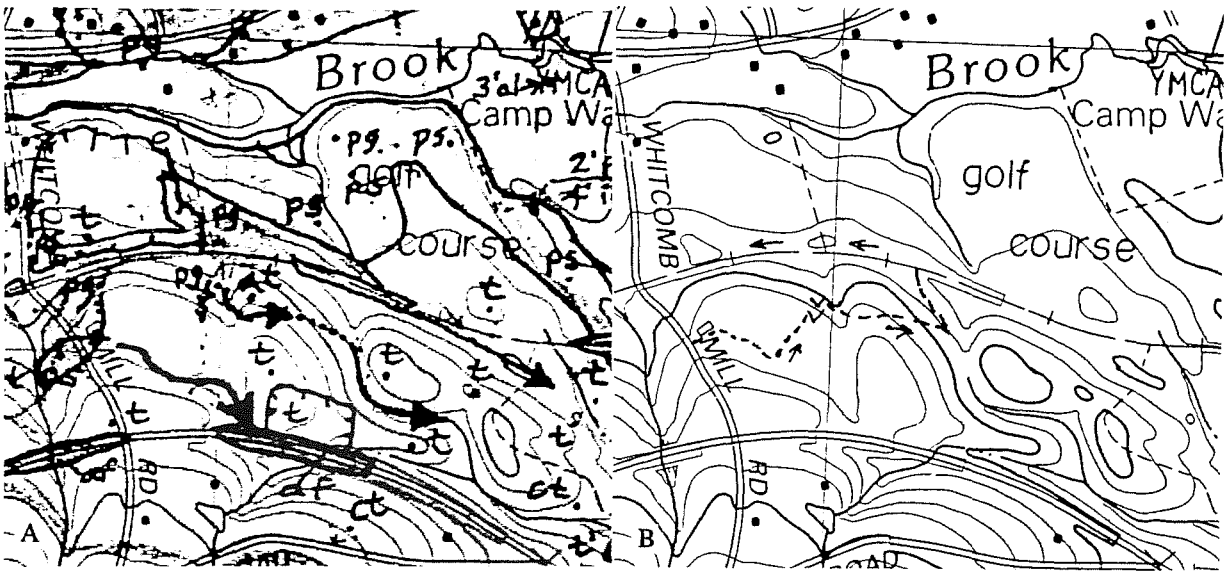


Figure 12. Portion of the Keene metric quadrangle showing location of Stop 8, ice-marginal channels: (A) geologic field map of surficial deposits and ice-marginal channels (arrows); t, till; at, ablation till; ct, compact till; pg, pebble gravel; ps, pebbly sand; af, artificial fill; (B) topographic map of same area showing approximate location of traverse to highest and middle channels. Scale 1 in = 315 m (1032 ft), contour interval = 6 m (20 ft), dark contour crossed by traverse in middle of map is 180 m ASL.

| | |
|---|---|
| | Make U-turn with caution. Proceed south on Whitcomb's Mill Road |
| 30.8 | Turn left (east) on Route 9. Road descends till-covered bedrock slope |
| 31.3 | Bedrock exposed on right |
| 31.7 | Continue straight on Route 9 at blinker and Base Line Road, leaving kame terrace gravel behind on right, enter lake-bottom deposits of Lake Ashuelot |
| 32.05 | Cross Ash Swamp Brook where Route 9 bends left to due east. For the next 1.9 mi the field trip route is on thin alluvium set into lake-bottom deposits. Most business establishments are on artificial fill. Note the high embankment of artificial fill on which Route 9 is located |
| 32.4 | Continue straight (east on Routes 10, 12 and 101) at first traffic light |
| 32.9 | Continue straight (east on Routes 12 and 101) at second traffic light |
| 33.4 | Cross Ashuelot River |
| 33.65 | Turn right (south on Route 12) at third traffic light |
| 33.9 | Cross bridge over The Branch |
| 33.95 | Start ascending front of Branch delta |
| 34.1 | Top surface of Branch delta rises gently to the east. Topset-foreset contact of the Branch delta was measured just south of green buildings at the right in St Joseph Cemetery |
| 35.0 | Road descends off edge of Branch delta at blinker, drops to stream and then rises across a series of kame terraces as described in the text and Figure 3 |
| 35.25 | "Dunham Shoes (?)" is on kame terrace ag-5 |
| 35.3 | "Nissan of Keene" is slightly higher and is on ag-4 |
| 35.5 | "Dry Firewood" is on wpg-3 |

35.7 Turn left (east) off Route 12 at sign "Great Meadows Homes" into old pit. Continue past slope with rip-rap on right
35.9 Old pit ends and road flattens on terrace cut in till. Test pits show 2 ft of pebble gravel over boulder till
36.2 Reverse direction at circle
36.6 Park on side of road in depleted gravel pit shown on Figure 13.

**STOP 9 HIGH KAME TERRACE (wpg-1), SITE OF COUNTY FAIRGROUNDS**

See Figure 3 for the vertical location of Stop 9 relative to the other kame terraces in this area. The south face of the pit has 15 ft of pebble-cobble gravel with boulders and is now covered by angular rip-rap. Formerly on the east side of the pit 7 to 8 ft of pebble-cobble-boulder gravel was observed resting on 4 ft of boulder till. The pit was excavated at the north end of a kame-terrace that extends to the southeast into the adjacent Marlborough metric quadrangle. The ice margin was oriented north-northwest to south-southeast on the west side of the kame terrace and forced high-energy meltwater streams to pass southeast through a low spot in the valley wall. The County Fairgrounds is located on a glaciofluvial surface that slopes southeast and is drained by a tributary of the South Branch of the Ashuelot River. A traverse due west from this site to Wilson Pond goes down the valley wall from (1) high kame terrace (wpg-1), to (2) middle kame terrace (wpg-3), to (3) low kame terrace (ag-4), and (4) across lake-bottom deposits of Lake Ashuelot to Wilson Pond (Fig. 13).

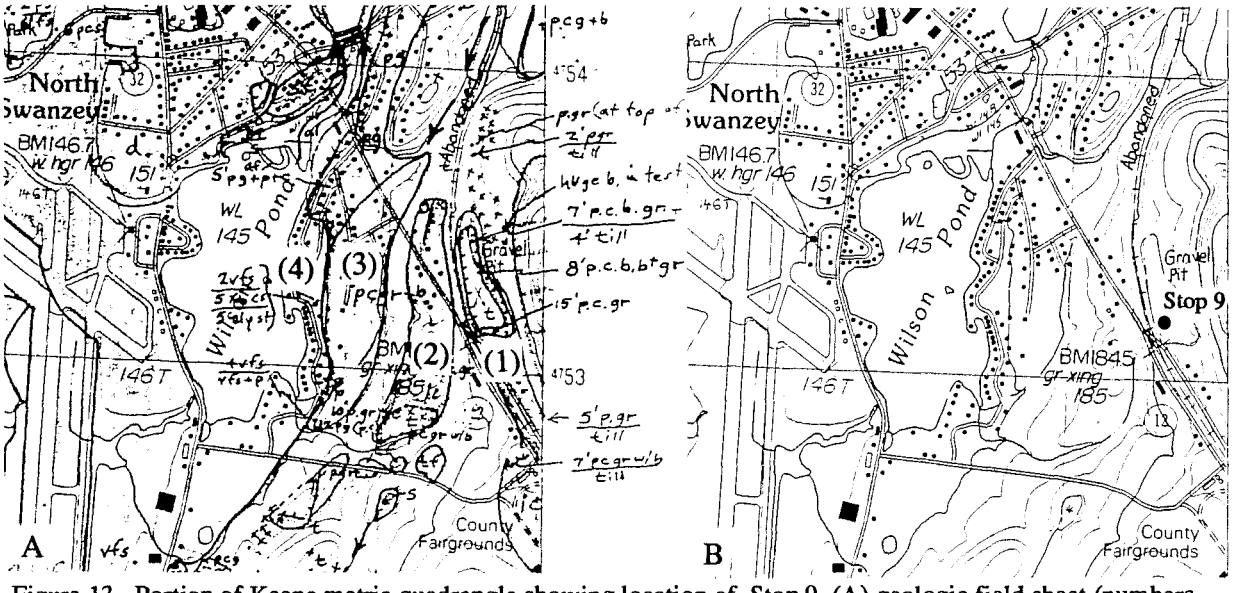


Figure 13. Portion of Keene metric quadrangle showing location of Stop 9, (A) geologic field sheet (numbers in parentheses represent sequence of formation as described above) and (B) topographic map of same area. Scale 1 in = 635 m (2082 ft), contour interval = 6 m (20 ft), dark contour above Wilson Pond is 150 m ASL.

36.7 Turn right on Route 12, retrace route north down over kame terraces, rise up on Branch delta
37.4 Turn left (southwest) on surface of Branch delta
37.8 Turn left (south) on Route 32
37.9 Road gradually drops off top of Branch delta. Airport runways at right are on artificial fill on lake-bottom/prodelta deposits. The level of Wilson Pond, a kettle, on the left is maintained by a dam at the south end
38.85 Road rises onto kame terrace, ag-2 (Fig. 3)
39.0 Road rises onto kame terrace, ag-1 (Fig. 3)
39.4 Road rises onto ice-contact delta, ad-5. Turn right into airport property. Proceed to Stop 10 at mile 39.75 (Fig. 14)

**STOP 10 WHITCOMB PIT IN ICE-CONTACT DELTA (ad-5)**

We've come full circle from ice-contact delta to ice-contact delta. In June, 1992, exposures in the Whitcomb pit were better than most others available because of recent excavation. Coarse-grained topset beds mostly have been removed from this deposit, leaving finer-grained foreset and bottomset beds. On the east face, collapsed south-dipping foreset beds are overlain by uncollapsed beds of fine sand that represent a small kettle filling. Some post-collapse beds dip toward the northwest the inferred direction of ice retreat. A former extension of the original delta to the northeast of Stop 10, as outlined by the 150 m contour, was excavated probably because it contained coarse-grained material. One could only guess that the former extension represents the locality where a meltwater stream in a subglacial tunnel debouched onto the surface of the delta. As shown in Figure 14 (A), this ice-contact delta, ad-5, was the second of four morphosequences formed in this area as the ice margin retreated from south to north.

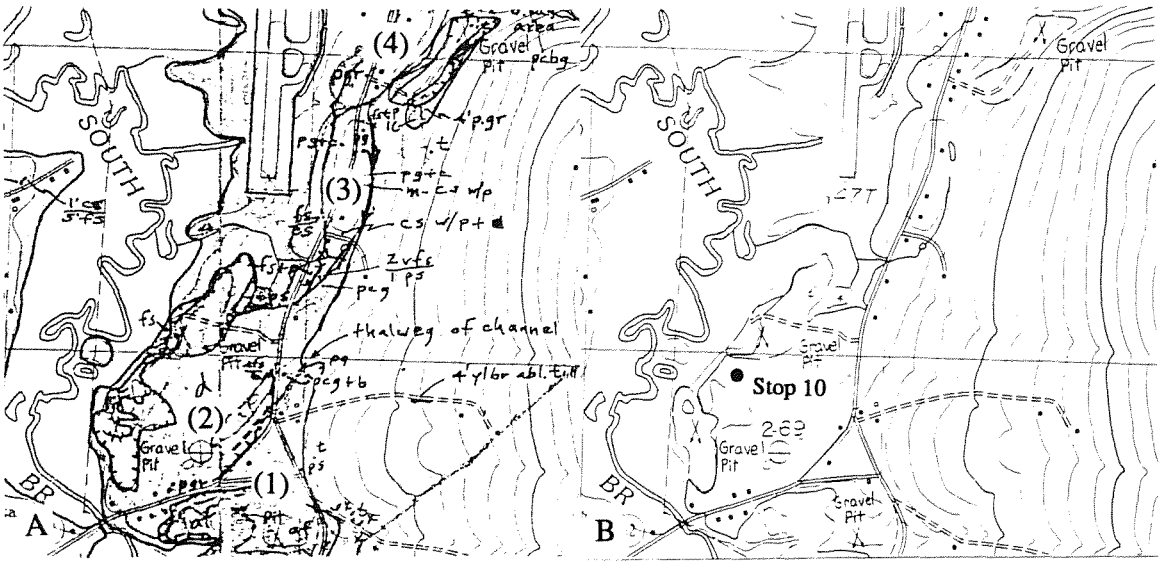


Figure 14. Portion of Keene metric quadrangle showing the location of Stop 10, (A) geologic field sheet, numbers in parentheses represent the order of formation of four morphosequences as the ice margin retreated from south to north, (1) ice-contact delta, ad-4; (2) ice-contact delta, ad-5; (3) kame terrace, ag-1; and (4) kame terrace, ag-2; (B) topographic map of same area, scale 1 in = 635 m (2082 ft), contour interval = 6 m (20 ft), dark contour just west of Stop 10 is 150 m ASL.

40.1 Route 32, turn left to Keene and points north, or turn right to Amherst and points south
40.2 Turning south, the road rises onto ice-contact delta, ad-4
40.4 Pit on left in ad-4
40. 7 Cross South Branch
40.95 Turn right off Route 32 at Mt. Caesar School. Note sign "2 mi to West Swanzey". Road rejoins road log at mile 21.4 in West Swanzey (Fig. 7).

## REFERENCES CITED

Antevs, Ernst, 1922, The recession of the last ice sheet in New England: American Geographical Society Research Series no. 11, 120 p.

Flint, R.F., 1957, Glacial and Pleistocene geology, New York, Wiley & Sons, 553 p.

Gustavson, T.C., Ashley, G.M, and Boothroyd, J.C., 1975, Depositional sequences in glaciolacustrine deltas, *in* Jopling, A.V., and McDonald, B.C., eds., Glaciofluvial and glaciolacustrine sedimentation: Society of Economic Paleontologists and Mineralogists Special Publication 23, p. 264-280.

Haley and Aldrich, Inc., 1976, Preliminary subsurface investigation for proposed arts center, Keene State College: Report to Shepley, Bulfinch, Richardson and Abbott, Boston, Massachusetts, 5 p. and boring logs

Jahns, R. H., 1953, Surficial geology of the Ayer quadrangle, Massachusetts: United States Geological Survey Geologic Quadrangle GQ-21.

Koteff, Carl, 1974, The morphologic sequence concept and deglaciation of southern New England, *in* Coates, D.R., ed., Glacial geomorphology: Publications in Geomorphology, Proceedings of Fifth Annual Geomorphology Symposium, Binghamton, New York, p. 121-144.

Koteff, Carl, and Larsen, F.D., 1989, Postglacial uplift in western New England: Geologic evidence for delayed rebound: *in* Gregersen, S., and Basham, P.W., eds., Earthquakes at North Atlantic passive margins: Neotectonics and postglacial rebound: Norwell, Massachusetts, Kluwer Academic Publishers, p. 105-123.

Koteff, Carl, and Pessl, Fred, Jr., 1981, Systematic ice retreat in New England: United States Geological Survey Professional Paper 1179, 20 p

Larsen, F.D., 1984, On the relative age of glacial Lake Hitchcock, glacial Lake Winooski, and the Champlain Sea: Geological Society of America Abstracts with Programs, v. 16, no. 1, p. 45.

Larsen, F.D., and Hartshorn, J.H., 1982, Deglaciation of the southern portion of the Connecticut Valley of Massachusetts, *in* Larson, G.J., and Stone, B.J., eds., Late Wisconsinan glaciation of New England, Kendall/Hunt, Dubuque, Iowa, p. 115-128.

Larsen, F. D., and Koteff, Carl, 1988, Deglaciation of the Connecticut River valley: Vernon, Vermont, to Westmoreland, New Hampshire, *in* Bothner, W. A., ed., Guidebook for field trips in southwestern New Hampshire, southeastern Vermont, and north-central Massachusetts (80th Annual Meeting of the New England Intercollegiate Geological Conference): Keene, New Hampshire, p. 103-125.

Lougee, R.J., 1939, Geology of the Connecticut watershed: New Hampshire Fish and Game Department, Biological Survey of the Connecticut Watershed, Survey Report 4, p. 131-149.

Lougee, R.J., 1957, Hanover in the ice age: Dartmouth Alumni Magazine, v. 50, p. 24-29.

Ridge, J.C., 1988, Quaternary geology of the upper Ashuelot, lower Cold River, and Warren Brook valleys of southwestern New Hampshire, *in* Bothner, W.A., ed., Guidebook for field trips in southwestern New Hampshire, southeastern Vermont, and north-central Massachusetts (80th Annual Meeting of the New England Intercollegiate Geological Conference): Keene, New Hampshire, p. 176-208.

Ridge, J.C., and Larsen, F.D., 1990, Re-evaluation of Antev's New England varve chronology and new radiocarbon dates of sediments from glacial Lake Hitchcock; Geological Society of America Bulletin, v. 102, p. 889-899.

Stone, J.R., Schafer, J.P., and London, E.H., 1982, The surficial geologic maps of Connecticut illustrated by a field trip in central Connecticut, *in* Joesten, R., and Quarrier, S.S., eds., Guidebook for field trips in Connecticut and south-central Massachusetts (74th Annual Meeting of New England of New England Intercollegiate Geological Conference): Connecticut Geological and Natural History Survey Guidebook 5, p. 5-25.

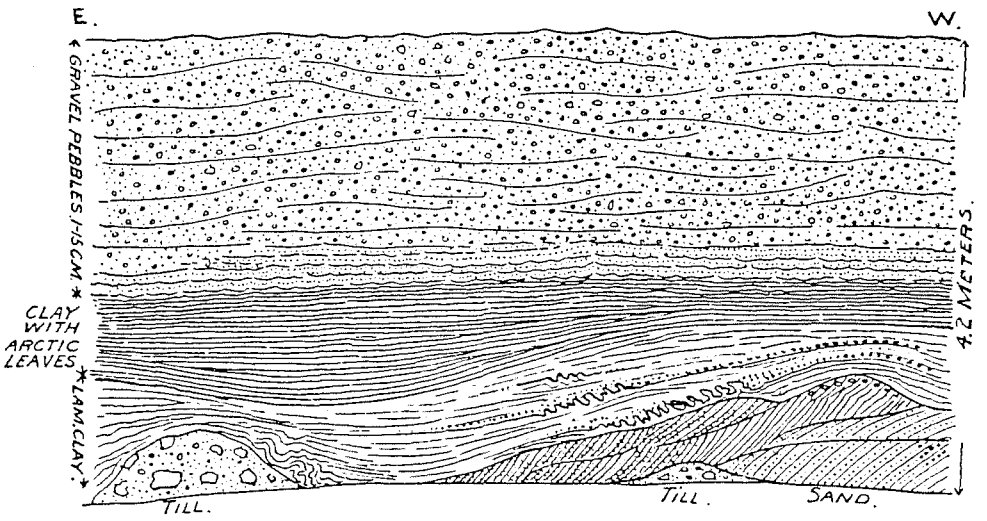


FIG. 38.—Enlarged section of the south side of cutting shown in fig. 37. The section represents the south side of the railroad cut beneath the bridge shown in fig. 37.

# THE MT. PROSPECT REGION OF WESTERN CONNECTICUT: MAFIC PLUTONISM IN IAPETUS-SEQUENCE STRATA AND THRUST EMPLACEMENT ONTO THE NORTH AMERICAN MARGIN

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## INTRODUCTION

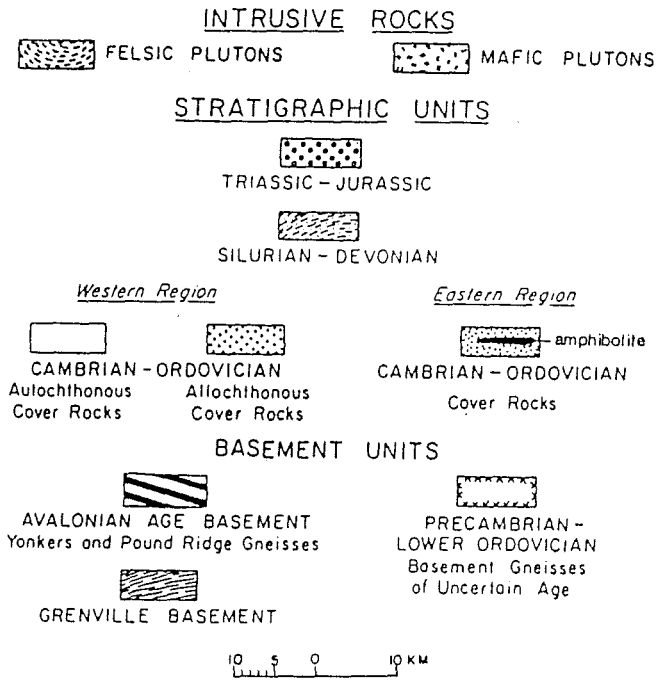
On this field trip, we will observe the structures and rock types along Cameron's Line in the vicinity of Litchfield, Connecticut. We will visit outcrops of Middle Ordovician Walloomsac Formation and Cambrian Manhattan Schist west of Cameron's Line, and the Cambrian/Ordovician Mt. Prospect Igneous Complex and surrounding Cambrian/Ordovician Hartland Group country rocks east of Cameron's Line. A number of outcrops seen on a previous NEIGC field trip (Panish and Hall, 1985) will be revisited, but additional outcrops of the Mt. Prospect Igneous Complex will also be seen. We will examine the field evidence for the inferred sequence of mafic igneous intrusion, deformation and metamorphism in a region affected by both the Taconian and Acadian orogenies. Figure 2 shows not only the field area mapped by the author (Panish, 1989), but an adjacent area to the west mapped by Dana (1978). Dana's area was added to show the stratigraphic section down to Precambrian basement and to better show the regional structural trends. Time constraints will prevent us today from visiting all the units shown in Figure 2.

## REGIONAL GEOLOGY

Cameron's Line is the map trace of a thrust fault that forms a major tectonic boundary in western Connecticut and southeastern New York (Fig. 1). Rocks west of Cameron's Line include all known exposures of Grenvillian basement, and also both autochthonous and allochthonous Cambrian and/or Ordovician cover rocks equivalent to the Middlebury synclinorium and Taconian sequences respectively. Rocks east of it are believed to have been deposited on oceanic crust and to have been transported westward along the Cameron's Line thrust fault onto North American continental crust during the Ordovician closure of the Iapetus ocean basin ( Hall, 1980; Hall and Robinson, 1982; Robinson and Hall, 1980). The Mt. Prospect Igneous Complex is a series of dominantly mafic plutonic rocks possibly derived from oceanic crust or mantle east of Cameron's Line and intrusive into the Iapetus-sequence rocks. Similar mafic intrusives are present east of Cameron's Line elsewhere in Connecticut and New York (Fig. 1). Some workers have favored early emplacement of the mafic intrusives prior to major motion on the Cameron's Line thrust fault (Robinson and Hall, 1980; Hall and Robinson, 1982), while others have argued that at least one of the mafic intrusives (i.e. the Hodges Mafic Complex) cross-cuts Cameron's Line and thus postdates the major thrust movement (Merguerian and Ratcliffe, 1977; Merguerian, 1983, 1985). Field evidence indicates that the Mt. Prospect Igneous Complex lies completely east of Cameron's Line and intruded early in the deformational sequence prior to major motion along the Cameron's Line thrust fault.

## STRATIGRAPHY

The stratigraphy of the Mt. Prospect region is most easily understood by considering the rocks in the North American plate separately from those of oceanic affinity (Fig. 3). All known occurrences of Grenvillian basement in western Connecticut are west of Cameron's Line (Fig. 1). These basement rocks are unconformably overlain by an autochthonous section of clastic and carbonate rocks, the Lowerre Quartzite and Inwood Marble, which were deposited along the margin of the North American plate during the Late Precambrian through Early Ordovician. Middle Ordovician sulfidic schists of Walloomsac were deposited unconformably upon these older autochthonous rocks. The Cambrian Manhattan Schist is a facies equivalent of the Lowerre Quartzite that was deposited on the continental slope east of the Lowerre. The Manhattan Schist was then tectonically transported westward, locally in three thrust sheets, onto the autochthonous section along thrust faults that root into the Cameron's Line thrust fault. The clastic and minor volcanic, Cambrian to Lower Ordovician Hartland Group rocks are considered, at least in part, to be a facies equivalent of the Manhattan Schist deposited east of the Manhattan Schist on oceanic basement. The Mt. Prospect Igneous Complex intruded into the Hartland section. The Complex is thus considered to have an oceanic crustal or possibly mantle derivation. The Hartland Group and the Mt. Prospect Igneous Complex were thrust westward as a package along the Cameron's Line thrust fault onto North American basement and its



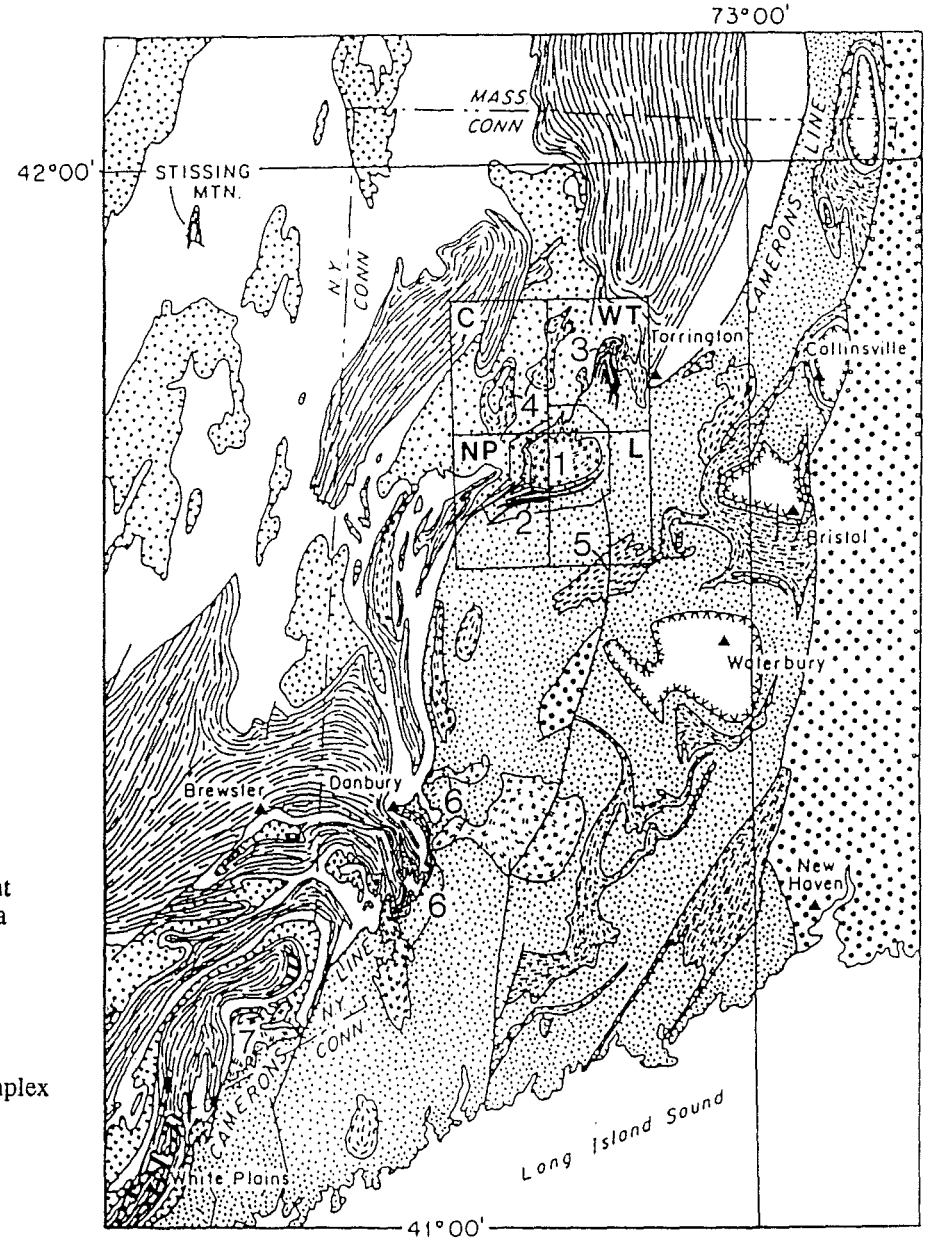
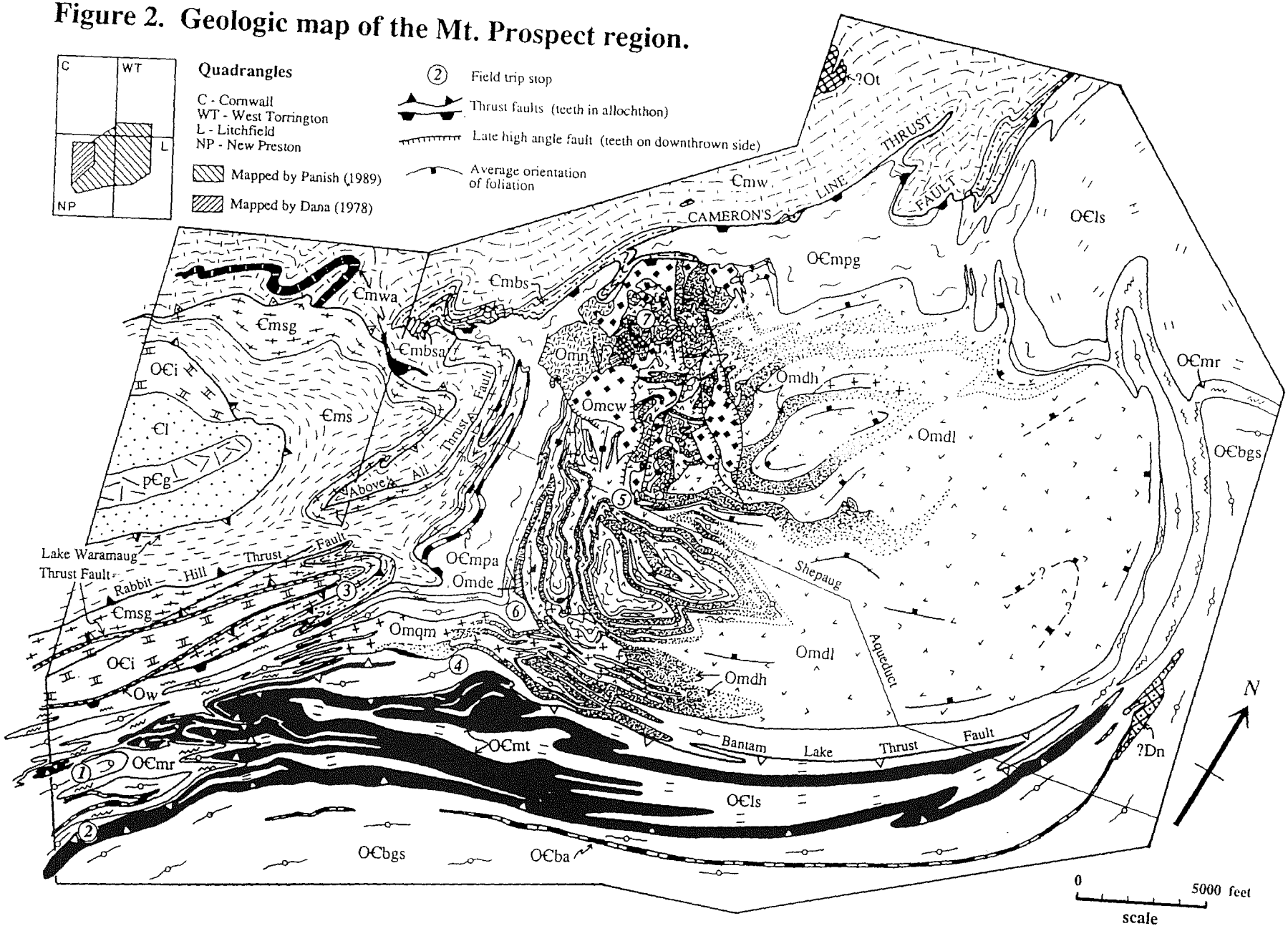




Figure 1. Generalized geologic map of western Connecticut and adjacent Massachusetts and New York (modified after Hall, 1980). The field area and adjoining 7.5 minute quadrangles are outlined.

Quadrangles: C - Cornwall; WT - West Torrington; L - Litchfield; NP - New Preston

1 - Mt. Prospect Igneous Complex
2 - Mt. Tom Amphibolite
3 - Hodges Mafic Complex
4 - Tyler Lake Granite
5 - Nonewaug Granite
6 - Brookfield Diorite
7 - Bedford Gneiss Complex

# Figure 2. Geologic map of the Mt. Prospect region.



Dev
Ordovician
?
?
Cambrian
Precambrian

?Dn Nonewaug Granite

?Ot Tyler Lake Granite

**CONTINENTAL PLATFORM**

Ow Walloomsac Formation

Middle Ordovician Unconformity

O€i Inwood Marble

€l Lowerre Quartzite

Unconformity

p€g Precambrian basement gneiss

**CONTINENTAL SLOPE**

Manhattan Schist

€msg Schistose granulite member

€ms Shepaug member

Biotite-garnet schist member

€mbs Schist unit

€mbsa Amphibolite unit

Warren member

€mw Schist unit

€mwa Amphibolite unit

**OCEAN BASIN**

Mt. Prospect Igneous Complex (see below)

Hartland Group

Bee Brook Formation

O€ba Amphibolite member

O€bgs Gneiss and schist member

O€mr Mt. Rat Schist

Mt. Prospect Formation

O€mpa Amphibolite member

O€mpg Gneiss member

O€mt Mt. Tom Amphibolite

O€ls Litchfield Schist

**Rocks of the Mt. Prospect Igneous Complex**

Omqm Quartz-bearing Monzonite

Omn Norite-Gabbro Association

Omcw Cortlandtite-Mafic Gabbro Series

Omp Pyroxene-Biotite-rich Gneiss

Diorite-Gabbro Association

Omdl Layered diorite-gabbro

Omdh Homogeneous diorite-gabbro

Omde Epidote-diopside gabbro

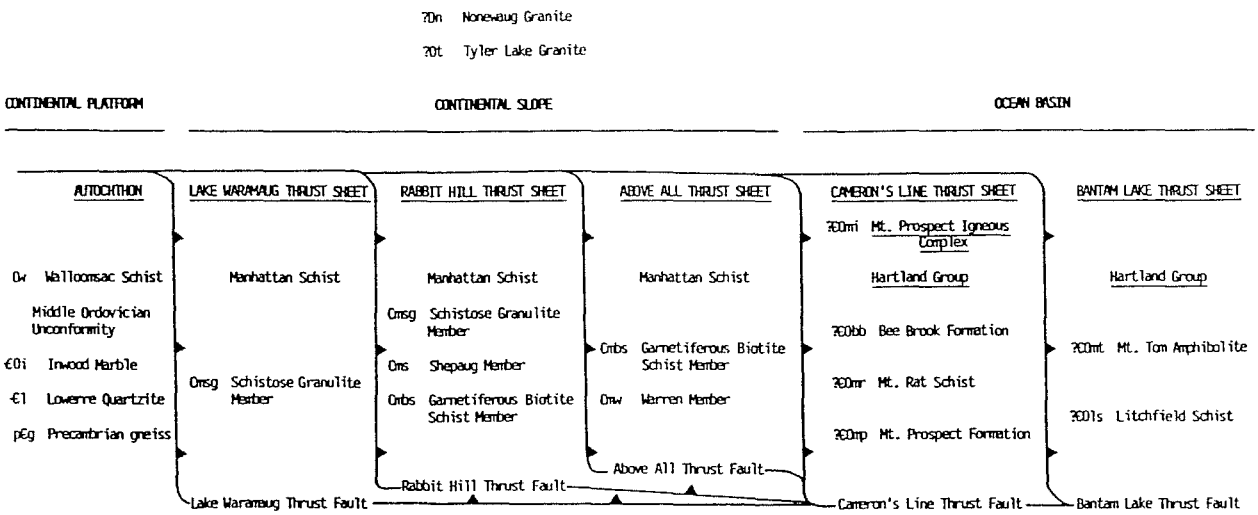



Figure 3. Stratigraphic summary chart showing the tectonic relationships among the rocks of the Mt. Prospect area.

autochthonous and allochthonous cover during the Taconian Orogeny. The Bantam Lake thrust fault is considered an early imbricate thrust fault within the Hartland section.

The bedrock of the Mt. Prospect area has been divided into 26 mappable units combined into six stratigraphic-tectonic groups consisting of Grenvillian basement and its autochthonous cover rocks, and the Lake Waramaug, Rabbit Hill, Above All, Cameron's Line and Bantam Lake thrust sheets (Fig. 2).

The rocks of the Grenvillian basement are dominantly gray, biotite-quartz-plagioclase gneisses. These gneisses are unconformably overlain by the Cambrian Lowerre Quartzite. The Lowerre Quartzite is made up of bedded, white to tan, locally slabby, quartzite, siliceous granulite and microcline-bearing schistose granulite. Overlying the Lowerre is the Cambrian-Ordovician Inwood Marble. The Inwood consists of well bedded, white, calcite-tremolite-dolomite marble, thin tremolite-rich beds, and white, calcite-cemented, dolomite marble. The Middle Ordovician Walloomsac Formation unconformably overlies the Inwood. The Walloomsac is a rusty-weathering, sillimanite-garnet-muscovite-plagioclase-biotite-quartz schist with minor, siliceous granulite beds. The Walloomsac is the only autochthonous unit that we will visit (Stop 3).

The Lake Waramaug, Rabbit Hill and Above All thrust sheets consist of four members of the Cambrian Manhattan Schist.

The Above All thrust sheet contains what are interpreted to be the stratigraphically lowest members, the Warren and Garnetiferous Biotite Schist members. The Warren member includes mainly interbedded, dark-gray, muscovite-garnet-chlorite-plagioclase-biotite-quartz schist, amphibolite lenses, siliceous granulite, and finely-layered schistose gneiss. A separate amphibolite unit consisting of lineated to foliated quartz-labradorite-hornblende gneiss was mapped within the Warren Member. The Garnetiferous Biotite Schist member is largely charcoal-gray, garnetiferous, sillimanite-muscovite-quartz-biotite schist with sillimanite nodules that protrude on the weathered outcrop. These schists are interbedded with minor thin beds of laminated gneisses. A separate amphibolite unit was mapped within the Garnetiferous Biotite Schist. This unit is a fine-grained, dark-green to black, plagioclase-biotite-hornblende gneiss with local thin layers of fine-grained, plagioclase-rich gneiss.

The Rabbit Hill thrust sheet contains in stratigraphically ascending order: the Garnetiferous Biotite Schist member, the Shepaug member, and the Schistose Granulite member. The Shepaug member consists of interbedded orthoclase-garnet-plagioclase-biotite-quartz schistose gneiss, rusty-weathering, schistose gneisses and schists with sillimanite rods, and subordinate granulite beds with sillimanite nodules similar to those in the Schistose Granulite member. The Schistose Granulite member is dominantly a fine-grained, well bedded, garnet-muscovite-biotite-

plagioclase-microcline-quartz schistose granulite or granulite with locally large, muscovite-sillimanite-quartz ellipsoids or nodules, and subordinate, rusty-weathering, biotite schist with sillimanite quartz rods.

The Lake Waramaug thrust sheet contains only the Schistose Granulite member.

The Hartland Group, which consists of Cambrian-Ordovician schists, gneisses, quartzites and amphibolites, has been divided into seven stratigraphic units within the Cameron's Line and Bantam Lake thrust sheets.

The Bantam Lake thrust sheet consists of the two stratigraphically lowest formations of the Hartland Group, the Litchfield Schist and the Mt. Tom Amphibolite. The Litchfield Schist is made up of silver-gray, garnet-biotite-oligoclase-quartz-muscovite schist with laminated quartzites, locally abundant staurolite and/or kyanite, and minor coticule layers. The Mt. Tom Amphibolite consists of dark-green to black, fine- to coarse-grained, massive, lineated, or foliated, quartz-andesine-hornblende gneiss.

The Cameron's Line thrust sheet contains, in stratigraphically ascending order: the Mt. Prospect Formation, the Mt. Rat Schist and the Bee Brook Formation. The Mt. Prospect Igneous Complex intrudes into these units. The Mt. Prospect Formation consists of gray to brown, commonly rusty-weathering, thinly layered or laminated, sillimanite-garnet-muscovite-plagioclase-quartz-biotite schistose gneiss with subordinate, commonly laminated, siliceous granulite, gray quartzite, biotite schist, local coticule, and garnet-cordierite-gedrite granulite. A mappable amphibolite unit within the Mt. Prospect Formation is a dark-green, slabby, well foliated, quartz-labradorite-hornblende gneiss. The Mt. Rat Schist consists of gray or silver-gray, rusty-weathering, laminated, garnet-plagioclase-muscovite-biotite-quartz schist that contains thin sandy quartz layers and irregular, red-staining quartz veins, black, rusty-weathering, highly fissile, biotite schist, and subordinate, siliceous granulites. The Bee Brook Formation is a thinly bedded unit of fine-grained, gray, massive to laminated, quartz granulites and quartzites, silver gray, muscovite schist, and subordinate, dark-gray, muscovite-biotite schist. Coticule layers are common. The Bee Brook Formation also contains a separately mappable amphibolite member.

## INTRUSIVE IGNEOUS ROCKS

The Mt. Prospect Igneous Complex is the oldest and by far the most areally extensive of the intrusive igneous rocks in the study area. Small bodies of granite and granitic pegmatite, of several ages and mostly unmapped, intrude the Mt. Prospect Igneous Complex and country rocks. Many are probably related to larger granite bodies outside the mapped area such as the Ordovician Tyler Lake and Devonian Nonewaug plutons. Also present are minor felsic veins associated with late jointing and minor faulting.

### MT. PROSPECT IGNEOUS COMPLEX:

The Mt. Prospect Complex is a highly deformed sheet of related igneous rocks covering 33 $km^2$ that intruded into rocks of the Mt. Prospect Formation and Hartland Group. In its northwest corner, the Complex is in thrust contact across Cameron's Line with the Manhattan Schist. The Complex was previously mapped by Cameron (1951) whose data were incorporated into quadrangle maps of Gates (1951) and Gates and Bradley (1952). Present mapping indicates that the intrusive rocks were emplaced in the following order:

1. **Diorite-Gabbro Association**: highly variable layered or homogeneous diorites to subordinate gabbros comprising the bulk of the Complex and consisting of three units:
Layered Diorite-Gabbro, Homogeneous Diorite-Gabbro and Epidote-Diopside Gabbro* (minor).

2. **Pyroxene-Biotite-rich Gneiss** (minor): a deeply-weathering, friable, biotite-rich gneiss only present on Mt. Prospect.

3. **Cortlandtite-Mafic Gabbro Series#**: mafic to ultramafic, commonly olivine-, augite-, and/or orthopyroxene-bearing rocks generally decreasing in age as follows:
a. Mafic gabbros and gabbronorites
b. Cortlandtite
c. Websterites and olivine websterites
d. Hornblendites.

4. **Norite-Gabbro Series#**: an undivided unit of felsic to mafic norites and subordinate hypersthene-bearing gabbros.

5. **Quartz-bearing Monzonite**: commonly porphyritic rocks with a matrix composition varying from quartz-bearing monzonite to diorite.

6. **Mafic Dike Rocks*** (minor): widespread dikes that include:
a. Diorite dikes and rare sills
b. Clinopyroxenite and hornblendite dikes.

* - minor units not shown in Fig. 2.
# - not subdivided in Fig. 2.

In addition to these main rock types, local assimilation and igneous brecciation have produced hybrid rock types. Textures and mineralogy have been further altered by metamorphism, the local introduction of ores, and deformation. The growth of metamorphic amphibole and biotite have been particularly extensive at many locations.

The Diorite-Gabbro Association, which makes up the bulk of the Mt. Prospect Igneous Complex, consists of highly variable layered and homogeneous diorites and subordinate gabbros. There is no consistent stratigraphy over the entire Complex and the various rock types repeat themselves at numerous places and in different thicknesses. The Layered Diorite-Gabbro consists of gray biotite-hornblende-andesine gneiss layers of different modal compositions and grain sizes, mafic augite-biotite-hornblende-labradorite layers, quartz-bearing monzonite layers, oligoclase microporphyry layers, and thin wispy biotite-hornblende layers. The Homogeneous Diorite-Gabbro consists of mafic, massive to foliated, biotite-augite-hornblende-plagioclase gneisses. This rock type can form continuous 15-100 m thick layers and thinner, discontinuous layers within the Layered Diorite-Gabbro, stock-like bodies, and margins around younger cortlandtite and norite intrusions. Some layers connect to the stock-like bodies. The Epidote-Diopside Gabbro is a dark gray, sphene-epidote-plagioclase-diopside-biotite-hornblende gneiss that forms 15 m thick linses and layers within the diorite-gabbros, and several large (<200 m across) inclusions within younger intrusions on Mt. Prospect.

The Pyroxene-Biotite-rich Gneiss is a minor unit of dark green, plagioclase-clinopyroxene-hornblende-biotite gneiss distinct from the older diorite units and only exposed as two irregular 350 m wide bodies near the summit of Mt. Prospect. The characteristic deep weathering and friability of the gneiss is largely due to the abundant biotite which increases toward the contacts with younger intrusives.

The Cortlandtite-Mafic Gabbro Series (undivided in Fig. 2) consists of spatially associated mafic gabbros and gabbro-norites, websterites, olivine websterites, and hornblendites that outcrop abundantly on Mt. Prospect and nearby hills, but are absent outside the northwest quarter of the Complex. Rocks of this series are typically medium- to coarse-grained, dark green or black, dark-brown weathering, massive, and homogeneous. Interstitial plagioclase, dark green augite, locally rusty hypersthene, and tan olivine typically form 2-3 mm wide grains. Plagioclase commonly comprises <25% of the gabbros. Poikilitic grains of hornblende and biotite are commonly several centimeters across. The cortlandtite and hornblendite are the result of varying detrees of hornblendization of the other rock types and are generally not separate intrusions. Hornblendization tends to be most advanced near contacts. Subdivisions of the series commonly proved impractical to map separately because the rock types can grade texturally and mineralogically into one another. Elsewhere contacts between the rock types are sharp and irregular or brecciated, but they are obscured by the mafic nature of the rocks. It is generally true within and between both the Cortlandtite-Mafic Gabbro and Norite-Gabbro series that the more felsic rocks intrude the more mafic rocks. Disseminated sulfides can be abundant near contacts. Attempts were made to mine the sulfides for nickel and for use as a flux beginning about 1835 (Cameron, 1943), but all that remains are eight abandoned pits and shafts. Ore rocks are quite rusty and friable.

The Norite-Gabbro Series consists of a variety of felsic to mafic norites and subordinate gabbros that, like the Cortlandtite-Mafic Gabbro Series, outcrop abundantly on Mt. Prospect and nearby hills. The various rock types are massive or, less commonly, foliated, All types have augite, biotite, amphibole, hypersthene and variably abundant plagioclase ($An_{35-65}$) as essential minerals. Quartz and rare orthoclase can be present in trace amounts; olivine is absent. Hornblendization is not as extensive as it is in the Cortlandtite-Mafic Gabbro Series. The rock types of the Norite-Gabbro Series can intrude and brecciate one another, in which cases the more felsic varieties are generally younger than the more mafic ones. In other places, variants grade imperceptibly into one another and consequently

this unit was not subdivided. The felsic varieties also locally grade into Quartz-bearing Monzonite. The Norite-Gabbro Series overall forms discordant bodies that are locally bordered by intrusive breccias; however, other contacts are likely faults. Outcrop relationships imply that this unit is generally younger than the Cortlandtite-Mafic Gabbro Series, but this sequence cannot be demonstrated for every body of norite and cortlandtite because there are many inferred contacts with uncertain relationships despite the abundant outcrops.

The Quartz-bearing Monzonite is typically an unlayered, medium- to coarse-grained, bioitite-K-feldspar-quartz-plagioclase ($An_{25-54}$) porphyritic gneiss with variable amounts of 2-5 cm long, euhedral phenocrysts of microperthitic microcline and orthoclase. Smaller plagioclase phenocrysts may also be present. This unit intrudes the Hartland Group country rocks and all the major igneous rocks of the Mt. Prospect Igneous Complex, especially the Layered Diorite-Gabbro. The major bodies of Quartz-bearing Monzonite include markedly discordant, irregular intrusions into the mafic igneous rocks on Mt. Prospect, grossly concordant lenses and sills within the Layered Diorite-Gabbro and along the major diorite/country rock contacts, and the large, discordant tabular body making up the southwest 'tail' of the Mt. Prospect Igneous Complex. In addition, the Layered Diorite-Gabbro is intimately intruded by countless minor sills, dikes and irregular bodies of quartz-bearing monzonite that commonly make up 10-20% of individual outcrops. To a much leser extent the quartz-bearing monzonite invades the Cortlandtite-Mafic Gabbro and Norite-Gabbro series, but it does, and that establishes the intrusive sequence.

There are numerous, simple and composite, diorite dikes and rare sills that intrude all the major units of the Mt. Prospect Igneous Complex and the Hartland country rocks. The dikes and sills are fine-grained and gray, have andesine, hornblende and biotite as the major constituents, and are typically a few centimeters to a few meters thick. Three dikes, however, are up to 30 m thick and 300 m long. The dikes are particularly abundant at norite contacts. Also present on Mt. Prospect are minor, massive clinopyroxenite and hornblendite dikes and veins that intrude rocks of the Cortlandtite-Mafic Gabbro and Norite-Gabbro series, Quartz-bearing Monzonite, and late diorite dikes. The dark green to black clinopyroxenite consists of interlocking anhedral clinopyroxene with minor interstitial hornblende, biotite and plagioclase. The black hornblendite, which consists of anhedral poikilitic hornblende, relict clinopyroxene, and minor interstitial plagioclase and biotite, apparently formed through late hydration of the clinopyroxenite.

## STRUCTURAL GEOLOGY

Truncation of the Precambrian gneisses at the unconformity beneath the Lowerre Quartzite demonstrates that Precambrian deformation occurred (Dana, 1978), however, this is not the subject of this field trip. We will instead focus on the deformation in the Paleozoic section.

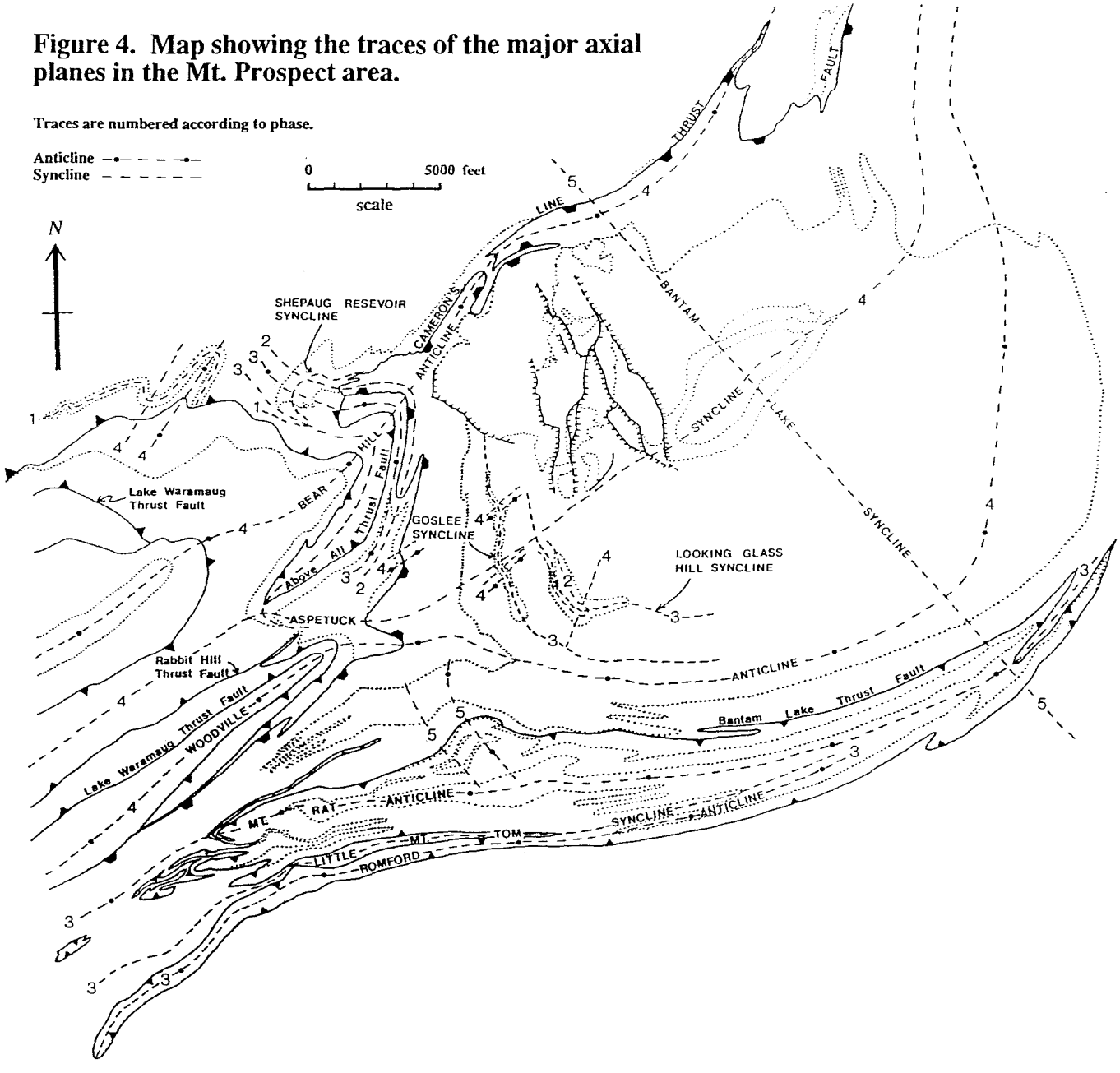
The Paleozoic rocks of the Mt. Prospect area has been subjected to three Taconian (D1-D3) and two Acadian (D4-D5) phases of deformation, and at least one late phase of brittle deformation (D6) that is probably Mesozoic (Table 1, Fig. 4). D1 includes: 1) an imbricate thrust fault (Bantam Lake thrust fault) within the Hartland section that is inferred to exist from the apparent loss of the Litchfield Schist section adjacent the Mt. Tom Amphibolite; 2) the Morris synform which causes the two terminations of the Mt. Tom Amphibolite in the southeast corner of the mapped area and which folds the Bantam Lake thrust fault; and 3) early folds in the Manhattan Schist that can be truncated by D2 thrusts (e.g. the Shepaug Reservoir syncline). D2 produced: 1) the dominant regional foliation; 2) the three thrust faults at the base of and within the Manhattan Schist (Lake Waramaug, Rabbit Hill and Above All thrust faults); 3) the Cameron's Line thrust fault; and 4) map scale isoclinal folds (e.g. Mt. Rat anticline and Little Mt. Tom syncline). Interference of F1 and F2 folds produced the complicated boomerang pattern in the Mt. Tom Amphibolite. All major igneous rocks of the Mt. Prospect Igneous Complex intruded prior to D2 deformation including the regional thrust faulting. The Cortlandtite-Mafic Gabbro series and younger rocks post-date and cut D1 features. D3 isoclinally to tightly folded D2 folds, faults and the dominant foliation. The infolds of country rock in the southwestern quarter of the Complex are the result of the interference of F2 and F3 (e.g. Goslee and Looking Glass Hill synclines) folds. D4 produced open to tight regional folds having generally NW-dipping axial surfaces and major anticlinal limbs steeply overturned to the southeast. These folds (e.g. Bear Hill and Woodville anticlines, Aspetuck syncline) dominate the map pattern and regionally overturn the stratigraphy. These major D4 folds diverge around the Mt. Prospect Igneous Complex which appears to have acted like a buttress. The Aspetuck syncline broadens out across the Complex and minor F4 folds help to accomodate the strain. D5 produced: 1) a regional NW-trending open fold (Bantam Lake syncline), 2) smaller map-scale folds in the Mt. Tom Amphibolite, and 3) conjugate, NE-trending minor folds seen in outcrop. The Bantam Lake syncline causes the F4 axial planes to shallow in dip and veer to the north along the east side of the mapped area. The plunge of the major F4 folds also varies from N20W-N30E in the west to N3-75W in the east. The plunge variation explains the local basin in the

Table 1: Summary of deformation and metamorphism in the Mt. Prospect area.

| Phase of Deformation | Structural Description: Major Features | Structural Description: Minor Features | Igneous Activity and Metamorphism |
|---|---|---|---|
| D5 (Acadian) | Conjugate set of NW- and NE-trending folds | | |
| | Map scale N10W to N60 W-trending, open folds. | Minor chevron folds and kink bands, and axial plane crenulation cleavage. A quartz-filled axial planar fracture cleavage is faint but widespread. | |
| | Possibly one SE-dipping open fold at Looking Glass Hill. | Minor NE- to E-trending, open folds are locally present. A fine, quartz-filled cleavage that parallels the axial planes is only locally present. | Late retrograde chlorite |
| D4 (Acadian) | Major open to tight, NE-trending, NW- to NE-plunging folds which dominate the map pattern. late muscovite is folded. | Axial planar crenulation cleavage; alignment of sillimanite, staurolite, and quartz rods parallel to fold axes. Some kyanite is aligned also. Some Late sillimanite grows. | Late muscovite<br>M2 regional peak<br>Kya pseudomorphs andalusite and overgrows sil. |
| | | | Andalusite forms? |
| D3 (Taconian) | Typically N-plunging, isoclinal to tight folds that are responsible for the refolding of F2 folds in the Mt. Prospect Igneous Complex/country rock contacts. | Minor isoclinal folds; local crenulation cleavage; alignment of sillimanite; quartz rods parallel fold axes. | |
| D2 (Taconian) | Late D2 isoclinal folds including folds in the Mt. Prospect Igneous Complex/country rock contacts.<br><br>Cameron's Line Thrust Fault post-dates other thrust faults.<br><br>Early D2 westward thrusting of the Manhattan Schist and imbricate faulting within the Manhattan Schist. | Dominant regional foliation. It consists of aligned muscovite and biotite plates. Minor isoclinal folds of compositional layering. Sillimanite, hornblende, and quartz rods parallel to fold axes. | Tyler Lake Granite<br>M1 regional peak<br>to sillimanite-Kspar grade. |
| | | | Quartz Monzonite intruded |
| D1 (Taconian) | Three map-scale folds including the Morris Synform in the Mt. Tom Amphibolite.<br><br>Bantam Lake Thrust Fault in Hartland. | Rarely, minor isoclinal folds in the compositional layering can be demonstrated to be F1 and not F2. A faint D1 foliation consisting of the planar alignment of fine-grained biotite, hornblende, and feldspar is present in the diorite, but is rare in the country rocks. Hornblende lineation in the Mt. Tom Amphibolite. | Intrusion of the mafic gabbros, norites, and peridotites with local development of a contact aureole with sil-kya-ksp-crd-ged assemblages. Early diorites of the Mt. Prospect Igneous Complex intruded with the local development of a contact aureole with sil-ksp-crd assemblages. |

**Figure 4. Map showing the traces of the major axial planes in the Mt. Prospect area.**

Traces are numbered according to phase.



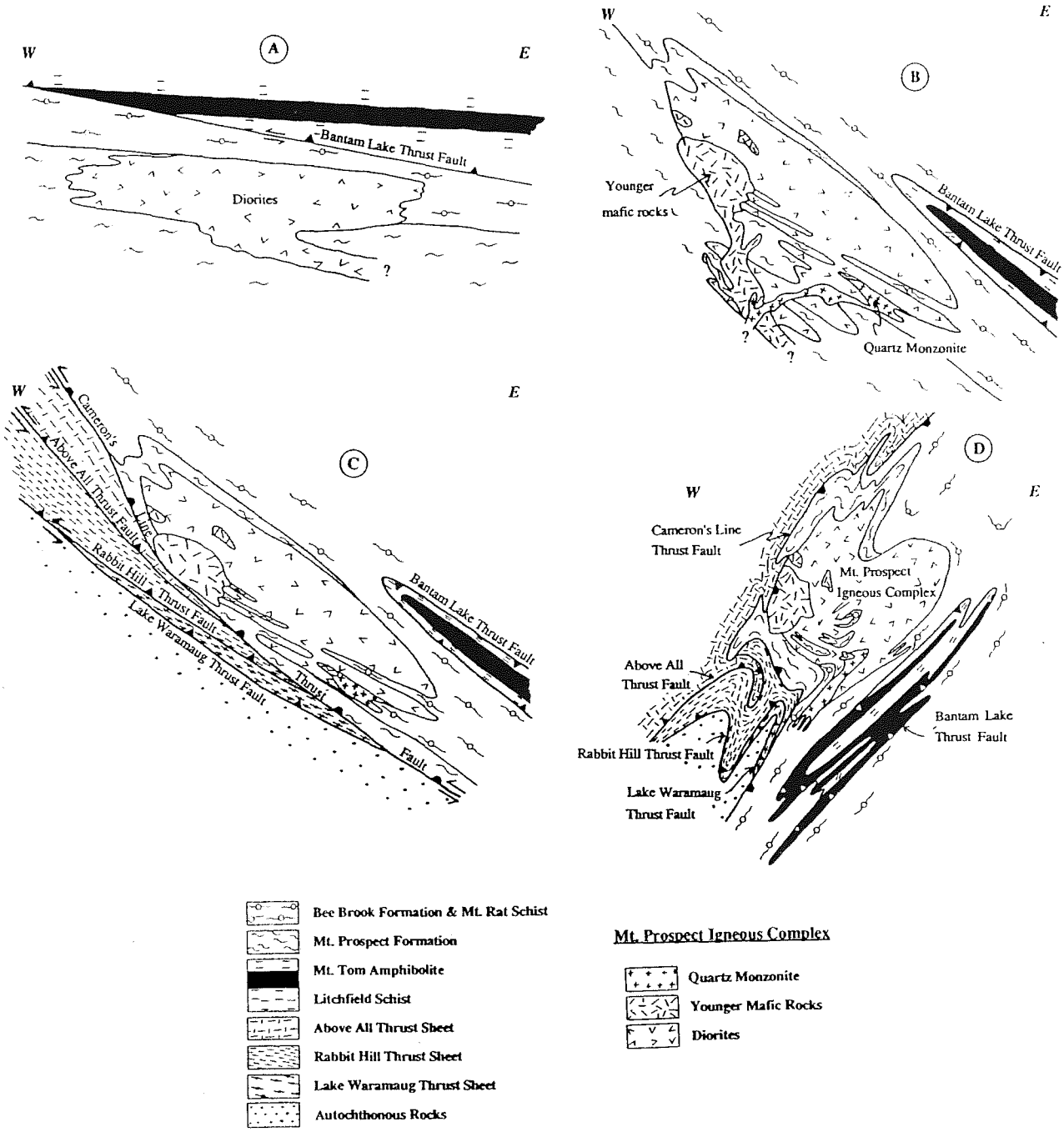



Figure 5. Diagrammatic cross sections illustrating the structural development of the Mt. Prospect area. A. Early D1: Emplacement of the Bantam Lake thrust sheet after the diorites of the Mt. Prospect Igneous Complex intruded along the Mt. Prospect Formation-Mt. Rat Schist-Bee Brook Formation contacts (the Mt. Rat Schist is not shown separately). B. Post-D1: The younger mafic rocks (not subdivided) and the quartz-bearing monzonite intrude after D1 isoclinal folding of the diorites and country rocks. C. Early D2: Major westward motion on the Cameron's Line thrust fault has emplaced the Hartland section and the Mt. Prospect Igneous Complex onto the Manhattan Schist. Thrust faults within the Manhattan section root into Cameron's Line. D2 folds have not yet deformed the thrust faults. D. D4: Major D4 folds develop in the Mt. Prospect Igneous Complex and country rock. The stratigraphic sequence is overturned.

center of the Complex. It should be stressed, however, that there are few outcrops in the eastern half of the Complex, and consequently the eastern map pattern is rather diagrammatic. Agar's (1927) subsurface measurements in the Shepaug aqueduct tunnel confirm a shallowing of the dips of layering and foliation in the keel of the Bantam Lake syncline. Minor folds, mylonitic foliation, crenulations, and mineral and intersection lineations are also associated with one or more of the D1-D5 deformations (Table 1).

There are also Paleozoic brittle-ductile structures with a wide range of ages within the Complex. Included are mafic dikes and veins that are associated with the major intrusives and that are therefore prior to most ductile deformation. Some high angle faults, mylonites, fault breccias, granite pegmatites and aplite dikes post-date Taconian structures, but are deformed by D4 and D5 structures. The high angle faults partly bound several igneous bodies of the Cortlandtite-Mafic Gabbro and Norite-Gabbro series and explain why these bodies cross-cut younger D2-D4 structures. Mesozoic minor faults, joints and felsic veins are typically planar structures that cross-cut all other features and that form well-oriented sets from which can be inferred principal stress directions in agreement with regional Mesozoic trends (Wise, 1981).

Critical to all regional tectonic scenarios is the age of intrusion of the Mt. Prospect Igneous Complex. Regional and local evidence imply that all major igneous units of the Complex had intruded the Mt. Prospect Formation and Hartland Formations before major movement on Cameron's Line. The igneous rocks of the Complex lie completely east of Cameron's Line, do not cut Cameron's Line, and contain inclusions of Hartland Group rocks but not Manhattan Schist. Regionally, mafic plutons (Fig. 1) similar to the Mt. Prospect Complex can also be interpreted to lie completely east of Cameron's Line. No one has interpreted any of these mafic plutons as lying completely west of Cameron's Line or as being bounded by Cameron's Line on their eastern margins. However, Merguerian and Ratcliffe (1977) and Merguerian (1983) interpret one of the mafic plutons northeast of Mt. Prospect, the Hodges Mafic Complex, as crosscutting Cameron's Line.

The D2-D3 folds in the diorite-country rock contact and the inferred first phase foliation in the diorite demonstrate that the diorites, which are the oldest major intrusives, intruded early in the deformation sequence. The quartz monzonite, which is the youngest major intrusive, locally displays the dominant D2 foliation and minor D2 folds (Stop 3), and is deformed by major D3 folds. The quartz monzonite thus predates D2 and D3; its age relative to D1 is uncertain.

The generally massive appearance of the norites and related rocks as well as their stock-like, map appearance suggest that they are the youngest intrusives. They cross cut the igneous layering of the diorites and D2-D3 infolds of the country rocks within the Complex (e.g. Goslee syncline). However, it is clear that many of the cross cutting norite boundaries are late high-angle faults. Furthermore, the quartz monzonite is seen in detail to intrude all the younger mafics. The younger mafic rock types are thus considered to be pre-D2.

The Mt. Prospect Igneous Complex is interpreted to be a thick, complexly folded sheet that owes its general shape to the interference of major D4 and D5 folds. The regional D4 Aspetuck syncline and Woodville anticline folded the sheet into a N to NW plunging cylinder with a "stout S-shaped" cross section. The E-W vertical cross section of Figure 5D is an oblique section through that cylinder. The Complex now lies in the keel of the open D5 Bantam Lake Syncline which changed the D4 axial plane trends, shallowed the dips of the D4 axial planes in the center of the Complex, and produced a shallow basin in the center of the Complex (Figs. 2 and 4).

## METAMORPHISM

The Mt. Prospect area has a complex metamorphic history involving local contact metamorphism adjacent the Mt. Prospect Igneous Complex followed by Taconian and Acadian regional metamorphism. The local contact aureole, which is confined to rocks of the Cameron's Line thrust sheet, typically has sillimanite-garnet-orthoclase-biotite-cordierite and kyanite and/or sillimanite-garnet-biotite-cordierite-gedrite contact assemblages which were altered by the later regional metamorphic events. The regional metamorphism can be described in three parts. A Taconian prograde phase, that was likely contemporaneous with D2-D3 deformation, reached sillimanite and sillimanite-K-feldspar grades and dominates the western part of the field area. This phase was transitionally overprinted to the east (the stippled transition zone of Fig. 6) by a prograde Acadian phase characterized by the growth of kyanite and some late sillimanite. This second phase produced the eastern sillimanite, sillimanite-staurolite, kyanite-sillimanite and kyanite-staurolite zones. These zones were increasingly modified eastward and southward by the third phase. The third phase is an Acadian retrograde phase that caused late staurolite and garnet

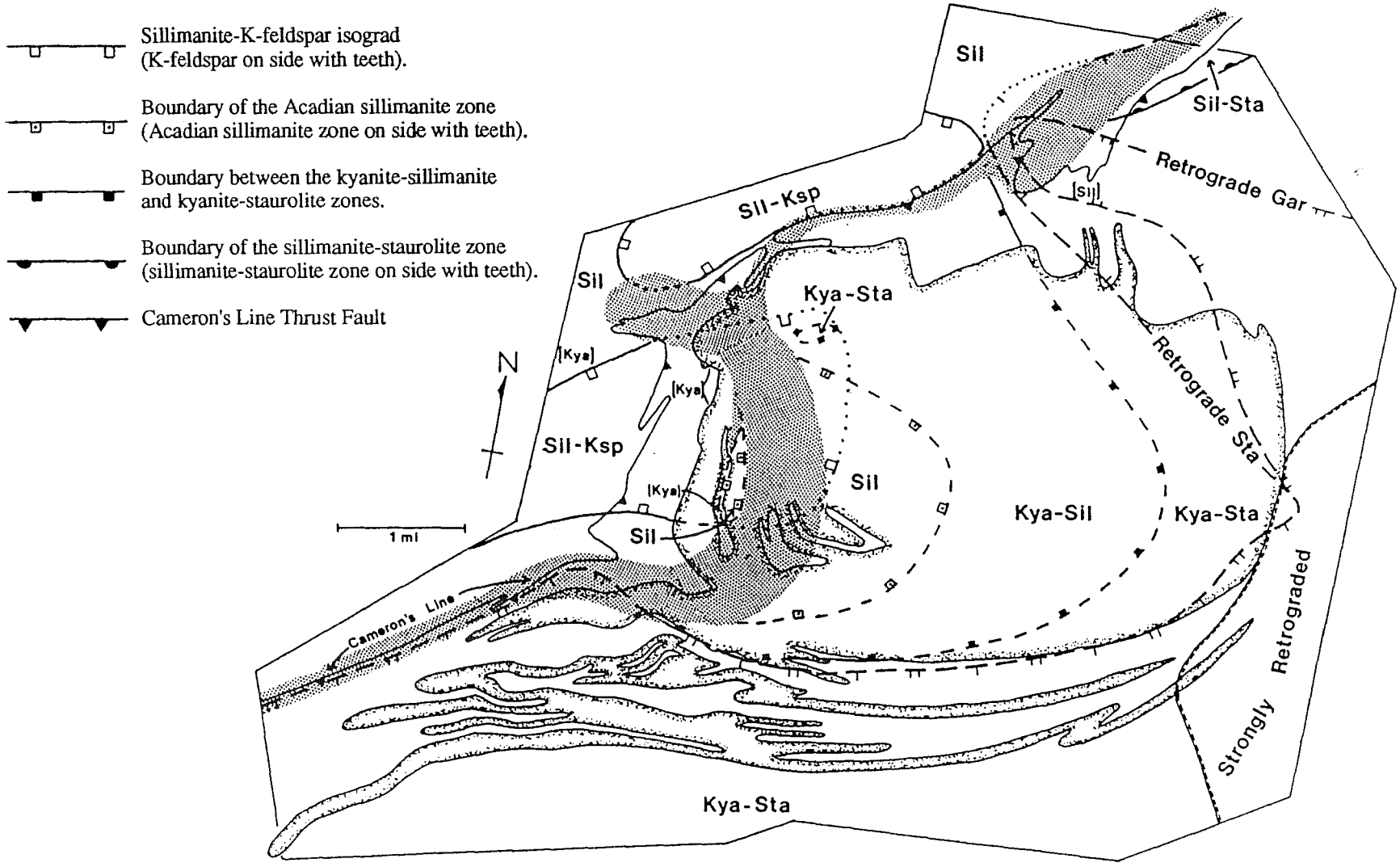



Figure 6. Metamorphic map of the Mt. Prospect area. Stippled pattern represents the transition zone between areas dominated by Taconian metamorphism to the west and Acadian metamorphism to the east. Mineral names in parentheses indicate isolated occurrences of those minerals. The retrograde staurolite and garnet lines are not true isograds, but just show the geographic limits of the minerals. Teeth mark the sides of the retrograde staurolite and garnet lines on which retrograde staurolite and garnet respectively appear.

growth and that locally retrograded rocks down to biotite grade. The grade and age of the dominant regional metamorphism generally decreases eastward and southward across the mapped area.

Most biotite and muscovite flakes are oriented in the dominant foliation although coarse retrograde muscovite, commonly containing remnant sillimanite, can cross-cut the foliation. Sillimanite and sillimanite-quartz-orthoclase rods are most common west of the transition zone, are oriented in the dominant foliation, and parallel not only F4 axes, but F3 and possibly F2 axes. These rods are considered primarily Taconian D2 and D3 features that were reoriented during the Acadian D4 deformation. Kyanite and staurolite, which are found in and east of the transition zone, parallel the dominant foliation. Staurolite locally forms lineations, but kyanite rarely does. Both minerals are considered Acadian. Kyanite commonly overgrows sillimanite, fibrolite and minor F2 folds. Local, 1-2 cm long "bricks" of coarse kyanite appear to be pseudomorphs after andalusite although andalusite itself has not been identified in the field area. There are several generations of garnet. Garnets east of the transition zone are, in particular, texturally and chemically complicated. Relict sillimanite can be present in the garnet, but absent in the kyanite-bearing matrix. These sillimanite inclusions are commonly oriented at high angles to the external foliation.

Regional metamorphism has also affected the rocks of the Mt. Prospect Igneous Complex. Large poikilitic hornblende and biotite grains have replaced pyroxene and olivine. Orthopyroxene-plagioclase-spinel-hornblende coronas have developed on other olivine grains. Metamorphic pigeonite lamellae are present in augite.

Pressure and temperature estimates based on microprobe data of selected samples are quite scattered as might be expected from a multiply metamorphosed terrane. Best estimates of Acadian peak conditions are based on those pelitic samples which clearly have Acadian kyanite with or without Acadian sillimanite. Garnet-biotite temperatures (Williams and Grambling, 1990, eq, 13 corrected) and GASP pressures (Koziol and Newton, 1988) using garnet rim compositions from these samples ranged from 500 to 600°C and 5.5 to 7.7 kb respectively.

## ACKNOWLEDGMENTS

I wish to thank the Connecticut State Geological and Natural History Survey and the Geological Society of America for financial support for dissertation field work back in the early 1980's. I also wish to thank all the local landowners for their cooperation and permission to enter their properties. Special thanks must be given to Prof. Peter Robinson of the UMass Department of Geology and Geography for being a patient principal advisor during the protracted writing of my Ph.D dissertation and for his continued valuable advice. I would also like to express my deep appreciation of the late Dr. Leo Hall of UMass who was my original principal advisor and who co-authored the 1985 NEIGC field trip to the Mt. Prospect area.

## ROAD LOG

**Meeting time and place: 10:00 AM,** Sunday October 11, 1992, at Stop 1 near New Preston, CT. Please note that the original registration notice mentioned a 9:30 AM meeting time. However the meeting time has been delayed 1/2 hour to 10:00 AM to allow a full 2 hours for driving from the UMass campus. For those wishing to return to Amherst after the trip, meet at 7:45 AM sharp at the parking lot north of the UMass stadium for consolidation purposes. Rest rooms and fresh water will be available at the lunch stop. **Please bring your lunch.** Two travel routes from the UMass campus are suggested:

Simplest route (repeated from the mailed registration information): From Amherst take Rt. 9 west for 6 miles and then take Rt. 91 south for 10.5 miles. Take Exit 16 onto Rt. 202 south at the Old Soldiers Home. Continue west and south about 36 miles on Rt. 202 through Westfield, MA to Avon, CT. From Avon continue west on Rt. 202 through Litchfield, CT to intersection with Rt. 47. Stop 1 is 0.5 miles south on Rt. 47 (not 0.1 miles as mentioned in the registration information).

Scenic route (which in heavy traffic is also the quicker route): Follow the directions of the first route until you reach downtown Westfield where at the common the Rt. 202 signs direct you to take first a right onto Court Street and then, almost immediately, a left off of Court Street. Do not take the left, but continue straight on Court Street past Nobel Hospital for about 1.1 miles. Take a left onto Granville Street immediately after Highland Elementary School. In another 0.9 miles, the road will curve to the left over a small bridge. Take a right immediately after the bridge which will keep you on Granville Road. In 0.7 miles you will reach a fork in the road. Bear right, continuing on Granville Road. After another 6 miles take a right onto Rt. 57 west. Continue west on Rt. 57 for 13 miles to New Boston. Turn left onto Rt. 8 south and continue 13 miles through Tolland, CT center where Rt. 8 becomes a divided highway. Continue 8 miles and take Exit 44 to Rt. 202. Continue on Rt. 202 west and south through downtown Torrington, Litchfield and Bantam until you reach the intersection with Rt. 47. Stop 1 is 0.5 miles south on Rt. 47.

**STOP 1:** Here are several roadcuts extending for 550 feet along the road that consist of two members of the Bee Brook Formation and the Mt. Rat Schist (Fig. 7). Note that I have locally subdivided the Bee Brook Formation into three members. Abundant minor folds are present in these outcrops. A discussion of the structural features will take place at the outcrop. Starting at the north end and walking south, the following rock units are encountered:

The Muscovite Schist Member of the Bee Brook Formation is exposed in the first 10 to 15 feet of outcrop. This unit is a fine- to medium-grained, silver gray, garnet-staurolite-chlorite-biotite-quartz-muscovite schist. Subordinately present are wispy, centimeter scale layers of a darker garnet-chlorite-muscovite-biotite schist and fine-grained, gray, siliceous granulite layers. Coarse-grained, massive, quartz lenses, pods, and bulbous fold noses <10 cm wide are abundant. Chlorite-biotite schists with local silky, greenish-gray foliation surfaces are also present. The schistosity is commonly crenulated.

The Gneiss and Granulite Member of the Bee Brook Formation is exposed over the next 350 feet of outcrop. This member consists of thinly-interbedded gneiss, quartzite, granulite and schist. Bedding is 1 cm to 2 m thick. Gray to slightly greenish-gray, thinly-bedded, fine-grained, garnet-biotite-chlorite-muscovite quartzites are interbedded with subordinate, gray and greenish-gray, thinly-bedded, fine- to medium-grained, chlorite-biotite-muscovite siliceous schists, and fine-grained, garnet-muscovite-chlorite-biotite-quartz schists. The quartzites are commonly massive, but may be foliated at minor fold hinges or layered on the millimeter to centimeter scale,the layering being due to slight variation in biotite and garnet. The muscovite schist is more abundant over the southern half on the interval. Here the muscovite schist has abundant chlorite patches, fined-grained sandy layers, and wispy biotite layers, Typically white, massive, medium- to coarse-grained quartz layers <10 cm thick are present. Atypically the quartz layers are pink due to abundant fine-grained garnets. These coarse, white quartz layers may be veins, judging from the local quartz veinlets that interconnect them. Minor folds within these quartz layers demonstrate that the massive fine-grained quartzites are internally folded. Evidence for a progressive increase in intensity of shearing toward the south is present in the form of numerous fine-grained foliated or laminated siliceous layers <30 cm thick.

The Mt. Rat Schist is exposed over the southern 185 feet of roadside. Over the first 25 feet the rock type is mainly a fine-grained, uniform, medium- to dark-gray, chlorite-muscovite-quartz-biotite schist and schistose gneiss with subordinate biotite-quartz granulite with coticule. The rocks are pin-striped with white, fine-grained, quartz layers and granular quartz layers. The coticule forms aphanitic red-brown layers and isolated ellipsoids <5 cm long. The last 160 feet of exposure consists of a very fine-grained, highly fissile, dark-brown-stained, rusty- to yellow-green-weathering, biotite-rich schist with a few rusty-brown <8 cm thick, irregular, quartz layers and pods.

**Accumulated mileage**

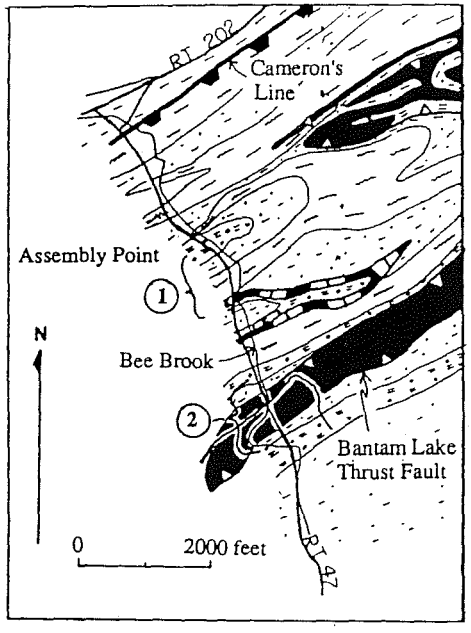
Return to the vehicles and continue south on Rt. 47 to Stop 2. If the field trip started late we will skip Stop 2 and proceed directly to Stop 3.

0.5 Stop 2: Bee Brook Bridge area. Park along the road north of the intersection of Rt. 47 and Buffum Road. The bridge over Bee Brook is south of the intersection. Walk to the northern contact of the Mt. Tom Amphibolite in the large roadcut north of Buffum Road. We will walk south along Rt. 47 to the southern contact of the Mt. Tom Amphibolite south of Buffum Road.

**STOP 2 (optional):** A 600 foot cross section of the Mt. Tom Amphibolite is exposed in two large roadcuts. The local contacts of the amphibolite with the Bee Brook Formation are concordant to the compositional layering in the country rock and the dominant foliation.

The Mt. Tom Amphibolite is a dark-green to black, fine- to medium-grained, biotite-andesine-hornblende gneiss with minor quartz, sphene, epidote and opaques. Thin epidote-rich veins are present. To the north the amphibolite is in contact with about 15 feet of muscovite schist; to the south the amphibolite is in contact with interbedded quartzite and siliceous granulite. The Mt. Tom Amphibolite ends about 800 feet to the west of these roadcuts in an apparent early fold nose. Lateral facies changes or faulting within the Bee Brook Formation may account for the lack of exact stratigraphic repetition across the axial surface of the early isoclinal fold. Largely for stratigraphic reasons to be discussed at the outcrop, the Bantam Lake thrust fault has been placed at the amphibolite contact. Two concordant, meter thick, amphibolite layers occur south of the main contact. It is uncertain whether the amphibolite is repeated in isoclinal folds, or minor thrusting, or whether these are separate amphibolite layers.

0.5 Return to your vehicles and head north on Rt. 47 to the intersection with Rt. 202



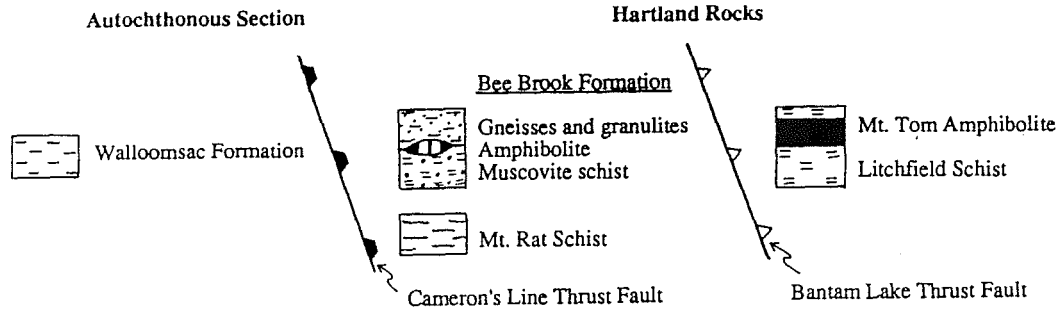
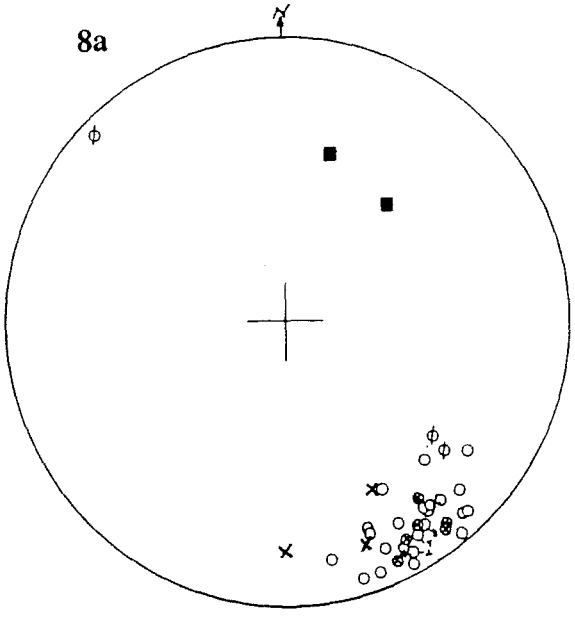




Figure 7. Geologic map of the area around Stops 1 and 2.



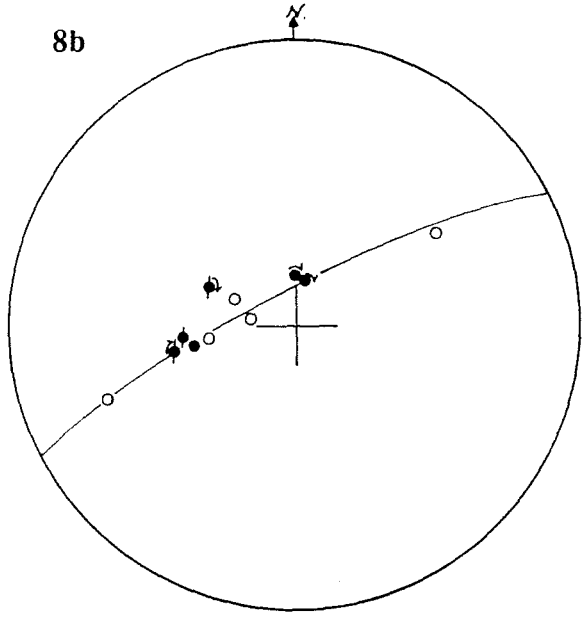




Figure 8a. Foliation data for area around Stops 1 and 2.
Poles to planar features = 45
× bedding
○ dominant D2 foliation
⊗ axial planes of minor D2 folds
ϕ axial planes of minor D3 folds
■ D5 foliation
N number of data that the symbol represents

Figure 8b. Lineation data from around Stops 1 and 2.
Lineations = 11
○ D2 mineral lineation
D2 fold axis
D3 fold axis

1.5 Turn right (east) at the intersection onto Rt. 202 and continue to Stop 3. The lowland immediately north of the Rt. 202 is underlain by the Inwood Marble which lies in the core of the major D4 Woodville anticline. The hills beyond are underlain by the Walloomsac Formation and Manhattan Schist. The Walloomsac is also present in the southern limb of the Woodville anticline immediately south of Rt. 202. The Manhattan is truncated by the Cameron's Line thrust fault in the Bee Brook valley, but in the direction of Stop 3 the Manhattan section appears and thickens between the Walloomsac and Cameron's Line.

4.5 The large rusty-weathering Walloomsac roadcut on the south side of Rt. 202 is the first station of Stop 3. Drive east of the roadcut to the intersection of Rt. 202 with Rt. 341 and Romford Road (not named on the New Preston quadrangle map). Park at the corner of Rt. 202 and Romford Road, and walk back to the roadcut.

**PLEASE WATCH OUT FOR TRAFFIC**

**STOP 3:** This traverse starts at the Walloomsac Formation roadcut near the nose of the Woodville Anticline and continues southward (Figs. 2, 9). Proceeding southward we will first cross the Lake Waramaug thrust fault into the Schistose granulite member of the Manhattan Schist, then cross the Rabbit Hill thrust fault into the Biotite-garnet schist member, and then across Cameron's Line into Hartland Group rocks and the Quartz-bearing monzonite member of the of Mt. Prospect Igneous Complex. The local section, which dips steeply northwest, lies on the overturned southeast limb of the regional D4 Woodville anticline. Please use Figure 9 in conjunction with the stratigraphic section of Figure 2.

**Station A:** The Rt. 202 Walloomsac road cut consists of fine-grained, slabby, rusty-weathering, sulfidic, sillimanite-muscovite-biotite-plagioclase-quartz schistose gneiss and fine-grained, siliceous granulites. We will now cross the unexposed Lake Waramaug thrust fault on the way to Station B.

**Station B (optional):** This is a small outcrop of the Schistose granulite member of Manhattan Schist. It is dominantly a thinly-layered, gritty-surfaced, fine-grained, garnet-muscovite-biotite-plagioclase-siliceous granulite with subordinate schistose gneiss layers. This weathered outcrop has a furrowed, gray-brown surface and a yellow- to burgundy-stained interior. We will now walk east toward Station C, crossing the Rabbit Hill thrust fault along the way.

**Station C:** Here are pegmatite-rich outcrops of an interbedded, gray, fine-grained, garnet-biotite-plagioclase-quartz schistose granulite and gray, fine- to medium-grained, rough surfaced, garnetiferous, muscovite-biotite-plagioclase-quartz schistose gneiss. White, sillimanite-quartz lenses and <4 mm garnets roughen the surface. Foliation surfaces may be rusty-brown-weathering. This outcrop has been placed in the Biotite-garnet schist member although it is not characteristic of that unit. We will now proceed southeast along the hillside and cross Cameron's Line to Station D.

**Station D:** Here are two roadside outcrops of the Mt. Prospect Formation. Here it consists of thinly-bedded, fine- to medium-grained, yellow brown to rusty-weathering, garnetiferous, muscovite-garnet-biotite-plagioclase-quartz schist or schistose gneiss. Sillimanite is apparent in coarser varieties. We will now head southwest recrossing Cameron's Line into the Manhattan Schist.

**Station E:** This is a large, characteristic outcrop of the Biotite-garnet schist member of the Manhattan Schist. It consists of medium-grained, dark-gray, garnetiferous, nubby-surfaced, muscovite-spangled, sillimanite-plagioclase-muscovite-quartz-biotite schistose gneiss. Nubs are sillimanite-quartz pods, lenses, and layers. Garnets may be up to pea size. There are granular, quartz layers < 3 cm thick that are boudinaged and lie along the foliation. At most, 5% of the outcrop is poorly foliated, biotite-plagioclase-quartz granulites with patchy mustard- or burgundy-staining, gritty surfaces <25 cm thick. We will now proceed south recrossing Cameron's Line.

**Station F:** Here are two small outcrops of Mt. Rat Schist. It is a fine- to medium-grained, in places laminated, rusty-brown, sillimanite-staurolite-garnet-muscovite-biotite-quartz-plagioclase schist. Muscovite is concentrated on foliation surfaces. Subordinately present are fine-grained, light-gray, granulites. We will now continue south.

**Station G:** Outcrops of the Bee Brook Formation are intruded by sills of quartz-bearing monzonite. The dominant country rock types are: thinly-layered, light-gray, fine-grained, siliceous granulite which is massive or laminated with thin white quartz layers, dark-gray, garnetiferous, sillimanite-garnet-muscovite-feldspar-biotite-quartz schist, and subordinate, fine-grained, white, laminated quartzite. Coarse-grained, massive, quartz layers are also

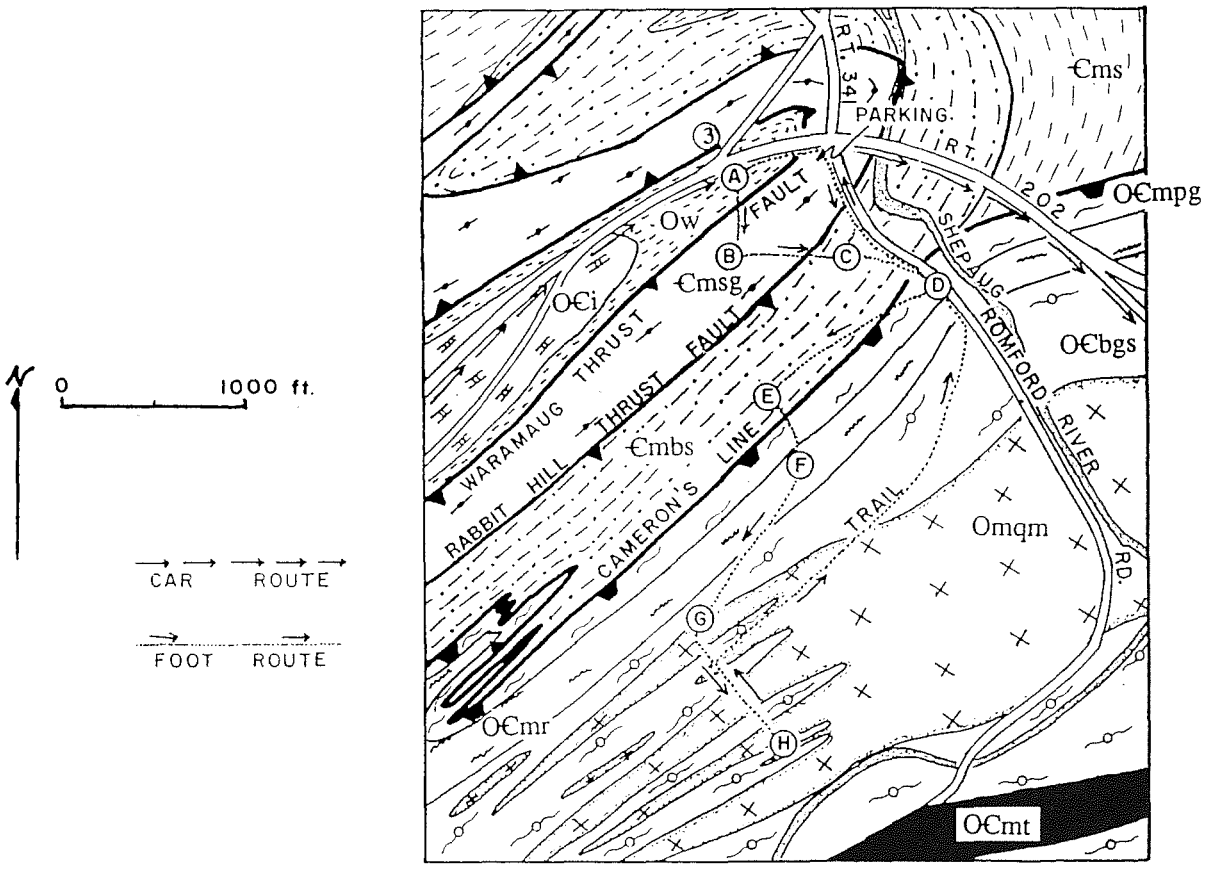



Figure 9. Geologic map of the Stop 3 traverse area.located west of the Shepaug River. Ornamentation is the same as in Fig. 2. Please refer to the stratigraphic section of Fig. 2 as needed.

present. A foot thick, concordant, amphibolite layer is present at one place. The main N60E trending Bee Brook Formation-Quartz Monzonite contact is approached obliquely as one proceeds 600 feet S45W along the line of outcrops (Fig. 9). The proportion of quartz monzonite increases to the southwest along the exposures. At the northeasternmost of the outcrops, the quartz monzonite is only present as a few < 10 cm thick porphyry sills that grade down to thinner medium-grained, wispy, non-porphyritic layers. The southeasternmost outcrop, through which the Bee Brook Formation-Quartz Monzonite member contact is drawn, is largely quartz monzonite.

**Structural Evaluation:** The southwesternmost outcrop displays evidence for four phases of deformation in both the country rock and quartz monzonite. The dominant northeast trending foliation, which is the regional D2, can be traced from the country rock into the porphyry. In the granulites and schists, the dominant foliation is a muscovite-biotite foliation. The foliation in the quartz monzonite is due to the alignment of biotite and 2-3 mm feldspar grains. Feldspar augen are in the schists adjacent to the igneous contacts. The dominant foliation wraps around the augen. This suggests that the augen and the associated quartz monzonite are older than the formation of the dominant second phase foliation. Since the quartz monzonite is the youngest major intrusive rock of the Mt. Prospect Igneous Complex, then the entire Mt. Prospect Complex was emplaced before or early in the second phase of deformation. Minor D3 isoclinal folds are present in the country rock. In at least one case a thin non-porphyritic quartz monzonite layer within the schist is folded by D3 folds and a faint foliation is axial planar to these folds. The ENE-trending regional D4 foliation lies at a low angle to the D2 foliation. In the country rock and quartz monzonite, this foliation is due to the parallelism of fine-grained quartz and feldspar. Minor north-plunging "S"-

shaped D4 folds are present in the country rock-igneous rock contacts. The rotation sense of these folds implies a major anticlinal fold to the north, the Woodville anticline. A vague northwest-trending D5 foliation is sporadically present. This foliation is a fracture cleavage with recrystallized minerals along it.

**Station H (optional):** An outcrop of the Bee Brook Formation of the Hartland intruded by quartz-bearing monzonite porphyry. The rocks at this outcrop also exhibit structural features that indicate that the quartz monzonite intruded either before or during the development of the dominant foliation. Feldspar porphyroblasts in the schist are probably the product of quartz monzonite contact metasomatism. The dominant foliation wraps around the porphyroblasts and thus probably postdates them. The dominant foliation of the schist continues into the quartz monzonite. The foliation in small Bee Brook inclusions in the quartz monzonite is continuous with that in the surrounding igneous rock. The inclusions are aligned parallel to this foliation. A wispy non-porphyritic quartz monzonite layer in the schist is folded and the dominant second phase foliation is parallel to the axial plane of the fold.

4.5 At the end of the traverse walk SE from Station G or NE from Station H until reaching the path north of the power lines. Follow this path northeast until reaching Romford Road. Walk NW on Romford Road to the vehicles. Drive east on Rt. 202. Manhattan Schist rocks, Cameron's Line, and Hartland Group rocks are crossed in this order over the next 0.9 miles.

5.1 Turn right and follow the signs for Mt. Tom State Park. We cross the folded southwest corner of the Mt. Prospect Complex and enter the Bee Brook Formation north of the entrance gate.

5.4 Pass through the entrance and follow the one-way road around a loop and to the western parking area.

6.0 Park in the western parking area.

**LUNCH**

Picnic tables, fresh water, and toilets are available.

After lunch walk from the parking lot to the four outcrops of STOP 4.

**STOP 4:** Several roadside outcrops of the Bee Brook Formation that display folded folds involving the D1, D2, D4 and D5 phases of deformation are exposed within the Mt. Tom State park about 400 ft. south of Mt. Tom Pond (Fig.10). The rock is composed of thinly-bedded, fine-grained, silvery-gray, staurolite-biotite-muscovite-quartz schist, slabby, gray, biotite-muscovite, siliceous schistose gneiss, and quartzite. Millimeter scale, fine-grained, white to slightly pink, quartz laminae are present as are coarser-grained, massive, irregular, quartz layers and lenses. If time permits we will also walk south along park trails to the summit of Mt. Tom. Along the way we will see major outcrops/cliffs of Mt. Tom Amphibolite separated by thin infolds of Litchfield Schist.

**Station A:** Here is evidence in the outcrop for five phases of deformation. A minor D1 isoclinal fold folded by a D2 fold is exposed on the north facing surface of the outcrop. The dominant schistosity with an average orientation of N75E 60NW (Fig. 11A) is axial planar to the D2 fold. A later foliation has an average orientation of N41E 53SE and consists of fine-grained quartz segregated into paper-thin, millimeter spaced folia. This later foliation cuts across the axial surface of this D2 fold and is axial planar to low amplitude, open folds in the lamination and dominant D2 foliation. This later foliation, which has been observed only south of Rt. 202, seems to maintain a SE dip (Figs. 11B,), but no known map scale folds are associated with it. This SE dipping foliation is considered to be conjugate to the NW-trending D5 folds. On top of the outcrop D5 open folds modify a hook pattern formed by the interference of D1 and D2 phase folds.

**Stations B and C:** The low angle intersection of very fine compositional layering and the schistosity produces a fine intersection lineation also marked by concentration of biotite and staurolite (Fig. 11A). A closely spaced crinkling of the schistosity seen at B is interpreted to be D3.

There are two important general features to note in Figure 11. The poles to bedding and to the dominant D2 foliation in the Mt. Tom area including STOP 4 are arrayed in a girdle (Fig. 11C), and the D2 and D3 lineations have been rotated along great circles to produce a girdle (Fig. 11D). These distributions are largely the result of D4 deformation although the effects of D3 folds, which are interpreted to subparallel the D4 folds, are uncertain. A great circle defined by the poles to schistosity around a minor open D4 fold at STOP 6 (Fig. 11A) has an orientation similar to the Figure 11C array as demonstrated by the near coincidence of the pole to the schistosity girdle (P in 11A) and the axis of the minor fold.

A kink band at B is interpreted to be a D5 feature (Fig. 11B). This kink band has a similar orientation to a sporadic, fine-grained NW trending D5 foliation seen elsewhere in the Mt. Tom area.. This foliation is due to thin

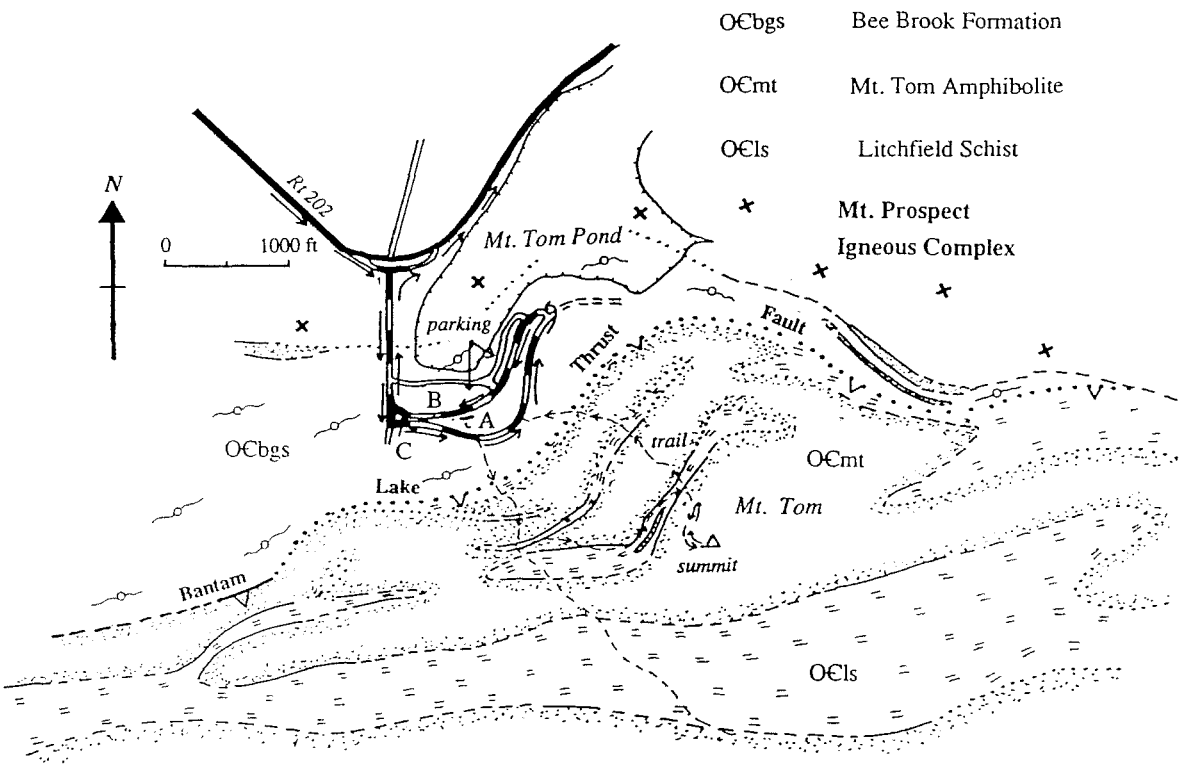



Figure 10. Geologic map of the Stop 4 area located at Mt. Tom State Park.

laminae rich in quartz and feldspar or to the preferred orientation of biotite in the schistose rocks, and to quartz-feldspar laminae or hornblende-quartz-feldspar laminae in the Mt. Tom Amphibolite. The broad, open, NNW map scale folds in the Mt. Tom Amphibolite at Mt. Tom are D5 features (Fig. 2).

6.0 Return to the vehicles, exit the park to the west, and head back to Rt. 202.

6.3 At the "T" intersection before Rt. 202 turn right (east).

6.4 Turn right (east) onto Rt. 202. The road at first subparallels the NE trending Mt. Prospect Complex/country rock contact. Primarily diorite and quartz monzonite underlie Mt. Tom Pond to the south, and the Mt. Prospect and Hartland Formations lie to the north. At the NE end of Mt. Tom Pond, the contact trends almost east and crosses the road.

7.1 As we pass the large STOP 6 roadcut we once again cross the Complex/country rock contact which at this point is trending northwest due to the Woodville anticline.

7.4 Just past the intersection with Looking Glass Hill Road, we cross the N-trending D3 Goslee syncline which is cored by the Mt. Prospect Formation. The low hills north and south of Rt. 202 have Mt. Prospect Formation at their crests and diorites on their flanks. In the 1985 NEIGC field trip the hinge area of the Goslee syncline south of Rt. 202 was visited to demonstrate major D3 folding of the Complex/country rock contact. The lack of time prevents us from visiting the hinge area today. Continue east along Rt. 202. Pass several diorite outcrops before reaching STOP 5. Mt. Prospect cannot be seen over most of the Rt. 202 route, but it is about 6000 feet north of STOP 5.

8.2 Park in the E.O. Phelps & Sons Co. driveway. Please do not block the access to the buildings. STOP 5 is the large diorite road cut and natural outcrop adjacent to the buildings.

**STOP 5:** A large roadcut along Rt. 202 of the Layered Diorite-Gabbro unit of the Diorite-Gabbro Association. The layered diorite gneiss is the most extensive unit of the Mt. Prospect Igneous Complex. At this site the rock is composed of typically 1 cm to 5 m thick layers of quartz-plagioclase-hornblende-biotite gneiss

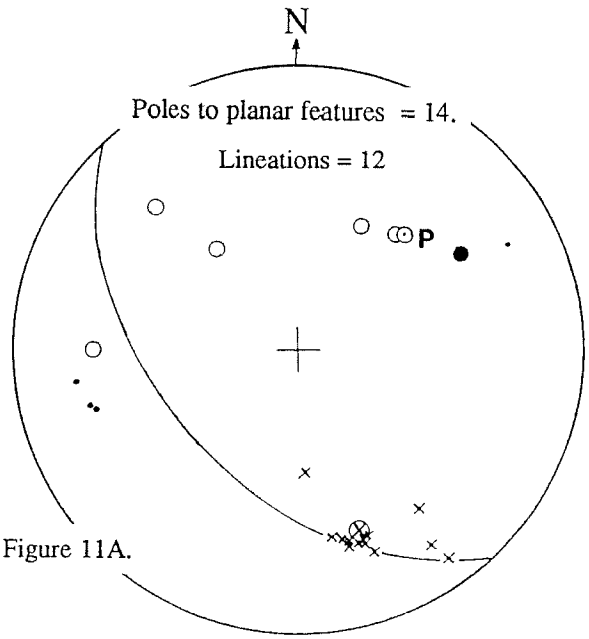



Figure 11A.

Data from STOP 4, Stations A, B, C.
- × dominant D2 foliation
- ⊗ D2 axial plane
- • D2 biotite and staurolite lineation
- ○ minor D2 fold axis
- ● D3 minor fold or crinkle of the dominant schistosity
- ⊙ axis (N48E 39) to minor D4 open fold
- P pole to the girdle of Fig. 11C

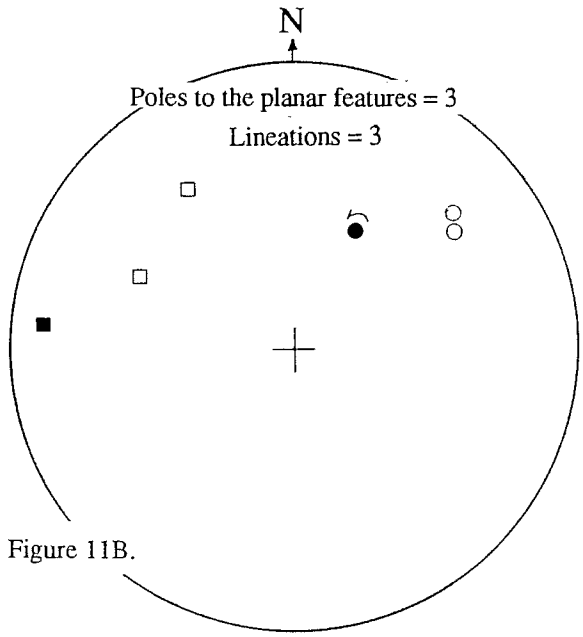



Figure 11B.

D5 data from STOP 4, Stations A and B.
- □ fine, fifth phase, quartz-feldspar foliation
- ■ D5 kink band axial plane (N5E 77E)
- ○ intersection of the D5 foliation with the D2 schistosity
- ● D5 kink band axis

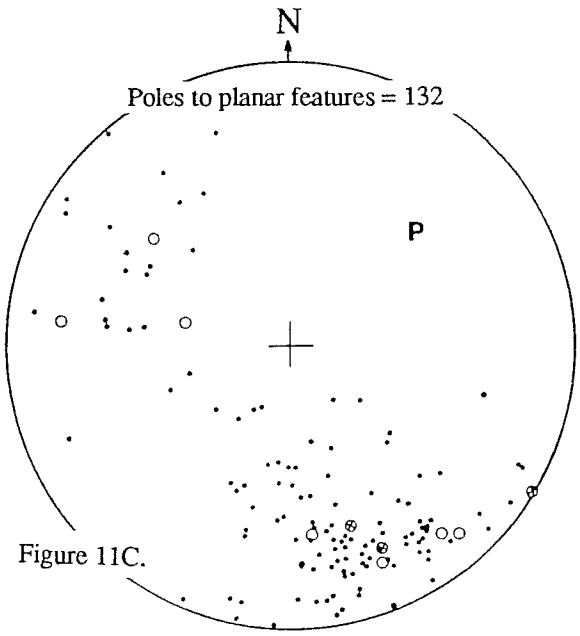



Figure 11C.

Foliation data for the Mt. Tom area including STOP 4.
- ○ bedding
- • dominant D2 foliation
- ⊗ axial plane of D2 minor fold
- P Pole (N48E 39) to the girdle defined by the poles to bedding and the D2 planar features

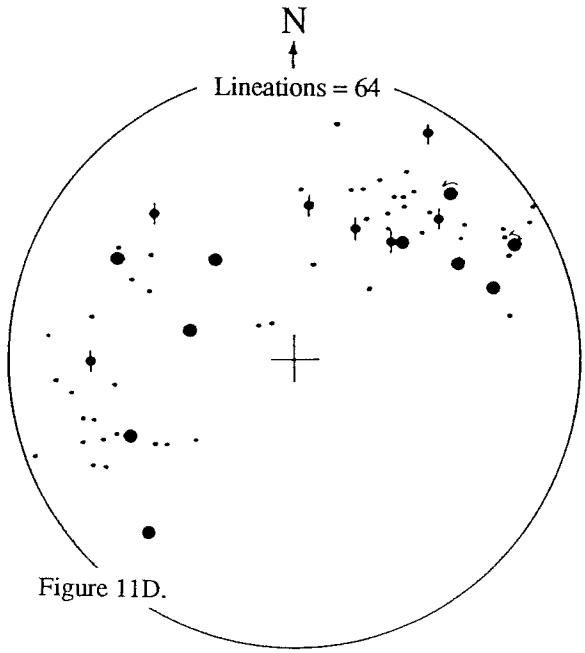



Figure 11D.

Lineation data from the Mt. Tom area.
- • D2 mineral lineation or intersection lineation
- ● D2 or D3 fold axis
- ⊕ D3 crinkle of the dominant D2 schistosity

locally with minor augite. The layering is due to differences in the relative proportions of the four major minerals, in color of the rocks from light-gray to dark-gray, and in grain size of rocks with this same mineralogy. Layer contacts may show low angle cross-cutting relationships with other layers. Mafic xenoliths are common in more felsic hosts but the reverse relationship is rare. Mafic biotite-hornblende lenses are floating in the diorite or locally concentrated along some felsic veins. The former appear to be partially assimilated xenoliths, while the latter may be cumulate patches.

Layered diorite outcrops contain at least minor amounts of intrusive biotite-andesine micropegmatite in the form of veins, irregular, wispy aggregates, discontinuous layers, sills, and rare dikes. Quartz monzonite layers < 2 cm thick are typically not porphyritic, but are rich in 2 - 4 mm, subhedral to euhedral plagioclase grains. Diorite xenoliths are common in the quartz monzonite. Though not present at this stop, microcline megacrysts are locally present in the diorite, especially near the contacts with the quartz monzonite porphyry. The localization of euhedral megacrysts only near contacts with the late intruding quartz monzonite, and the random orientation of the megacrysts in foliated diorite suggest that the microcline megacrysts in the diorite are porphyroblasts.

Foliations are present consisting of 1-3 mm, aligned hornblende, biotite and andesine grains or aggregates of grains. The hornblende and andesine grains may be disseminated in the rock, which is the more general case, or concentrated in wispy lenses. The general impression is that these foliations are subparallel to the roughly E-W trending, compositional layers, but in fact low angle intersections among the foliations and compositional layers are common. The layering and foliations together produce the dominant, broadly curved surfaces in the outcrops. Locally the hornblende weathers out to produce either a pitted surface or a crude parting in the rock. Most lineations plunge moderately N to NE. This alignment is due to D4 deformation, and the major D4 Aspetuck syncline lies north of STOP 5 (Fig. 2). These lineations roughly parallel the overall plunge of the Mt. Prospect Igneous Complex.

There are several, NE- to NW-trending, minor faults in the layered diorite on both sides of the road which locally show an actinolite lineation or slickensides. Displacement is difficult to demonstrate at this outcrop, but these faults postdate the previously mentioned foliations. The faults are in turn cut by randomly oriented, 1 mm wide, felsic veins which may be single and straight or in anastamosing groups.

8.2 Return to the vehicles and proceed west on Rt. 202 to STOP 6.
9.4 Drive to the west end of the large STOP 6 roadcut and park along the side of the road.

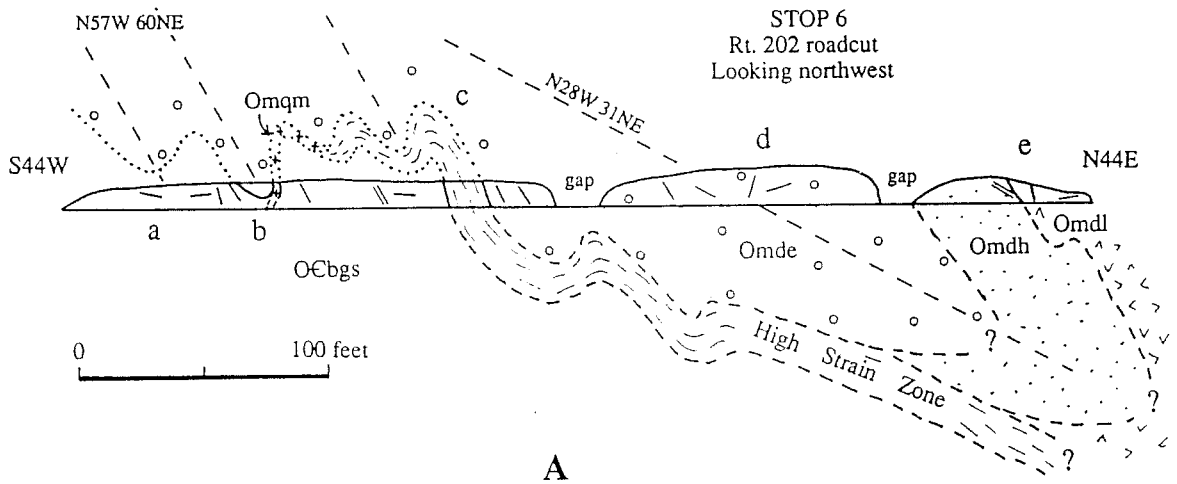
**This is a very interesting outcrop but also a very dangerous one due to the lack of a shoulder to the road, SO PLEASE BE CAREFUL. If a large group of people attends this field trip, we may have to forego STOP 6.**

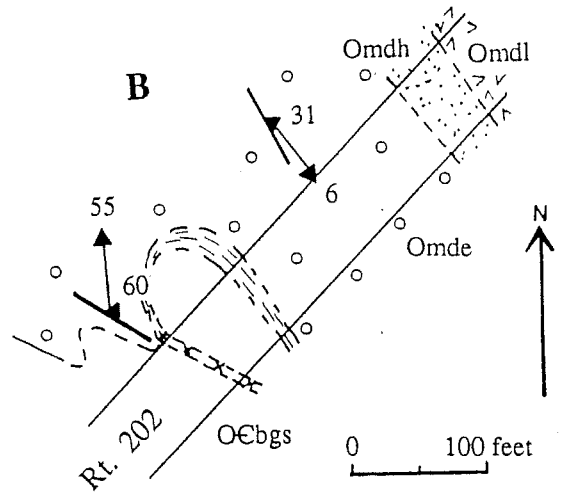
At STOP 6 we will observed a sheared contact between the Bee Brook Formation and Diorite-Gabbro Association rocks of the Mt. Prospect Igneous Complex. While this contact has not been mapped as a major fault, it is unclear how much displacement has taken place. The contact, the country rocks and the Complex rocks have been deformed by D3 and D1 or D2 folds.

The roadcut covers a distance of 400 feet along both sides of Rt. 202 (Fig. 12). Rocks of the Bee Brook Formation are exposed over the southwest third of the roadcut; rocks of the Mt. Prospect Igneous Complex are exposed over the northeast two thirds of the roadcut and at point 'b' (Fig. 12A). The sheared country rock/Complex contact is exposed at 'c' and at 'b' where the igneous rocks are interpreted to be in the core of a small syncline. Another syncline only involving the country rocks is present at 'a'. The contacts and structures project southeast to the other side of the roadcut, however the igneous rock core at 'b' is not present across the road, it being interpreted to project above the outcrop. The basic structure then is interpreted to be an E-dipping country rock/Complex contact that has been folded by NW-trending, NW-plunging folds.

The Bee Brook Formation here consists of fine-grained muscovite-biotite-feldspar siliceous granulites, gneisses and schists. In places where the biotite and muscovite are fine-grained and disseminated, the granulite appears massive. Elsewhere centimeter-thick biotite-rich layers and thin coticule layers are locally present.

The Epidote-Diopside Diorite-Gabbro member, which is the dominant igneous rock at STOP 6, is in contact with the Bee Brook. This member is a fine- to medium-grained biotite-hornblende-plagioclase gneiss with minor epidote, diopside, sphene, scapolite and sulfides. The diopside, which is a distinctive bright green is both



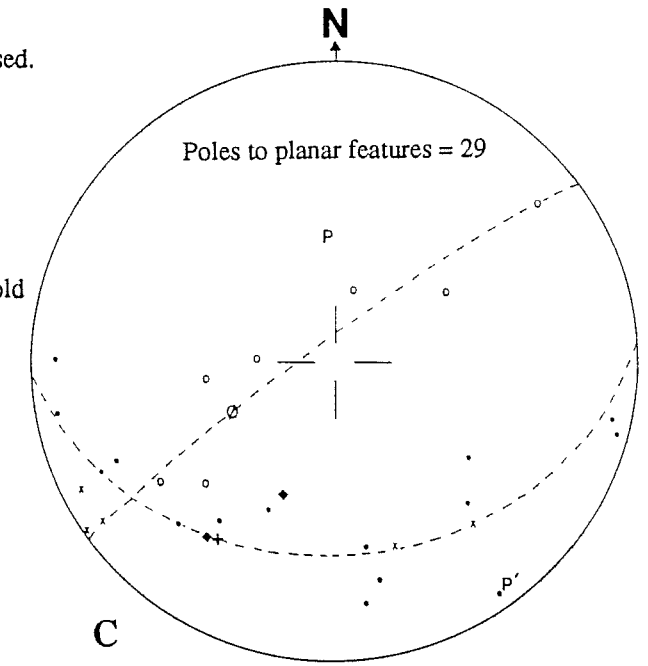




Figure 12. The STOP 6 roadcut.

A. Sketch of the roadcut on the northwest side of Rt. 202 with an interpreted cross section superimposed. Formation abbreviations as in Fig. 2. Reference points a to e are used in the text.

Traces of foliation on plane of section:

- — Compositional layering and early foliation
- = D3 foliation

Speculative axial planes:

- _ _N57W 60NE_ _

B. Sketch map of the STOP 6 roadcut. Foliation and lineation symbols refer to the axial planes and fold axes of speculative early and D3 folds

C. Equal area net plot of foliation data from STOP 6.

Poles to compositional layering and early foliation in the country and Complex rocks:

- • Data over the interval from a to c and around e.
- o Data over the interval from c to e
- × Poles to the country rock/Complex contacts and contacts between igneous members.
- ◆ Poles to D3 foliation.
- P Pole (N3W 55°) to the great circle fitted to the data from the interval a to c and around e.
- P′ Pole (S37E 6°) to the great circle fitted to the data from the interval c to e.
- + Best fitted D3 axial plane (N57W 60NE)
- ø Speculative early axial plane (N28W 31NE)

disseminated through the rock and in irregular, centimeter-scale patches with plagioclase and sphene. Plagioclase and biotite contents are quite variable. Weathered surfaces show how plagioclase rich some varieties are. Locally, such as the core of the syncline at point b, the rock is a biotitite. The Epidote-Diopside Diorite-Gabbro member grades to the northeast into a massive, medium-grained hornblende-plagioclase-biotite gneiss with minor garnet and sphene but without the distinctive green diopside. Overall this unit, the Homogeneous Diorite-Gabbro, is more feldspathic and biotite-poorer than the Epidote-Diopside Diorite-Gabbro. The Homogeneous Diorite-Gabbro sharply grades to the northeast across a contact complicated by faulting into rocks of the Layered Diorite-Gabbro unit. Here this unit is a gray, quartz-plagioclase-hornblende-biotite gneiss with irregular andesine-rich patches and layers.

The dominant early foliation in the country rock is a muscovite-biotite foliation that is parallel to the compositional layering or intersecting it at a low angle, that is axial planar to minor isoclinal folds in the layering and that is folded along with the layering and the igneous-country rock contact by NW-trending folds. The foliation is seen at 'b' to be parallel the country rock/Complex contact.

Within 5-10 meters of the country rock/Complex contact at point 'c' the country rock consists of a variety of biotite-feldspar-quartz-muscovite-bearing mylonitic rocks interlayered on the 1-10 centimeter scale. Included are fine-grained, laminated gray quartzite layers, layers of anhedral feldspar porphyroclasts in a biotite-rich matrix, plus thinly laminated schistose gneiss and fine-grained, rusty-olive-weathering phyllite with feldspar porphyroblasts. Within half a meter of the contact the igneous rock is biotite-rich and well-foliated. At one place this foliation is axial planar to a minor fold in the igneous layering. How much younger the mylonitic foliation is than the regional foliation away from the contact is uncertain, both could be D2 or D1. The country rocks at 'b' do not seem as mylonitic. Instead a relatively unstrained 50 cm dike of porphyritic diorite with large feldspar megacrysts is present at the igneous contact at 'b'. The dike is also present across the road away from the folded igneous contact which projects into the air.

An equal area plot (Fig. 12C) includes the compositional layering and early folded foliation of the country rock, the country rock/Complex contact, and the igneous layering and foliation. This data from points 'a' to 'c' and point 'e' fit a great circle with a pole P of N3W 55°; this is interpreted to be the average axis of the folds deforming the igneous contact and foliation. Based on projection across the road the corresponding axial planes have a northwest trend. A fine NE-dipping quartz/feldspar foliation (shown as diamonds in Fig. 12C) that is sporadically present in the country rock and a feldspar foliation in the layered diorite is interpreted to be the axial plane foliation of these folds. A best fit axial plane of N57W 60 NE is shown in Figs. 12B and C. These folds, which fold the dominant foliation, are considered D3. Minor open folds with NW-dipping axial plane foliation and NW-plunging axes (not shown in Fig. 12C) are considered related to the regional D4 Aspetuck Syncline and Woodville anticline. Additional, faint SW- and SE-dipping foliation, sporadically present in the country rock are considered D5.

Several observations suggest that there has been a section faulted out at the country rock/Complex contact. First, except for the diorite dike, there is no intrusive igneous contact exposed in this outcrop. Second, there are no country rock inclusions in the igneous rocks. Third, this is the only location where the Epidote-Diopside Diorite-Gabbro unit is in direct contact with country rock. Typically this unit is within the Homogeneous Diorite-Gabbro unit; the Layered Diorite-Gabbro and Quartz-bearing Monzonite members are commonly in contact with the country rock.

The poles to foliation and igneous layering in the interval between 'c' and 'e' have a completely different great circle distribution with a pole P' of S37E 6°. Despite the structural measurements taken, the Epidote-Diopside Diorite-Gabbro is overall massive in appearance and it is difficult to make out any outcrop-sized structure. One possible interpretation is that a major early D2 fold with shallow S37E 6° plunge is present in the eastern half of the roadcut and this speculative interpretation using a N28W 31NE axial plane is illustrated in Fig. 12. In this interpretation the deformed foliation would be D1.

9.4 Return to the vehicles and drive east again on Rt. 202.
10.2 Before reaching STOP 5 turn left onto Old Turnpike Road.
10.3 Turn north onto Prospect Road and continue north.
11.7 Round a sharp corner to the right and park by the abandoned sulfide mine.

**STOP 7:** STOP 7 involves traverses on both sides of Prospect Road. We will examine the country rocks west of Prospect Road to demonstrate that the Complex does not cross Cameron's Line. The rest of our time will be spent on an easterward traverse up the Mattatuck Trail to the summit of Mt. Prospect. We will see a number of

what country rocks units are in contact with the Mt. Prospect Igneous Complex, to evaluate whether the contacts are the result of igneous intrusion or fault emplacement, and to determine if the Complex cross cuts the Cameron's Line thrust fault. Detail maps of the geology around Mt. Prospect will be made available at that time. Some of the more important stations are now briefly described.

**Station A:** The abandoned, flooded sulfide mine east of Prospect Road. Dark greenish-gray, massive, jointed olivine norite is present at the southern entrance. At the northwest end of the pit the walls are rotten and rusty from weathered, disseminated sulfides. Above the flooded shaft to the east is a breccia in which the host rock has been intruded by a sulfide rich phase. The host rock at this point is hornblende-rich. Late fine-grained diorite dikes have intruded the olivine norite and may have brought in the sulfides.

**Station B:** This is a group of small outcrops lying along 150 feet of the Mattatuck trail about 800 feet southeast from Station A. Here is a somewhat bewildering range of igneous rock types that have intruded one another. Basically, there are olivine norites and hornblende-rich norites/gabbros that share both sharp and more obscure gradational contacts. An igneous breccia is present in one outcrop that consists of a norite that has been fragmented and intruded by a coarse-grained feldspathic magma possibly derived from nearby quartz-bearing monzonite. Within this breccia are small exotic fragments of country rock quartzites.

**Station C:** At the summit of Mt. Prospect several discordant Quartz-bearing Monzonite Porphyry bodies intrude rocks of the Norite-Gabbro Association, the Cortlandtite-Mafic Gabbro Series, Layered Diorite-Gabbro and Epidote-Diopside Gabbro, although many of the contacts are not well exposed. The contact with the Norite-Gabbro Association rocks are especially obscure being gradational over tens of meters. As the norite is approached the microcline megacrysts in the porphyry disappear and the matrix becomes more mafic. Here at the summit the Quartz-bearing Monzonite forms irreugular bodies that intrude along contacts of the mafic rock types and engulf small igneous bodies and country rock inclusions.

**Station D:** Further southeast along the crest of Mt. Prospect where abundant outcrop permits us to observe locally well developed Diorite-Gabbro Association layering. The layering consists of 1-40 meter thick layers of Layered Diorite-Gabbro, Homogeneous-Diorite Gabbro, Epidote-Diopside Gabbro, tabular country rock inclusions up to 100 meters long and concordant sills of quartz-bearing monzonite porphyry. Minor high angle faults displace the layering.

Additional stations will be described and all stations located on the geologic map to be handed out during the field trip. The outcrops visited today only begin to illustrate the complex geology of the Mt. Prospect Igneous Complex and surrounding country rocks. Additional outcrops are described in the 1985 NEIGC guidebook (Panish and Hall, 1985).

**I hope you have had an interesting and enjoyable field trip! Thank you!**

## REFERENCES CITED

Agar, W.M., 1927, The geology of the Shepaug aqueduct tunnel, Litchfield County, Connecticut: Connecticut Geological and Natural History Survey, Bulletin 40, 46 p.

Cameron, E.N., 1943, Origin of sulphides in the nickel deposits of Mount Prospect, Connecticut: Geological Society of America Bulletin, v. 54, p. 651-686.

Cameron, E.N., 1951, Preliminary report on the geology of the Mt. Prospect Complex: Connecticut Geological and Natural History Survey Bulletin 76, 44 p.

Dana, R.H., Jr., 1978, Stratigraphy and structural geology of the Lake Waramaug area, western Connecticut: MS thesis, Department of Geology and Geography, University of Massachusetts, Amherst, 108 p.

Gates, R.M., 1951, Bedrock geology of the Litchfield quadrangle: State Geological and Natural History Survey of Connecticut, Miscellaneous Series No. 3, 13 p.

Gates, R.M. and Bradley, W.C., 1952, The geology of the New Preston quadrangle: Connecticut Geological and Natural History Survey Quadrangle Report No. 2 (Miscellaneous Series 5), 55 p.

Hall, L.M., 1980, Basement-cover relations in western Connecticut and southeastern New York. In Wones, D.R. (Ed.), The Caledonides in the U.S.A.: I.G.C.P. Project 27: Caledonide Orogen, 1979 Meeting, Blacksburg, VA, Virginia Polytechnic Institute and State University, Memoir No. 2, p. 299-306.

Hall, L.M. and Robinson, Peter, 1982, Stratigraphic-tectonic subdivisions of southern New England. In St. Julien, P. and Beland, J. (Eds.), Major Structural Zones and Faults of the Northern Appalachians, Geological Association of Canada Special Paper 24, p. 15-41.

Koziol, A.M. and Newton, R.C., 1988, Redetermination of the anorthite breakdown reaction and improvement of the plagioclase-garnet-$Al_2SiO_5$ geobarometer: American Mineralogist, v. 73, p. 216-223.

Merguerian, C.M., 1983, Tectonic significance of Cameron's Line in the vicinity of the Hodges Complex - an imbricate thrust model for western Connecticut: American Journal of Science, v. 283, p. 341-368.

Merguerian, C.M., 1985, Geology in the vicinity of the Hodges Complex and the Tyler Lake Granite, West Torrington. In Tracy, R.J. (Ed.), Guidebook for Connecticut and adjacent areas of New York and Rhode Island, N.E.I.G.C., 77th Annual Meeting, p. 411-442.

Merguerian, C.M. and Ratcliffe, N.M., 1977, A reinterpretation of the Hodges Mafic Complex and its relation to deformation along Cameron's Line in West Torrington, Connecticut: Geological Society of America Abstracts with Programs, v. 9, p. 301.

Panish, P.T., 1989, Tectonic and metamorphic setting of the Mt. Prospect Igneous Complex, western Connecticut: Ph.D. dissertation, Department of Geology and Geography, University of Massachusetts, Amherst, 292 p.

Panish, P.T. and Hall, L.M., 1985, Geology of the Mt. Prospect region, western Connecticut: In Tracy, R.J. (Ed.), Guidebook for Connecticut and adjacent areas of New York and Rhode Island, N.E.I.G.C., 77th Annual Meeting, p. 443-489.

Robinson, Peter and Hall, L.M., 1980, Tectonic synthesis of southern New England. In Wones, D.R. (Ed.), The Caledonides in the U.S.A.: I.G.C.P. Project 27: Caledonide Orogen, 1979 Meeting, Blacksburg, VA, Virginia Polytechnic Institute and State University, Memoir No. 2, p. 73-82.

Williams, M.L. and Grambling, J.A., 1990, Manganese, ferric iron, and the equilibrium between garnet and biotite: American Mineralogist, v. 75, p. 886-908.

Wise, D.U., 1981, Fault, fracture and lineament data for western Massachusetts and western Connecticut: United States Regulatory Commission, NUREG/CR-2292, 253 p.

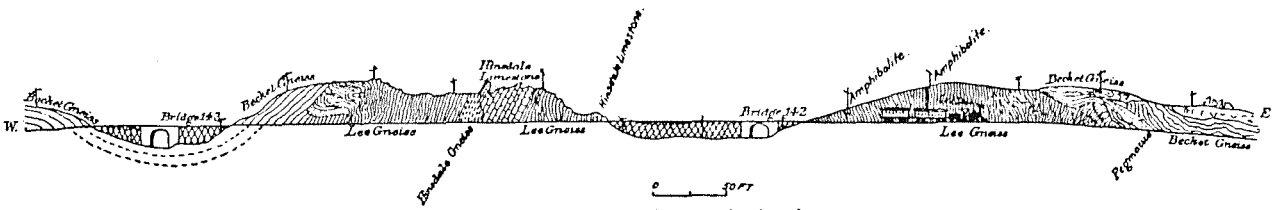



FIG. 1.—Algonkian section at Coles Brook.

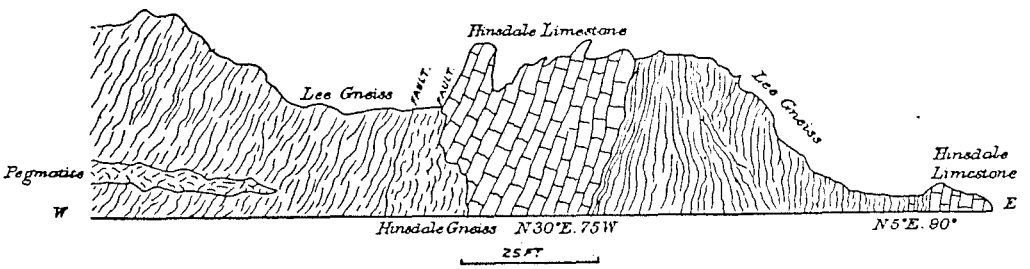



FIG. 2.—Detailed section of the limestone at Coles Brook.

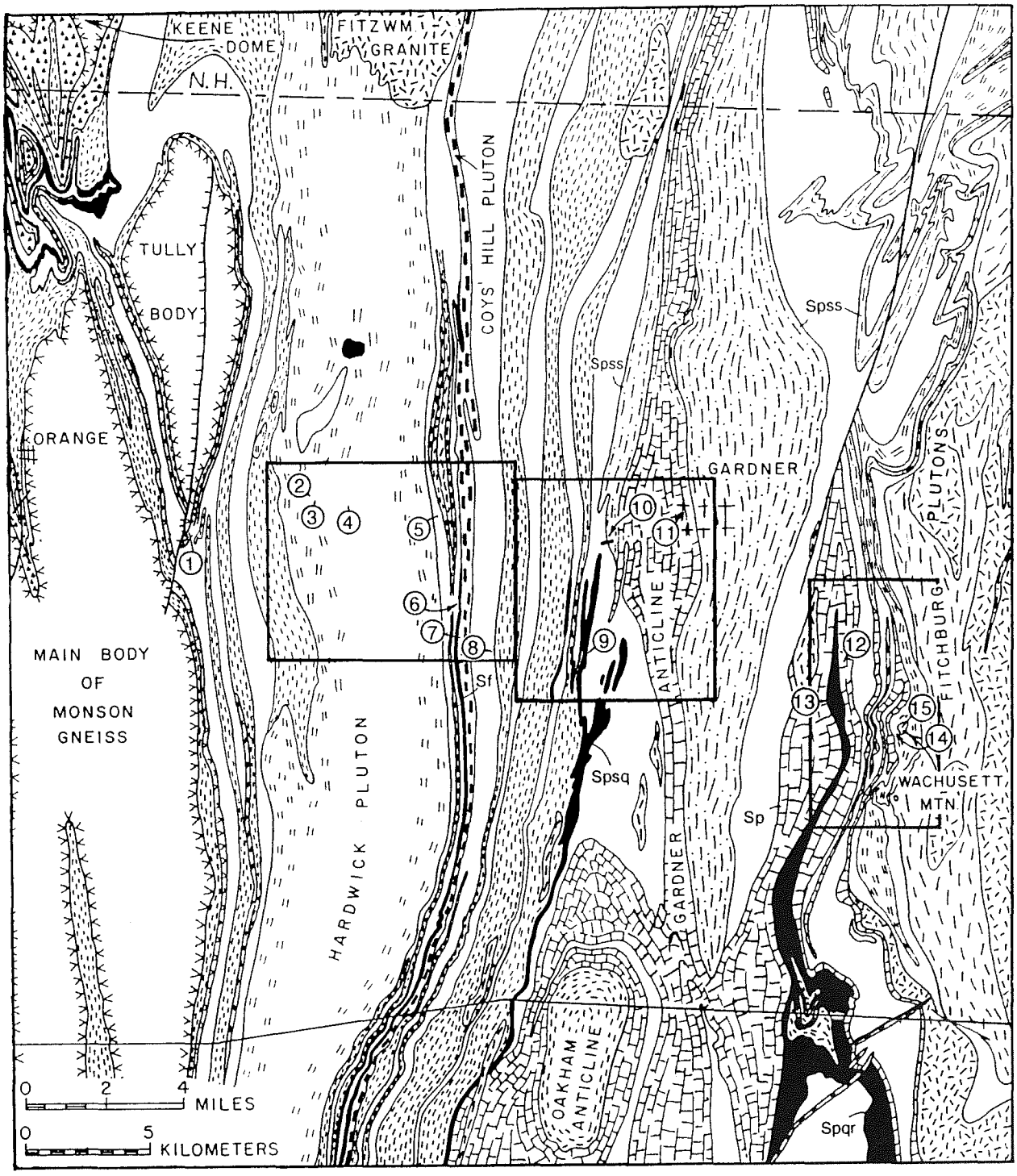



Figure 1. Generalized geologic map of the northern part of central Massachusetts showing the field trip area, stop locations, and locations of more detailed maps (modified from Robinson, 1979 and Robinson and Goldsmith, 1991). Section line in southern part of map is that of Section D-D' of the State bedrock map (Zen et al., 1983).

# A TRAVERSE ACROSS THE "WILD UNKNOWNS" OF NORTH-CENTRAL MASSACHUSETTS FROM THE BRONSON HILL ANTICLINORIUM TO THE FITCHBURG PLUTONS

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## PURPOSE OF TRIP

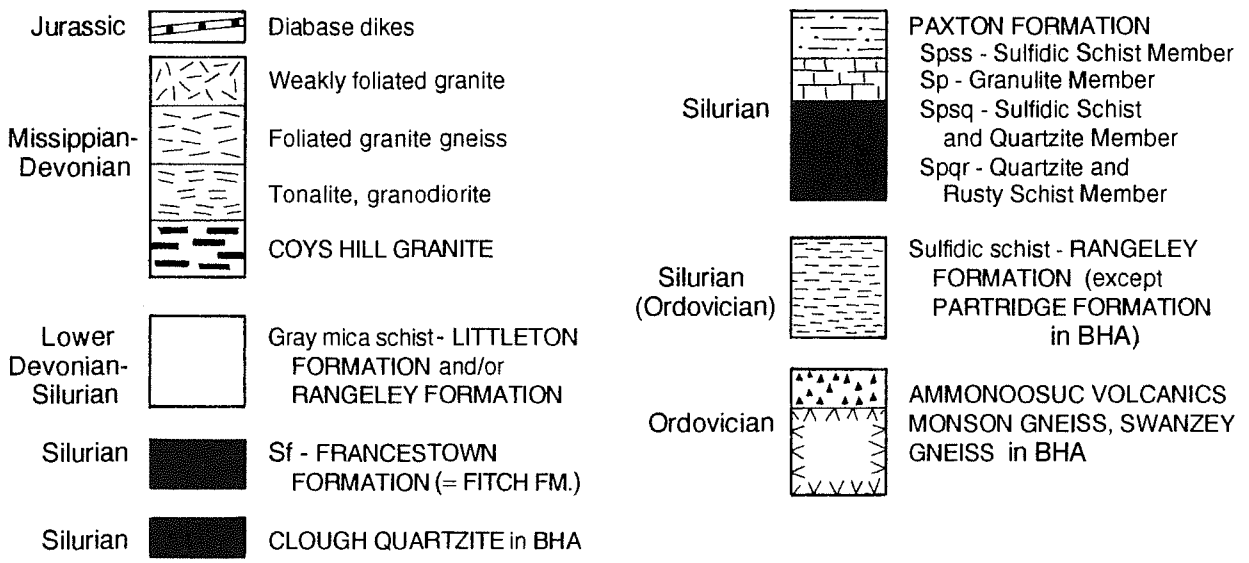
The purpose of this trip is to review what is known about the metamorphic and igneous rocks of north-central Massachusetts, between the Bronson Hill anticlinorium and the west margin of the Fitchburg plutons, in the northern part of the central Massachusetts metamorphic high (Figure 1). Much of this information was gathered during general and detailed reconnaissance done in preparation for the Bedrock Geologic Map of Massachusetts (Zen et al., 1983). In contrast with southern Massachusetts, except for the Athol Quadrangle (Mook, 1967), the Barre area (Tucker, 1977), the Ashburnham-Ashby area (Peterson, 1984), and the Royalston area (Springston, 1990 no quadrangle-scale geologic maps have ever been completed, although the authors did devote many days to detailed mapping in certain favorably exposed locations. The reconnaissance work has been supplemented by topical studies on the petrology and geochemistry of the Hardwick pluton by Shearer (1983; Shearer and Robinson, 1988), on the structure and petrography of the igneous breccia at Brooks Village by Morton (1985), on the petrology and geochemistry of the Fitchburg plutons by Maczuga (1981), and on the phase relations of sulfide-rich schists by Tracy and Robinson(1988) . All of these works will be referred to here, and the first three will be available to all participants.

The trip can be divided geologically into four parts: A) The geology at the extreme eastern edge of the Bronson Hill anticlinorium (Stop 1); B) Igneous rocks, contact metamorphism and structural relations near the Hardwick pluton (Stops 2-5) C) Stratigraphic relations, metamorphism and structure on the west limb and center of the Gardner foliation arch (Stops 6-11); and D) Stratified rocks and overlying Fitchburg plutons on the east limb of the Gardner foliation arch (Stops 12-15). Most of these stops (except Stop 12) can be found without guidance, and several will probably be omitted on the actual NEIGC trip, depending on the interests of participants.

## STRATIGRAPHIC-TECTONIC SUBZONES

In its simplest aspect, this field trip may be thought of as a traverse across an enormous open anticline, the Gardner arch. This seems incongruous, in that the area is commonly referred to as the Merrimack synclinorium. At the western end of this traverse the strata of the Bronson Hill anticlinorium are overturned to the east at an angle of

Explanation of map patterns in Figures 1, 2, 3, and 5.



about 60°, and in fact this eastward overturning is characteristic all along the east margin of the anticlinorium in Massachusetts. This dip flattens eastward to the center of the Gardner arch, and from there eastward to the center of the Fitchburg plutons, the foliation dips to the east, so that the plutonic rocks structurally overlie the stratified rocks. In southern Massachusetts, in contrast, west dips prevail except in very limited areas, and a major unsolved problem is to understand what happens to the broad foliation arches and downwarps of northern and central Massachusetts as they are followed south (see Berry, this guidebook).

For convenience of description, the rocks of north-central Massachusetts have been divided into stratigraphic-tectonic subzones (Robinson, 1979; Robinson and Goldsmith, 1991). The subzone boundaries are somewhat arbitrary and may ultimately have little tectonic significance. They are modified somewhat here from earlier publications.

### Ware Subzone

In this report, the Ware subzone is taken to include strata and intrusions as far west as the Brennan Hill thrust (see Robinson and Elbert, this guidebook) and as far east as the westernmost occurrences of the Paxton Formation. It includes the Hardwick Tonalite intrusion (Shearer, 1983; Shearer and Robinson, 1988), the largest single exposed intrusive unit in Massachusetts, that is equivalent to the Spaulding plutons in New Hampshire (Duke, 1978). It also includes the Gneiss of Ragged Hill (Field, 1975; Peterson, this guidebook), and the long, attenuated Coys Hill Granite (Field, 1975) that is identical to the Kinsman Granite in New Hampshire (Clark and Lyons, 1986, P. J. Thompson, 1985).

The Ware subzone is dominated by alternating belts of gray and rusty-weathering pelitic schists. These were originally taken respectively to be Devonian Littleton Formation in isoclinal synclines and Ordovician Partridge Formation in isoclinal anticlines (Zen et al., 1983). These schists are now mostly taken to be members of the Lower Silurian Rangeley Formation, including the westernmost gray schist in close proximity with Monson Gneiss of the Bronson Hill anticlinorium, that was previously described as the easternmost Littleton Formation of the Connecticut Valley belt (Hatch et al., 1988) and is now known to be Rangeley Formation on the basis of detailed mapping and conglomerate lenses (P. J. Thompson, 1985; Robinson et al., 1988; Springston, 1990). Although the older stratigraphic assignments are changed, the synclinal nature of some of the gray schist belts east of the Coys Hill Granite is locally confirmed by graded bedding (Tucker, 1977; Robinson et al., 1982).

Within the Ware subzone, the strata immediately west of the Coys Hill Granite are special because they are representative of the upper part of the Silurian sequence as defined by P. J. Thompson (1985) in the Monadnock area. These include the distinctive Francestown Formation pyrrhotite calc-silicate rocks called Fitch Formation (Sfs) at Stop 7 and the big garnet gneisses at Stops 6 and 7 that are now tentatively correlated with an upper member of the Warner Formation (Robinson and Goldsmith, 1991), although shown as Littleton Formation on the State Map. The main belt of big garnet gneiss is proved to be synclinal on the basis of graded bedding at Ragged Hill (Robinson and Goldsmith, 1991). As compared to the Monadnock area along strike to the north, the Perry Mountain Formation and normal members of the Warner Formation are apparently absent. Based on Thompson's interpretation of the Monadnock area, his Chesham Pond thrust is tentatively placed along the west margin of the Coys Hill Granite. In keeping with his work in New Hampshire, the highest grade metamorphic rocks in Massachusetts lie east of or a short distance west of this postulated nappe-stage thrust that carried Lower Silurian Rangeley Formation and Kinsman Granite over folded Silurian and Lower Devonian strata to the west. However, Field (1975) has located several small lenses of Francestown Formation on the east side of the Coys Hill Granite.

### Gardner Subzone

The Gardner subzone is drawn to include the westernmost occurrences of members of the Paxton Formation and extends east to an ill-defined fault zone along the west margin of the Wachusett Mountain subzone (Peterson, 1984). It includes rocks on both sides of the broad foliation arch that dominates the region. The name Paxton Schist of B. K. Emerson (1917) was adapted as Paxton Formation for the Massachusetts map (Zen et al., 1983) to include all units of then presumed Silurian age in this part of the State.

A key marker is the sulfidic schist and quartzite member (Spsq) of the State map, which in northern Massachusetts is confidently correlated with the Smalls Falls Formation of Maine and New Hampshire (Hatch et al., 1983). This unit consists of interbedded sillimanite quartzites and schists with high pyrrhotite and locally pyrite content, extremely Mg-rich biotite and cordierite, and rutile. The western contact against less extremely sulfidic

garnet-bearing schist may be a normal stratigraphic contact with underlying Rangeley Formation. The unit is continuous from southern Massachusetts to a point just south of Route 2 where it pinches out (details in Figure 3).

The next unit to the east is a rather massive gray garnet-biotite-sillimanite schist, locally with calc-silicate layers and lenses, which was assigned to the Littleton Formation on the state map, but has many characteristics of gray members of the Rangeley. The contact of this unit with the structurally overlying Smalls Falls Formation will be seen at Stop 9 and the unit itself will be viewed in an extensive road cut at Stop 10. It pinches out to the south between the Smalls Falls Formation to the west and the normal Paxton Formation Granulite Member to the east, giving rise to speculation that this unit is bounded on one side by a thrust (Berry, this guidebook).

The next unit is one of sulfidic mica-garnet-sillimanite schists with subordinate gray biotite granulite with green calc-silicate rock. This is assigned to a sulfidic schist member (Spss) member of the Paxton Formation and will be seen on Stop 10. Near Stop 10 it contains one boudin of metamorphosed gabbro with coarse relict ophitic texture. This unit pinches out just south of Route 2 between the gray schist to the west and the normal Paxton granulite to the east, again leading to speculation concerning faults.

The center of the Gardner anticline (Stop 11) is occupied by the Granulite Member of the Paxton Formation, normally a well layered biotite-plagioclase granulite with subordinate layers of green or pink calc-silicate granulite, and in some regions interbedded with sulfidic mica schist. The typical Paxton bears considerable resemblance to the Upper Silurian Warner Formation of the Monadnock area and the Upper Silurian Madrid Formation of northwestern Maine. When in sequence with Smalls Falls rock types, it is tempting to think of such a correlation for the Paxton Granulite Member. However, similar rock types are also known in the Lower Silurian of central Maine (Osberg, 1980) in such units as the Vassalboro Formation, etc., and, in southern Massachusetts, Berry (1989 and this guidebook) has included granulites with interbedded schists in a granulite-rich member of his Lower Silurian Rangeley Formation. Taken together, these ideas may suggest that the Paxton Formation includes eastern calcareous facies of much of the Silurian succession. This is a problem that will persist, as it has in Maine, where the rocks are only at low metamorphic grade.

On the east limb of the Gardner anticline the Paxton granulite is physically overlain by a similar succession of sulfidic schist (Spss) and gray schist (Dl on the State map) which marks the general eastern limit of the Gardner subzone. The intrusive rocks of the Gardner subzone are mainly tonalitic to granitic gneiss sills that have been little studied petrographically.

**Wachusett Subzone**

The strata of the Wachusett subzone generally dip east beneath the Fitchburg plutons, a series of quasi-concordant sheets of variably foliated granites and tonalites occupying the center of a broad syncline in foliation. The repetitions of strata are believed to be in a series of isoclinal folds rooted to the west, i.e., on the west limb of the Gardner arch. The strata have close affinities to the Smalls Falls Formation and adjacent strata of the western part of the Gardner subzone.

The tentatively oldest stratigraphic unit is the quartzite and rusty schist member of the Paxton Formation (Spqr). This occupies a narrow belt interpreted as an isoclinal recumbent anticline. The belt appears to terminate to the north (Figure 1) and to the east south of the Fitchburg plutons (outside area of Figure 1), indicating that the proposed fold axis trends approximately N20W and the fold is directed from the west toward the east. The center of this poorly exposed belt is dominated by 5-10 cm beds of impure rusty-weathering quartzite alternating with pyrrhotite mica schist. Although generally similar to the Smalls Falls Formation to the west, the schists are not known to contain, on very limited petrographic data, the extremely magnesian assemblages found in the Smalls Falls. In those few places where contacts are exposed, the quartzite appears to grade either upward or downward over a thickness of 5 to 15 meters into predominantly sulfidic schist adjacent to the contact of the Paxton Granulite Member (Sp). This is true particularly at Paxton Falls, outside the area of Figure 1, which is taken to be B. K. Emerson's type locality for the Paxton, where the Paxton Granulite Member structurally underlies the Quartzite and Rusty Schist Member (Spqr). The same relations will be seen in abbreviated form at Stop 13.

The Paxton Granulite Member (Sp), as exposed below and above the Quartzite and Rusty Schist Member (Spqr) in the Wachusett subzone (see Stops 12 and 13), is dominated by slabby biotite-feldspar granulite with numerous biotite-rich partings and subordinate 2-6 cm layers of green diopside calc-silicate. It lacks the abundant interbeds of

sulfidic schist found in some other parts of the Paxton. The Granulite Member is in turn structurally overlain and underlain by gray mica schist (Dl) with minor micaceous quartzite.

Taken together, the succession just described closely resembles in attenuated form a sequence described by Ludman and Griffin (1974) and Osberg (1980) on the Kennebec River in the Skowhegan area of Maine. The lowest unit in the Maine sequence is an unnamed rusty-weathering quartzite tentatively correlated with the Perry Mountain Formation. This is overlain by the sulfidic schists of the Parkman Hill Formation that is correlated with the Lower to Middle Silurian Smalls Falls Formation, then by the biotite-feldspar granulite of the Fall Brook Formation that is correlated with the Upper Silurian Madrid Formation, and finally by the gray mica schist of the Carrabasset Formation that is correlated with the Lower Devonian Littleton Formation. Several workers from Maine have agreed that the unit "Spqr" bears considerable resemblance to a "distal portion of the Perry Mountain". If this scheme of correlation is correct, then the Paxton Granulite Member of the Wachusett subzone would indeed correlate with the Madrid Formation, and the gray schists would belong to the Lower Devonian Littleton Formation and not to the Lower Silurian Rangeley Formation.

Structurally above and to the east of the recumbent anticline just described there are additional belts of Paxton Formation Granulite Member with some zones of sulfidic schist, alternating with belts of gray schist with minor quartzite, tentatively assigned to the Littleton Formation. At quadrangle scale the granulite and sulfidic schist zones can be mapped separately, but not at the compilation scales used for Figures 1 and 5. These eastern belts of alternating biotite granulite with sulfidic schist, and gray schist with minor quartzite become intricately involved with the intrusive sills of the Fitchburg Complex.

The structurally lowest parts of the Fitchburg Complex are a series of granitic to tonalitic sills (grg) that have not been studied petrographically. These are overlain by an extensive unit of biotite tonalite to biotite granodiorite (Dfgd), with rare lenses of hornblende tonalite (Dft) that has been extensively studied by Maczuga (1981) on both the east and west limbs of the Wachusett syncline. Structurally higher is a major sheet of muscovite-biotite granite gneiss (Dfgrg). Eastward and southward the Complex is dominated by much less deformed muscovite-biotite granite (Dfgr), which is characteristic of the classic granite quarries at Rollstone Hill in Fitchburg. A sample from Rollstone Hill has yielded a zircon age of 390 ±15 Ma (Zartman and Naylor, 1984). It is not presently clear whether the weakly developed foliation and lineation in this rock is because it is younger than the more foliated members of the Complex, or because the rock itself was more resistant to deformation.

## INTRUSIVE IGNEOUS ROCKS

Tectono-stratigraphic settings of the intrusive igneous rocks have been discussed above, and some aspects of their structural geology are covered below. The intrusions are the Hardwick Tonalite and related intrusions, the Fitzwilliam Granite, the Coys Hill Granite, the Fitchburg Intrusive Complex, and a host of minor granitic and tonalitic gneissic sills.

The Coys Hill Granite (Stop 7) has not been studied in detail, but it has identical characteristics, though generally more metamorphosed and deformed, to the Kinsman Granite of New Hampshire that has been studied extensively by Clark and Lyons (1986). It is a classic "S-type" granite in the sense that it is extremely peraluminous with primary garnet, cordierite, and sillimanite, indicative of a source from the melting of pelitic sediments. However, its characteristics indicate it is not a result of in situ melting of the enclosing Silurian-Devonian pelites, but instead suggest a deeper source with some interactions with mantle-derived mafic magmas.

The Hardwick Tonalite (Stop 2) and more mafic associated rocks were characterized in detail by Shearer (1983, Shearer and Robinson, 1988). The hornblende and some of the biotite tonalites have "I-type" characteristics including mildly oxidized compositions supposed to be derived by fractional crystallization from magmas produced in a setting of mantle subduction. However, there is a continuous gradation to the muscovite tonalites that are mildly peraluminous and reduced, suggesting an "S-type" or mixed source. Garnet tonalites occur only at contacts and are clearly a product of country rock contamination. Despite the "I-type", subduction-related characteristics, the Hardwick Tonalite and related bodies are unusual in their high K, P and Mn contents (Shearer and Robinson, 1988), suggesting a possible genetic connection to alkali basalts. The microcline-granite porphyry pods within the Hardwick Tonalite (Stop 3) are unusual in their low silica content, tonalitic matrix, and common sillimanite pseudomorphs after andalusite with corundum rims. They may represent interstitial liquids of the tonalite that have been contaminated by included pelitic schist. Shearer (1983) has also characterized other minor binary granite bodies associated with the Hardwick Tonalite, including the post-tectonic Fitzwilliam Granite.

The Fitchburg Complex, characterized petrologically and geochemically by Maczuga (1981), includes three major phases; an extensive lower sill of biotite granodiorite to tonalite, with three small masses of hornblende tonalite, that is very similar overall to the Hardwick Tonalite; a much larger strongly foliated biotite-muscovite granite gneiss that is mildly peraluminous; and a large area of weakly foliated binary granite. The last occurs in the classic quarries at Rollstone Hill in Fitchburg where a U-Pb zircon age of 390 ±15 Ma was obtained (Zartman and Naylor, 1984).

## STRUCTURAL DEVELOPMENT

Structural implications of the stratigraphic relationships in the region have been mentioned above, including the implications of several occurrences of graded bedding. It appears there has been an early phase of west-directed isoclinal fold nappes, followed by thrust nappes, the latter based on the work of P. J. Thompson (1985) and H. N. Berry (1989) to the north and south respectively. For a long distance east of the Bronson Hill anticlinorium the axial surfaces of the fold nappes are overturned to the east. Where the west-dipping foliation passes over the Gardner foliation arch, so that it dips east, it is natural to believe that the recumbent fold axial surfaces are backfolded and overturned beyond recumbency, so that the folds now appear directed from the west. This appears to be true, and furthermore the folds appear to be rooted on the west side of the foliation arch.

The earliest intrusion that can be demonstrated is the Coys Hill Granite, equivalent to the Kinsman Granite in New Hampshire, where P. J. Thompson (1985) has shown that it was intruded and then truncated by the Chesham Pond thrust. The abundant tonalitic intrusions were probably somewhat later. Berry (1989), for example, has mapped an intrusive sheet that cuts across fold and thrust nappes, but was itself involved in peak regional metamorphism. The east-west trending linear shear fabric associated with the backfold stage is implanted in most of the intrusions. Exceptions are the Fitzwilliam Granite of southern New Hampshire, which appears to be post-tectonic and has been tentatively assigned to the Mississippian (P. J. Thompson, 1985), and possibly the eastern granitic phase of the Fitchburg Complex.

The general period of backfolding of the earlier fold and thrust nappes appears to be associated with the development of an east-west trending linear fabric that overprints the peak metamorphic minerals of the stratified rocks as well as nearly all of the intrusive rocks. Where this fabric has been studied in detail (Robinson et al., 1986; Finkelstein, 1987; Berry, 1989), especially by Peterson (1992 and this guidebook) it is invariably associated with west-over-east thrusting. In a few locations, east-west trending folds appear to be associated with this lineation (Tucker, 1977, 1978; Peterson, 1984), as for example isoclinal folds in the Wachusett tonalite southeast of Wachusett Mountain (see Figures 6 and 7). These may be folds of the backfold stage rotated into the transport direction associated with the west-over-east shearing. A more comprehensive study of kinematic indicators in this widespread shear fabric is greatly needed, of the type completed by Peterson (1992), .

The next phase of structural development has been referred to as the "dome stage", because the associated north- to northeast-trending folds and lineations appear related to dome formation in the eastern part of the Bronson Hill anticlinorium. Peterson (1992 and this guidebook) has proved that this lineation is related to right-lateral longitudinal shear in the Conant Brook shear zone along the east margin of the Bronson Hill anticlinorium, but no shear senses have been determined elsewhere within this strain field. The lineation is closely associated with a series of north- to northeast-trending asymmetric minor folds. Beginning on the east side of the Tully dome (Stop 1), these folds have an east-side-up sense of asymmetry (Field, 1975; Tucker, 1977) for a distance of about 15 km, or approximately 90% of the distance across the Ware stratigraphic-tectonic subzone. East of this point to the eastern limits of the area, the same folds show a west-side-up or top-side-east sense of asymmetry, as spectacularly shown by the big fold at Church Rock (Stop 12), and many other folds in the Gardner (Stop 10) and Wachusett subzones. The complex outcrop pattern of the Smalls Falls Formation south of Stop 10, for example, is caused by the interaction of such folds with topography.

The broad foliation arches and depressions in the region appear to be yet younger than the "dome-stage" folds, and could be as young as late Paleozoic. This is proved by the way the dome-stage folds and lineations are deformed across the Gardner arch, plunging southwest on the west limb and northeast on the east limb.

## METAMORPHISM

No detailed studies of metamorphic petrography or petrology have been completed in the region except for the descriptive work of Tucker (1977),Peterson (1984), Mook (1967), and Springston (1990); the study of the coarse garnet-cordierite gneiss at Philipston (Stop 6) by Richardson (1975; Lasaga et al., 1977); the detailed work on the Hardwick Tonalite and surroundings by Shearer (1983, Shearer and Robinson, 1980); and the detailed study of the Smalls Falls sample from Stop 9 by Tracy and Robinson (1988). The schists at Stop 1, and for some distance to the east, contain sillimanite and muscovite. They are assigned to Zone III, lacking both staurolite characteristic of Zone II and orthoclase characteristic of Zone IV. Shearer (1983) found schists adjacent to the Hardwick Tonalite are in Zone III in the northwest, near Athol; and change progressively to Zone VI in the southeast near Ware and Barre. However, muscovite is an important constituent of one member of the Tonalite, and in this he found it persisting into Zones V and even VI. He speculated that this muscovite might be stabilized to higher temperature by its high Mg/(Mg+Fe) ratio. He also found a relict contact aureole adjacent to an augite tonalite sill in Zone V (Shearer and Robinson, 1980). The contact aureole seems to have produced a local gradient in $\mu H_2O$ over a distance of a few meters, that resulted in production of different assemblages during subsequent regional metamorphism, low-pyrope garnet and low-Mg biotite far from the contact, and higher-pyrope, higher-Mg biotite close to the contact, all in a sillimanite-orthoclase assemblage and all apparently at the same temperature. Shearer reported numerous occurrences of sillimanite pseudomorphs after andalusite near contacts and in schist inclusions in the Hardwick Tonalite and related rocks (Stop 4). However, these are also widespread in the region away from igneous contacts, occurring for example in the Smalls Formation at Stop 9, and very widely in the gray mica schist of the Wachusett subzone.

The highest reported metamorphic grade is in the sillimanite-orthoclase-garnet-cordierite gneisses just west of the Coys Hill Granite at Stop 6, which marks the northern tip of the Zone VI granulite facies region of southern Massachusetts. East and north of this the grade falls gradually and at Stops 9 and 10 the rocks contain the assemblage sillimanite-orthoclase-muscovite typical of Zone IV. Most of the Wachusett subzone is characterized by the sillimanite-muscovite assemblage typical of Zone III, although Peterson (1984) found an unusual sillimanite-muscovite-garnet-cordierite-beryl assemblage where it is speculated that the cordierite is stabilized by Be. A xenolith within the Fitchburg Complex has andalusite pseudomorphed by sillimanite, with rims of fine sillimanite and orthoclase, indicating progradation of former muscovite rims on andalusite into Zone IV. The relations east of the Fitchburg complex are beyond the scope of this paper, but the grade decreases sharply over a few hundred meters going downward structurally from the contact of the plutonic rocks.

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## ROAD LOG

The field trip will depart the parking lot north of the U. Mass. Football Stadium at 8:00 A. M. Those wishing to return to the starting point (about 5:30 P.M.) should consolidate vehicles. Those planning to travel to points east and north following the field trip should proceed **direct to stop 1**, where we will gather about 8:30 A. M. The trip will end at the summit of Wachusett Mountain about 4:15 P. M. For those traveling from the stadium, the following notes are offered to the beginning of the road log.

**Mileage**

0.0 Exit U. Mass. Stadium. Turn left (north) on University Drive.

0.2 Turn right (east) on Fearing Street.

0.8 Turn right (southeast) on North Pleasant Street.
1.0 Stop lights. Go straight on Triangle Street.
1.5 T junction and lights. Turn left (east) on Main Street.
2.1 Stop lights. Go straight on Pelham Road which becomes Amherst Road in Pelham.
7.5 Flashing light at T junction with Route 202. Turn left (north) on Route 202.
8.1 Overlook for Quabbin Reservoir and Wachusett Mountain (Stop 15). Proterozoic Pelham Quartzite on left.
19.9 Junction Routes 202 and 122. Stay on 202.
20.8 Another junction, Routes 202 and 122. Again stay on 202.
21.6 Right entrance ramp for Route 2 East. This is also Route 202. BEGIN FORMAL ROAD LOG
0.0 Junction of Routes 2 and 202. Proceed east on Route 2.
2.7 Major trench road cut. It is safest to watch traffic and pull off on broad grass strip on left (north) side. This is firm in all weather and gives a much better view of the outcrop.

**STOP 1. SOUTH END OF TULLY BODY OF MONSON GNEISS CUT BY MESOZOIC NORMAL FAULT** (15 MINUTES) See complete description of this stop under Robinson and Elbert, this guidebook, Stop 8 (Their Figures 8 and 9). An addition point for the purpose of this field trip is that the dome-stage asymmetric folds on the east limb of the Tully dome, with west-dipping axial surfaces, show an east-side-up sense of asymmetry. This sense is characteristic of a broad belt across central Massachusetts (Field, 1975), including everything to be seen on this field trip as far east as stop 8. Beyond that the asymmetry changes to west-side-up or top-side-east, a pattern that continues at least as far east as Mount Wachusett and is typified by exposures to be seen at Stop 10 and at Stop 12, Church Rock. These folds appear to predate the Garner Arch, in that they generally plunge southwest, west of the arch, and northeast, east of the arch.

Continue east on Route 2. Take great care entering traffic from north side even though visibility is good.
4.3 Take exit 17 for Route 32.
4.4 Stop sign end of ramp. Turn left (north) on Route 32.
5.5 Stop sign and junction with Route 2A. Turn left (east) on Route 2A.
6.7 Entrance ramp for Route 2 West. Stay left on Route 2A.
6.9 Philipston Town Line.
7.4 Pull well off on right just before underpass beneath Route 2. Leave flashers on. Carefully cross Route 2A to east side and climb up embankment to large cut (with many trees) on north side of Route 2. Stay off pavement of Route 2.

**STOP 2. HORNBLENDE TONALITE IN THE HARDWICK PLUTON** (20 MINUTES) The dark rock in this exposure is characteristic of the most mafic phase of the Hardwick pluton (Figures 1 and 2) and occupies four moderate-sized areas within the larger area dominated by biotite tonalite (Shearer, 1983, and his Figure 7). The rock is massive to weakly foliated and in this area consists of 40-42% plagioclase An 42-36, 19-22% quartz, 21-30% biotite, 1.7-10% hornblende, 1.3-1.9% sphene, 1.2-2.5% apatite, 0.4-1.2% orthoclase, and 1.9-2.4% opaques including magnetite, ilmenite and pyrite. Modes plot as either tonalite or quartz diorite. Chemical analyses show about 57% silica, with a range of chemistries from 6% normative diopside to 0.3% normative corundum, depending mainly on the amount of hornblende present. All of the biotite tonalites show normative corundum. The rock is much less foliated than the bulk of the biotite tonalite. The rocks in this vicinity are more magnetic than most others in the Hardwick pluton and produce an unusual aeromagnetic anomaly indicating the rocks have a reversed remanent magnetization. Shearer completed detailed electron probe analyses and some wet chemical analyses for FeO of hornblendes and most other minerals. Using Shearer's results Zen (1989) computed an estimated pressure of intrusion of 6 kbar. This is compatible with the estimated minimum pressure of regional metamorphism, not the postulated pressure of contact metamorphism, suggesting the hornblende may be re-equilibrated. Interestingly the scarce crystals of orthoclase contain 3.6 to 4.5 mol% celsian which correlates with a common Ba content of 1000-2000 ppm, about ten times that of surrounding country rocks.

The outcrop contains a network of felsic dikes. Many of these closely resemble the microcline granite porphyry to be seen at Stop 3. At one point a dike contains xenocrysts of sillimanite after andalusite. Shearer reports that such xenocrysts commonly contain corundum on their margins.

Return down embankment to Route 2A, taking special care when crossing back to vehicles. Proceed east on Route 2A.
7.6 Pull off on right next to wooded-over cut that is directly opposite entrance ramp for Route 2 East.

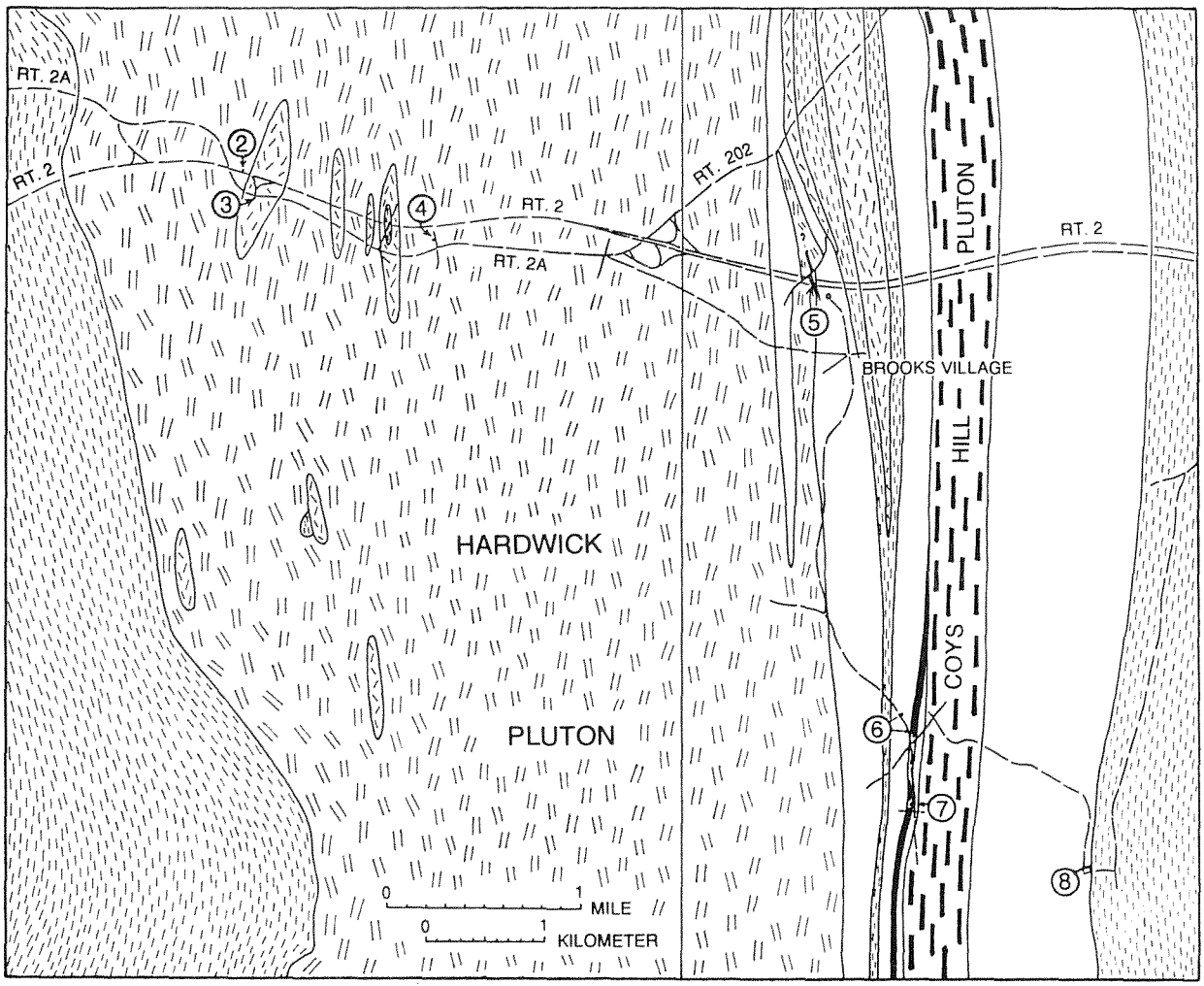



Figure 2. Local geologic map of part of the Hardwick pluton and stratified rocks in the central part of theWare subzone, showing the locations of Stops 2 through 8. Compiled from reconnaissance data of Robinson and Tucker, published mapping by Morton (1985), and unpublished field data of C. K. Shearer.

**STOP 3. MICROCLINE PORPHYRY OF THE HARDWICK PLUTON** (10 MINUTES) This brief stop directly on the road, will acquaint participants with the typical aspect of this rock type. The rock contains 1-20% microcline phenocrysts in a matrix of 34-54% plagioclase An 28-12, 9-31% quartz, trace to 30% microcline, 8-27% biotite, tr-4% muscovite, and accessory sphene, allanite, ilmenite, magnetite, apatite, zircon and pyrite. Xenoliths of normal biotite tonalite are common, as are xenocrysts of sillimanite after andalusite. Of the 7 samples studied modally, 3 class as granites, 3 as quartz monzonites, and 1 as granodiorite. However, minus the phenocrysts, 4 are granodiorites, 1 quartz monzonite, and 1 tonalite. Chemical analyses show these rocks typically contain 64-68% silica, substantially lower than most granites, and 1.3-2.5% normative corundum. Shearer points out that the matrix mineralogy, as well as major and trace elements agree closely with the biotite-muscovite tonalites of the pluton, suggesting a genetic relationship. Further, if the matrix is considered as a liquid composition, it is inappropriate to have crystallized the enclosed microcline, suggesting they may be xenocrysts, not phenocrysts. These observations are similar to those made by Clark and Lyons (1986; see also Lyons, 1988) for the Kinsman Granite, who point out "that the K-feldspars cannot have crystallized from an initially homogeneous melt" and that "geochemically Kinsman compositions can be accounted for on the basis of varying mixtures of leucogranite, assimilated pelite, and restite". The abundant xenocrysts of sillimanite pseudomorphs after andalusite in this Hardwick porphyry strongly favor such a model here also. The corundum rims on the pseudomorphs strongly suggest that the liquids were not silica-saturated until very late in their crystallization history.

Continue east on Route 2A, passing by Highland Avenue.

8.6 Crossroad. Blake Corner Road. Turn left (north) and proceed up hill.
8.7 Dead end. Turn around in truck yard and park. Blake Corner road may be temporarily blocked.

**STOP 4. SCHIST INCLUSIONS IN HARDWICK TONALITE WITH COARSE SILLIMANITE PSEUDOMORPHS AFTER ANDALUSITE.** (10 MINUTES) Small outcrops and/or slightly moved blocks on the west side of the road just south of the truck yard show very coarse sillimanite schist in contact with typical Hardwick biotite tonalite. Sillimanite pseudomorphs after andalusite are abundant throughout central Massachusetts east of the Bronson Hill anticlinorium, but are particularly striking in and near the Hardwick pluton. They are indicative of an early low pressure metamorphism, in part associated with contact metamorphism, that was overprinted by a higher pressure regional metamorphism. In one example described by Springston (1990), at the west contact of the Hardwick pluton, the pseudomorphs have delicate rims of fine euhedral staurolite.

Return south down Blake Corner Road.
8.8 Junction with Route 2A. Turn left (east).
9.7 Right exit toward Philipston Center. Do not take this. Stay on Route 2A for less than 0.1 mile and then take next right on unmarked paved road toward Brooks Village.
11.1 Center of Brooks Village (free-style intersection). Turn sharp left (north) into dead end road.
11.4 Paved turn-around (Figure 2). Park and walk north through roadblock to south side of Route 2. Walk west to outcrops of Stop 5. Say off of pavement on south side of highway. Do not cross to median strip.

**STOP 5. TECTONIC BRECCIA IN TONALITE AT BROOKS VILLAGE** (20 MINUTES) The outcrops at this stop were described in detail by Morton (1985) and the results of his study are paraphrased here. The road cut exposes a 120-meter-thick north-trending tonalite sill that is similar to the biotite-muscovite tonalite of the Hardwick pluton and here is separated from it by approximately 270 meters of country rocks. A breccia zone approximately 5-12 meters thick cuts across this sill. The breccia consists of angular to rounded ellipsoidal fragments of fine-grained gray tonalite and pegmatitic granodiorite in a biotite-muscovite tonalite matrix. Elsewhere in the area, early pegmatites are concordant with early nappe-stage foliation and are folded by the back-fold stage folds. They are interpreted to be equivalent in age to the pegmatitic clasts in the breccia.

The breccia is interpreted to have formed as the result of the brittle behavior of early tonalite and pegmatite dikes while the mica-rich tonalite sill was behaving ductilely at the time of regional mylonitization that occurred late in the backfold-stage of Acadian deformation (see this trip Stop 10 and Peterson, this volume). Clasts that are still attached to one another and clasts with tails provide evidence that the tonalite sill was intruded by dikes that were subsequently brecciated, and they provide evidence that the breccia is not of primary igneous origin as previously proposed (Shearer, 1985, p. 29-33). The matrix tonalite consists of 40% plagioclase An 33, 21% quartz, 3% orthoclase, 32% biotite, 2.5% muscovite, 0.3% sphene, and accessory apatite, zircon, ilmenite, pyrite and calcite. It contains 62.49% silica, 2.62% normative corundum, and is similar in every respect to the normal biotite-muscovite tonalite of the Hardwick pluton.

A later generation of pegmatite dikes cuts across the breccia zone. These dikes are folded into asymmetric minor folds with axial planes parallel to a late foliation that is interpreted to have formed in the dome stage. Many of these show the same east-side-up asymmetry as seen on the east limb of the Tully dome at Stop 1. Because these dikes were not affected by the breccia but were folded during the dome stage, the breccia-forming process definitely predates the dome stage. A later generation of 3-4 cm thick granite sills concordant with the dome-stage foliation are apparently undeformed.

Return south toward Brooks Village.
11.8 Brooks Village again. Go straight through and continue south.
13.2 T junction at millstones. Turn left (southeast).
13.8 Small road cut on left. Park on right.

**STOP 6A. SCHIST OF THE BIG GARNET SYNCLINE** (10 MINUTES ) Medium-sized outcrops of coarse gray schist and gneiss on the left and then on the right of the road for about 0.1 mile (Figure 2). The rock unit is tentatively correlated with the uppermost member of the Upper Silurian Warner Formation in central New Hampshire (J. B. Lyons and N. L. Hatch, personal communication, 1982; Robinson and Goldsmith, 1991) and lies in direct contact with the underlying middle Silurian Francestown Formation (see Stop 7). It was correlated with the Lower Devonian Littleton Formation in the Bedrock Geologic Map of Massachusetts (Zen et al.,

1983). These rocks lie a few hundred feet west of the Coys Hill Granite and the postulated trace of the west directed Chesham Pond thrust as proposed by P. J. Thompson (1985) in the Monadnock area, New Hampshire.

The rocks lie in the sillimanite-orthoclase-garnet-cordierite zone (Zone VI) of the central Massachusetts high and are the northernmost documented exposure of this zone. Two types of coarse garnet-cordierite- sillimanite gneiss are present here. At Stop 6A coarse garnets are set in patches rich in orthoclase and are believed to have grown to large size in the presence of felsic melt. This is very similar to a rock exposed in a pasture in the same belt many miles to the south (Robinson et al. , 1986) where the felsic melt segregations containing garnet are very obvious. At Stop 6B coarse garnets are in a matrix of cordierite, plagioclase, and quartz without orthoclase, and may be a residuum from partial melting. This rock is unusual in containing traces of magnetite. Some years ago Allan Thompson named the rocks at Stops 6A and 6B "pasturite" and "philipstonite" respectively. Fresh samples from the philipstonite were used by Richardson (1975; Lasaga et al., 1977) in a study of retrograde diffusion between garnet and cordierite, but he did not present complete mineral compositions. Mineral compositions from Zone VI generally suggest peak temperatures of 670-740°C at pressures of 6.2-6.4 kbar (See Thomson, Peterson, Berry, and Barreiro, this guidebook). At these outcrops there is abundant evidence of production of sillimanite and biotite by retrograde hydration.

13.9 Move vehicles south to outcrop on right.

**STOP 6B. PHILIPSTONITE** (10 MINUTES) Coarse granular rock with garnets up to walnut size in a matrix of dark blue-gray cordierite.

Continue south.

14.0 Four way junction and stop sign. Cross Route 101 and proceed straight onto narrow gravel road.

14.3 Rusty natural outcrop directly beside road on right. Pull over on right where house and wide driveway is visible directly ahead.

**STOP 7. FRANCESTOWN FORMATION, WARNER FORMATION(?) BIG GARNET GNEISS AND CONTACT OF COYS HILL GRANITE** (25 MINUTES) The rocks to be seen at this stop (Figure 2) are characteristic of a narrow zone west of the Coys Hill Granite that has been traced from the Winchendon area southward through the Ware area, and has been recognized by Peterson (this guidebook) in the Palmer area. These rocks come the closest in detail to the sequence recognized in the Monadnock area, New Hampshire by P. J. Thompson. In that area they all lie west of the Kinsman Granite that is interpreted to be east of the nappe-stage Chesham Pond thrust. Although the Kinsman Granite and the Coys Hill Granite cannot be directly connected in the Monadnock area, their rock type and structural position appear to be identical. For this reason the western contact of the Coys Hill Granite in Massachusetts is postulated as the trace of the Chesham Pond thrust.

The key stratified unit for correlation with New Hampshire is the sulfidic calc-silicate of the Francestown Formation well exposed directly beside the road. On the state bedrock map this was indicated as Fitch Formation (Sfs) for reasons described by Robinson and Goldsmith (1991). It is medium to dark gray when fresh and light gray with a brown crust when weathered. It tends toward slabby weathering. According to Field (1975) the rock typically consists of 60-80% quartz and labradorite-bytownite, 3-35% diopside, 1-6% pyrrhotite, 1-2% graphite that is conspicuous in hand specimens, and 1-2% pink pleochroic sphene. The rock type both to the east and to the west is the big garnet rock already seen and discussed at Stop 6, that is now considered to belong to an upper member of the Upper Silurian Warner Formation. The belt of big garnet rock to the east of the Francestown Formation lies in a highly sheared pegmatitic contact zone with the Coys Hill Granite. The belt to the west is in an isoclinal syncline, west of which is commonly found a second belt of Francestown Formation followed by rusty schist of the Rangeley Formation. It was proved that this belt is truly a syncline and that the big garnet rock is younger than the Francestown, on an NEIGC field trip in 1982 when graded bedding near the Francestown contact was found at a stop on Ragged Hill in the Ware area on both sides of the big garnet rock. A consequence of this interpretation is that the Francestown belt observed at this station, and for a long way south, should be an isoclinal anticline of major amplitude.

After examining the Francestown outcrop walk south to driveway and then directly west about 300 feet through overgrown clearing to knob with superb outcrops of big garnet rock west of the Francestown. Return same way to driveway area and observe a second set of outcrops of big garnet rock in the narrow belt east of the Francestown. Now walk east along south side of small stream to outcrops on the right. At first these are very messy feldspathic

rocks and garnet schists, but at the far end is a superb peeled exposure of strongly foliated Coys Hill Granite. Return by same route to gravel road and return to vehicles.

14.4 Drive south to wide driveway and turn around. Return north.
14.9 Junction with Route 101. Turn right (northeast) on Route 101.
15.0 Four-way junction. Turn right (southeast) on Burntshirt Road. Becomes Stone Bridge Road.
15.6 Cross the Stone Bridge.
16.0 Three-way junction. Turn right (south) on Henshaw Road.
16.5 Park near right hand L turn and walk to outcrop in field to northeast.

**STOP 8. COARSE GRAY GARNET-SILLIMANITE GNEISS WITH GRADED BEDDING AND LARGE CALC-SILICATE PODS** (10 MINUTES) This beautiful outcrop lies at the east edge of a belt of gray weathering schists east of the Coys Hill Granite (Figure 2) that has been traced in reconnaissance entirely across the State of Massachusetts. It includes the rocks at Mount Hitchcock in Wales described by Thomson, Peterson, Berry and Barriero (this guidebook). It was shown as Littleton Formation (Dl) on the state bedrock map, and several such belts were interpreted as isoclinal synclines surrounded by sulfidic schists of the Ordovician Partridge Formation (Ops). This view was encouraged by a number of observations of graded bedding, particularly by Tucker (1977), showing that some of these belts truly are synclinal. A more recent view is that all of these rocks may belong to the Lower Silurian Rangeley Formation. At any rate the grade bedding at this outcrop suggests that the strata are right side up and face away from the sulfide-rich schist that lies just in the trees to the east. The calc-silicate pods are very similar to those mapped in members of the Rangeley Formation in the Monadnock area to the north.

Continue (now east) on Henshaw Road.
16.7 T junction with Barre Road. Turn left (north).
17.0 View of Mt. Monadnock ahead.
17.4 Road cut on right. Sulfidic mica schist with granite pods; probably Rangeley Formation.
17.6 Second road cut in sulfidic schist and granite. Northeast-trending sillimanite lineation.
17.7 Stop sign. Turn right (northeast) on Route 101.
18.0 Center of Templeton (Figure 3). Country Store. Go to stop sign and turn right (east) on Route 2A then immediately bear right (southeast) on Hubbardston Road.
20.2 "Cross Road" enters from left. Go straight on Hubbardston Road.
20.7 Hubbardston Town Line.
20.9 Small tarn on right with road cut beyond. Park on right just before tarn and walk south to road cut.

**STOP 9. CORDIERITE-SILLIMANITE-PYRITE-PYRRHOTITE SCHIST OF SILURIAN SMALLS FALLS FORMATION** (20 MINUTES) This distinctive rock type consists of interbedded sillimanite schists and sillimanite quartzites extremely similar to the type Smalls Falls Formation in northwestern Maine and through much of central New Hampshire (Hatch, Moench and Lyons, 1983). On the state map this was called the Sulfidic Schist and Quartzite Member of the Paxton Formation (Spsq). At the south end of the exposure this physically overlies a gray garnet-mica schist, either Littleton Formation or Rangeley Formation, that will be extensively seen at Stop 10.

The rock is unusual in having extremely high sulfide content, at this locality both pyrrhotite and pyrite, that has led to unusual metamorphic reactions, with the production of extremely Mg-rich silicates and rutile rather than the ilmenite that is characteristic of most schists of the region. Because of the easy weathering of pyrrhotite, this is one of only a few localities where fresh samples have been obtained, and only one of two where rocks contain both pyrite and pyrrhotite and essentially pure Mg end-member cordierite. The locality is T1F of Tracy and Robinson (1988). The assemblage is quartz, plagioclase An37, orthoclase, sillimanite, muscovite, biotite, cordierite, rutile, pyrrhotite, pyrite, graphite. Some of the sillimanite is in pseudomorphs after andalusite. The euhedral pyrite is rimmed by anhedral pyrrhotite consistent with prograde reactions in which Fe is removed from silicates and oxides, and combined with S in pyrite to make pyrrhotite. The ratios of Fe/(Fe+Mg) for biotite and cordierite are 0.053 and 0.002 respectively, and the cordierite contains S reported as 0.93 Wt% $H_2S$. Fluid inclusions contain substantial $H_2S$ which accounts for the smell when fresh samples are broken. Sulfur isotope measurements of both pyrite and pyrrhotite show $\partial^{34}S = -27.3$. Such isotopically light sulfur is typical of bacterially reduced sulfur, as contrasted with values around -12 for normal sulfidic schists, presumed to have formed from protolith shales deposited under conditions where bacterial reduction was much less efficient.. These features suggest anoxic conditions of deposition in this region in the late Early Silurian, either in a restricted basin or more probably in an oceanic anoxic event.

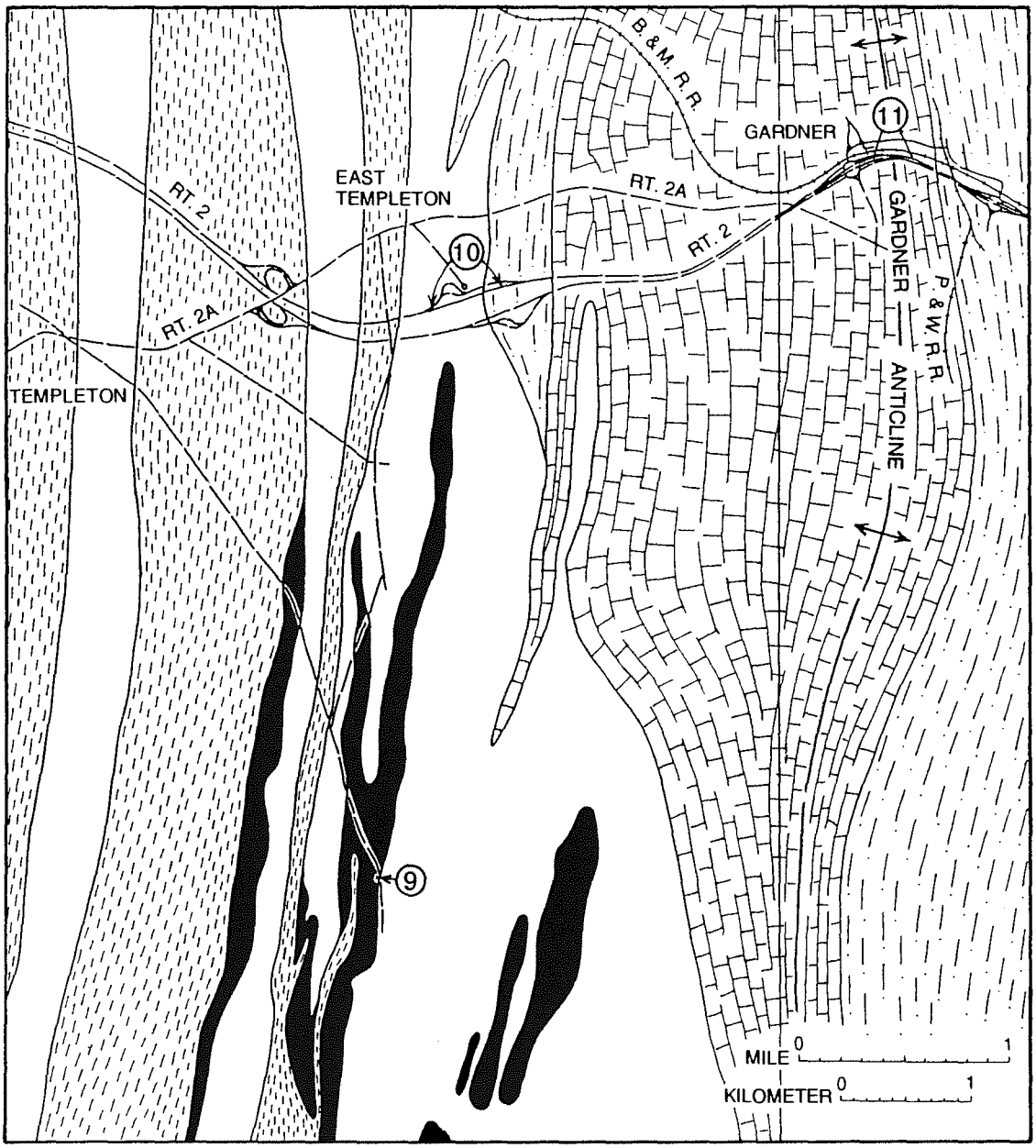



Figure 3. Local geologic map of part of the eastern part of the Ware subzone and western part of the Gardner subzone, showing the locations of Stops 9, 10, and 11. Complied from reconnaissance data of Robinson and Tucker, an unpublished ground magnetic survey by D. W. Klepacki, 1977, and unpublished geologic mapping by Heather B. Stoddart, 1977.

Proceed south to turnaround point.

21.0 Turn around using driveway on left. Return north toward Templeton past Stop 9.
21.8 Turn right (northeast) on Cross Road.
22.5 Junction with South Main Street. Bear left (north).
23.1 Four way junction. Stay straight on South Main Street.
23.8 Overpass over Route 2.
24.3 Junction with Route 2A in East Templeton. Turn right (east) on 2A and proceed for two blocks.
24.5 Turn right (south) on Cottage Lane (dead end).

24.9 Cul de sac (Figure 3) at end of Cottage Lane. Park and walk south around fence into Route 2 rest area parking. Walk west and down hill on left-hand side of west-bound re-entry ramp to west end of huge rock cut. Throughout this stop remain on the grass strip on the **north side** of the highway away from the edge of the pavement. Under no circumstances cross the highway because of heavy high-speed traffic. Please do not proceed as far east as the rest area entrance ramp, because we will exit via the rock west of that

**STOP 10. DEFORMATION FEATURES IN GRAY AND SULFIDIC SCHISTS INTRUDED BY GRANITE AND PEGMATITE** (80 MINUTES INCLUDING LUNCH) The rocks to be seen at this stop (Figures 3 and 4) include gray schists that may be assigned to the Littleton or the Rangeley Formation and sulfidic schists with calc-silicates that are assigned to the Paxton Formation. They lie on the west limb of the Gardner foliation arch, which will be observed at Stop 11. The structural and metamorphic features are typical of the Gardner stratigraphic-tectonic subzone in northern Massachusetts.

The gray schist at this stop lies within a narrow belt that could be interpreted as a recumbent syncline. In contact with it structurally above is the Smalls Falls Formation (White Schist Member of the Paxton Formation, Spsq, on the state map). The contact between this unit and the Smalls Falls was seen at Stop 9. In contact with the gray schist, structurally below and to the east is the Sulfidic Schist Member of the Paxton Formation, consisting predominantly of pyrrhotite-biotite-sillimanite schist with lenses of well bedded feldspathic calc-silicate granulite. This schist contains a dismembered metamorphosed gabbro dike at the extreme east end of these cuts. The Sulfidic Schist Member, which pinches out a short distance to the south, overlies the Granulite Member of the Paxton (Sp) that occupies the center of the Gardner arch as seen at Stop 11.

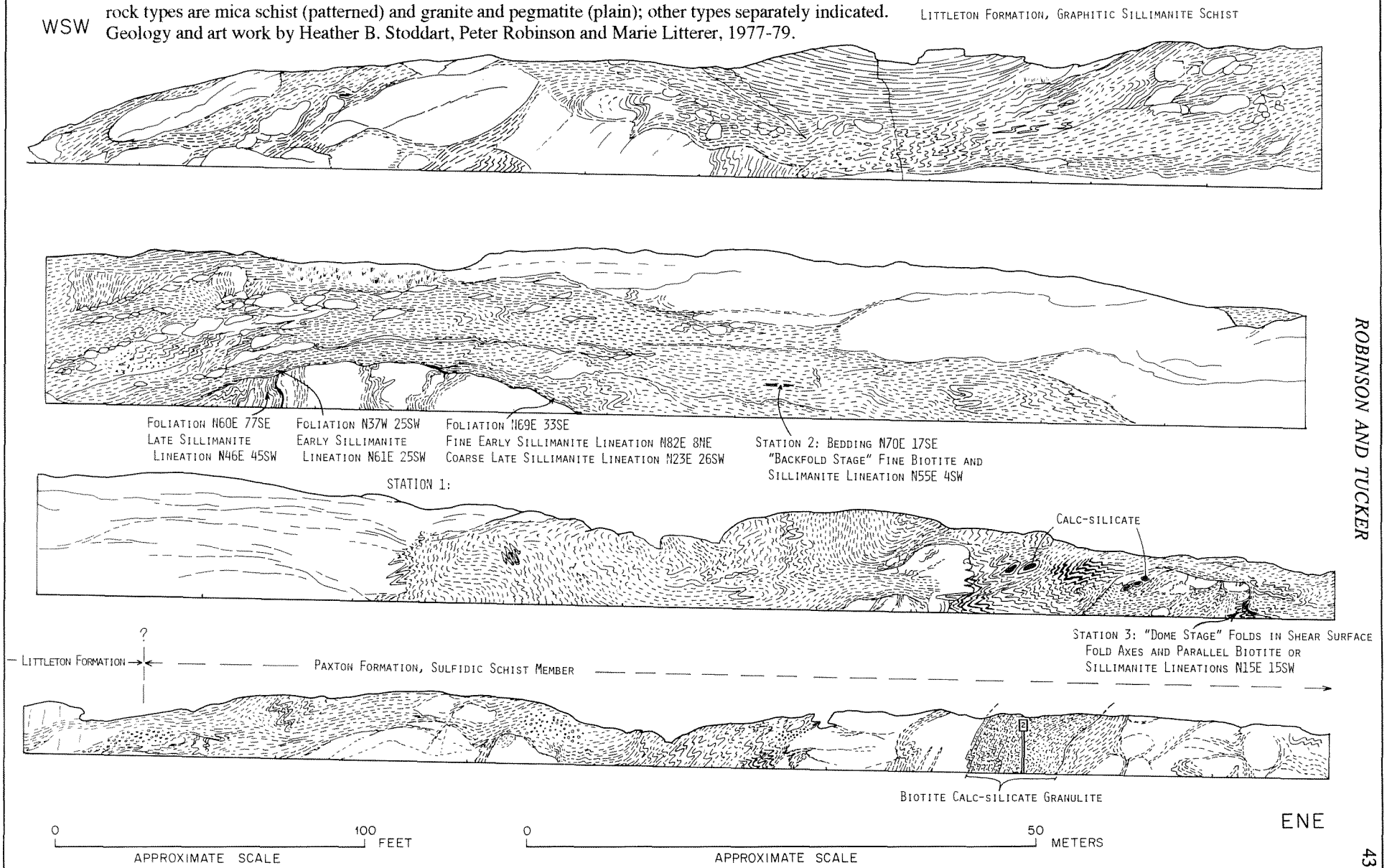
The gray schist at this stop consists predominantly of quartz-plagioclase-biotite-muscovite- K feldspar-garnet schist, typical of the zone in which muscovite breakdown is in progress but not yet complete (Tracy , 1975). In this location we are on the northeast flank of the central Massachusetts metamorphic high. Profuse feldspathic streaks suggest partial melting was also in progress. Beds and large pods of calc-silicate rock, a few with preserved carbonate cores, are abundant. Pegmatite and fine-grained garnet granite are profuse. Some of the granite has an equigranular texture suggesting it is late, but more careful examination of contact relations proves that is was solidified before genesis of the two different sillimanite lineations associated with the two most obvious structural episodes.

The apparent structural and plutonic chaos at this stop may help to explain the extreme scarcity of stratigraphic and structural research in north-central Massachusetts in the past! These rocks must already have been highly deformed and metamorphosed before intrusion of the granites, although no minor folds of the early episodes can be definitely recognized. The granites and pegmatites with their schist inclusions are severely deformed, boudinaged, and bounded by zones of strongly foliated schist of mylonitic aspect. A prime example of such a boundary is at Station 1 (near large Templeton sign) where strongly foliated sillimanite schist clearly truncates the internal structure of a large granite block. This bounding schist contains fine biotite and sillimanite lineation trending N50-80E that is characteristic of the **backfold** and/or mylonitic **stage** in the region. Superimposed on this throughout the outcrop is a southwest-plunging set of lineations clearly parallel to and related to the more obvious eastward overturned folds in the outcrop and definitely related to the Acadian **dome-stage** deformation in the region. The pattern of these later southwest-plunging, eastward-overturned folds is reflected on a larger scale by the outcrop pattern of unit contacts to the south and southwest (Stoddart, unpublished data, 1977; Klepacki, unpublished data, 1977). The early E-W lineation has also been measured at Station 2. An excellent example of a shear zone involved in a later overturned fold can be examined at Station 3. In folds in the thick calc-silicate rocks at the east end of the traverse, both lineations may be seen together on the same bedding surface.

From the thick calc-silicate rocks climb over outcrop and past end of fence onto entrance ramp for rest area. Walk up ramp (west) back to rest area parking and go around road block to vehicles. Return north on Cottage Lane.

25.3 Junction with Route 2A. Turn left (west). Leave East Templeton and cross another overpass of Route 2.
26.1 Make sharp left turn into entrance ramp for Route 2 East.
26.5 Enter Route 2 East.
29.0 Take Exit 22 for Route 68.
29.4 Around rotary and exit for Route 68 North. Stay right beyond railroad underpass.
29.5 Take sharp right turn beyond unmarked red building beyond Speedee Oil Change.
29.6 Proceed to end of factory driveway beside railroad tracks (Figure 3). Walk directly south across tracks to outcrop on south side of Route 2 west exit ramp. Watch for traffic.

Figure 4. Detailed drawing based on photographs of part of the East Templeton cuts on Route 2 at Stop 10. Shown is west end of the cut on north side of west-bound lane, near west-bound rest area. Dominant rock types are mica schist (patterned) and granite and pegmatite (plain); other types separately indicated. Geology and art work by Heather B. Stoddart, Peter Robinson and Marie Litterer, 1977-79.



**STOP 11. PAXTON FORMATION IN CENTER OF GARNER ARCH** (20 MINUTES) Two closely spaced outcrops at this stop lie on opposite limbs of the Gardner arch, which separates west-dipping strata from here west to the Bronson Hill anticlinorium (Stops 1-10) from east-dipping strata from here east to the Fitchburg plutons (Stops 12-15). Both outcrops are typical biotite-feldspar granulite with minor calc-silicate of the Paxton, with significant layers of sheared granite and pegmatite. The Gardner arch is believed to be younger than the minor folds and lineations associated with the Acadian dome stage, because these structural features trend more to the northeast and are deformed across the arch (see Peterson, this guidebook) The fate of the Gardner arch, the Oakham anticline and related open folds as one proceeds southward to the Connecticut border is poorly understood.

The first outcrop, on the south side of the west-bound exit ramp, shows west-dipping foliation on the west limb of the arch. After observing this, return to railroad and walk east along south side of railroad to rock cut and to outcrop on north side of Route 2. The outcrop on Route 2 is cleaner. Both show consistent gentle east dips on the east limb of the arch. Return west along south side of railroad and then cross active track back to vehicles.

Exit from factory lot. Turn right (east).
30.0 Stop sign. Turn right (southeast).
30.2 Stop light. Junction with Route 140. Go straight on Route 140 South.
30.4 Underpass beneath railroad. Enter rotary.
30.6 Enter ramp for Route 2 East.
30.7 Enter Route 2 East. Big cut on right in sulfidic mica schist (Spss). East-dipping foliation.
32.6 Low cut in sulfidic schist and granite. Cross drainage divide into Merrimack River watershed.
33.8 "Digital" outcrop on left. Gently east-dipping gray granulite with calc-silicate and sulfidic schist of Paxton Formation. Shows E-W lineation and some isoclinal folds, especially in the median strip. Dangerous for a group. View of Wachusett Mountain on right. Descend long dip slope.
35.1 Take Exit 25, Route 140 South (Figure 5).
35.3 Stop sign, four way junction. Turn left on Route 140 South. Outcrop near lights shows sulfidic schist of Paxton Formation dipping gently northeast beneath gray schist of Littleton Formation. Gentle dip and consistently oriented foliation are characteristic of the Wachusett Mountain zone.
36.0 Turn sharp right (northwest) onto Worcester Road.
36.4 Big ledge of gray schist (Littleton Formation ?) on right. Turn left at junction onto West Princeton Road.
36.9 Fork in road. Take left on West Princeton Road.
37.8 Low outcrop of gray schist on low rise on right. Park and follow flagged trail through laurel and minor poison ivy for about ten minutes to Church Rock. Use same trail on return.

**STOP 12. DOME-STAGE RECUMBENT FOLD IN PAXTON FORMATION AT CHURCH ROCK WITH OLDER REFOLDED E-W LINEATION** (45 MINUTES) This large outcrop was first examined by Tucker and Robinson in 1976, but it was so heavily bushed that much of it could not be seen. In 1980 Robinson, Peter J. Thompson, and Stuart Michener cut a trail to the outcrop, cleared it of trees and brush, mapped it at 10 feet = 1 inch using a specially built wooden frame system, and made structural measurements assisted by a 25-foot aluminum ladder. The results are portrayed in Figure 6. The rock is typical Paxton Formation biotite granulite with minor calc-silicate and with numerous thin, deformed pods of granite and pegmatite. The folds are dome-stage folds in foliation that plunge gently northeast and show transport toward the east. This sense is typical of a huge range of such folds over a province that covers much of east-central Massachusetts with an eastern limit that is not well known. Locally there is a late foliation parallel to fold axial surfaces. Several foliation surfaces parallel to bedding show an earlier E-W trending lineation that is deformed by the folds, but not enough measurements could be made on steeply dipping limbs to determine a transport direction in detail.

Continue south on West Princeton Road.
38.4 Three way junction. Stay right on West Princeton Road.
38.9 Right at fork on Lanes Road. Cross divide back to Connecticut River watershed.
39.6 Left at three-way junction. Stay on Lanes Road.
39.7 Low outcrops and overhanging ledge on left.

**STOP 13. PAXTON QUARTZITE AND RUSTY SCHIST (Spqr) OVERLYING PAXTON GRANULITE (Sp)** (15 MINUTES) The unit Spqr is dominated by 5-10 cm beds of impure rusty-weathering quartzite alternating with pyrrhotite mica schist. It is not known to contain the extremely magnesian assemblages found in the Smalls Falls Formation to the west. The quartzite appears to grade downward over a thickness of 5 to

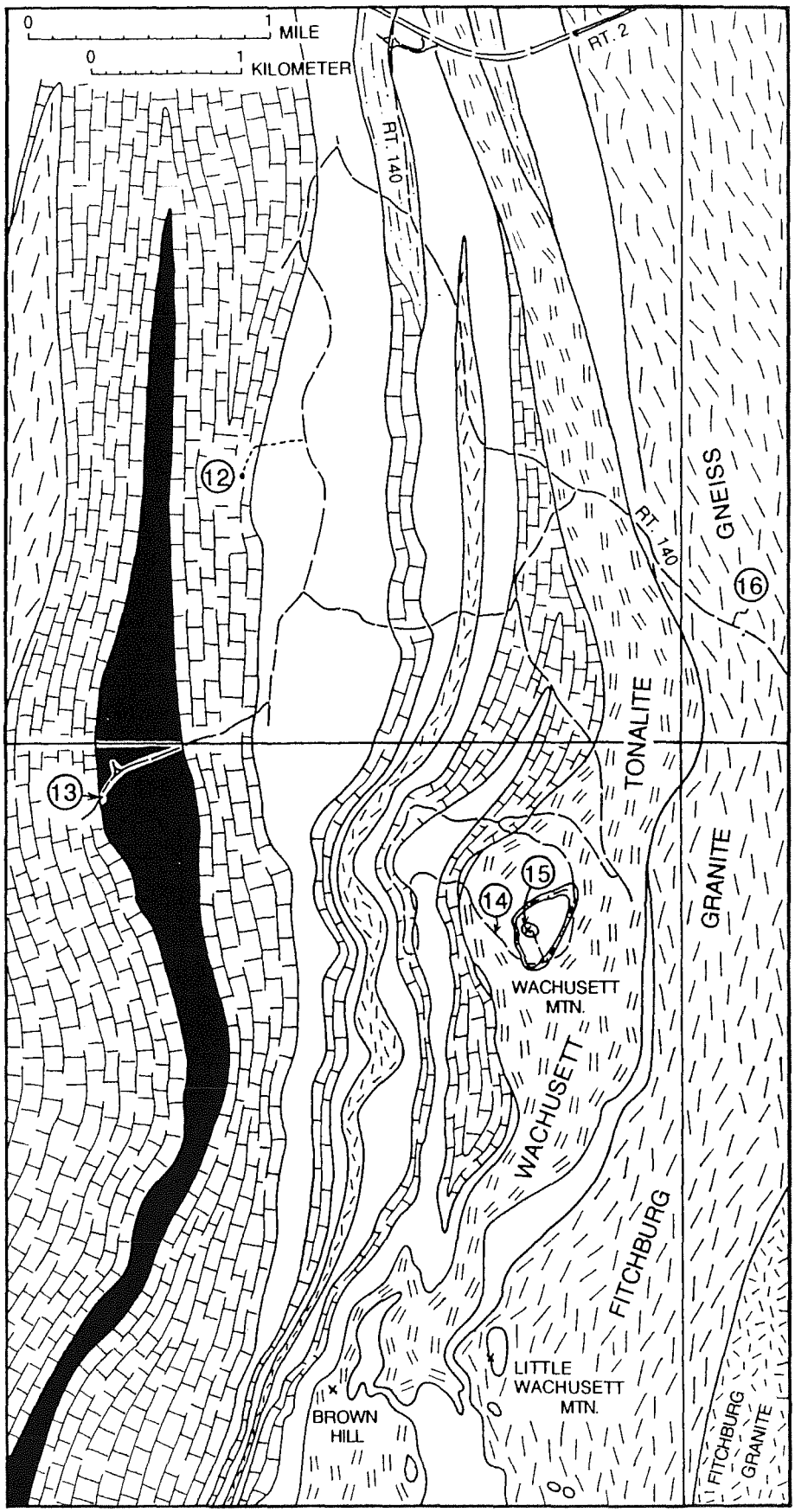
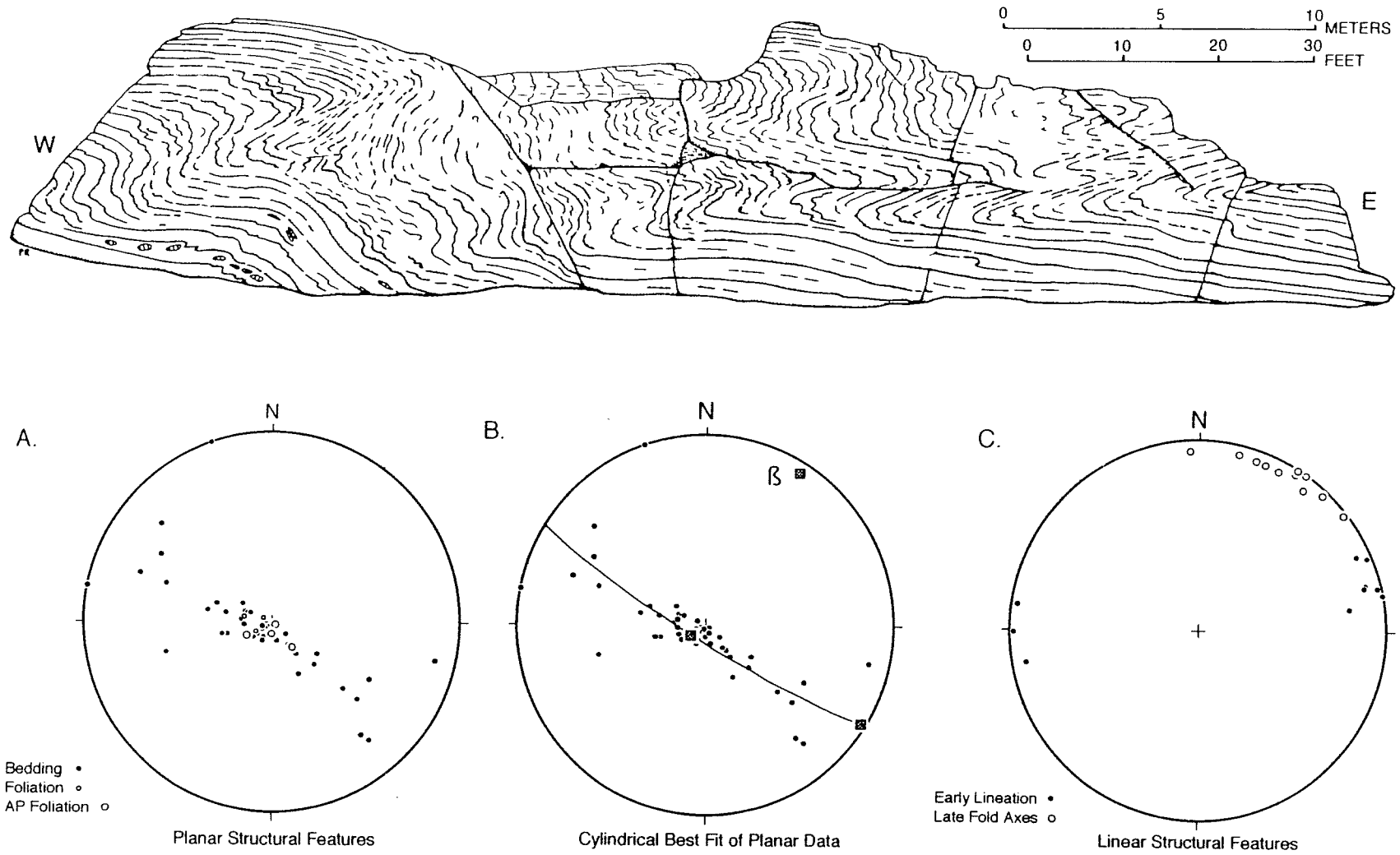



Figure 5. Local geologic map of part of the Wachusett subzone showing locations for stops 12, 13, 14, and 15. Complied from reconnaissance and detailed mapping by Tucker and Robinson, 1976, and detailed descriptions of igneous rocks by Maczuga (1981).

Figure 6. Detailed field drawing of the outcrop of Paxton Formation Granulite Member at Church Rock. Prepared October 1980 by Peter Robinson assisted by Peter J. Thompson and Stuart R. Michener. Contacts were located using a 5 x 30 foot wooden grid positoned repeatedly against the vertical outcrop. Insets A, B, C show equal area diagrams of planar and linear features measured with the assistance of a ladder.



15 meters into predominantly sulfidic schist, just above the Granulite Member (Sp) contact. This section is interpreted to be on the inverted limb of a major east-directed recumbent anticline, in which the Quartzite and Rusty Schist Member occupies the core.

Continue on Lanes Road.
39.8 Turn around at farmhouse on left. Retrace route on Lanes Road past Stop 13.
40.1 Junction. Bear right on Lanes Road.
40.8 Junction. Bear left toward Westminster. Recross into Merrimack watershed.
41.3 Sharp right on Bolton Road.
42.3 Junction with Mountain Road near Wachusett Ski Area. Turn right (south).
43.5 Entrance to Wachusett Mountain State Park. Sharp right turn.
43.6 Talus blocks on left composed of gneissic Wachusett Tonalite.
43.9 Outcrops left show east-dipping base of Wachusett Tonalite sill overlying sulfidic schist of Paxton Formation (Spss) on north face of mountain. For next two miles follow tortuous route on west and southwest sides of mountain eventually approaching main peak from the west (Figure 5). During this tour we will cross once more in and out of the Connecticut watershed which runs about 0.5 mile west of the main peak, but which crosses over Little Wachusett Mountain.
45.4 Cuts in intersection on left. Sulfidic schist of Paxton Formation.
45.5 Low outcrops by tarn, both sides. Gray schists of Littleton Formation in isoclinal syncline.
45.6 Low cuts on right during steep ascent. Sulfidic schist (Spss) and biotite granulite (Sp) of Paxton.
45.7 Basal contact, left, of Wachusett Tonalite sill against gray schist.
46.0 Parking area in quarry on left.

**STOP 14. BIOTITE TONALITE OF FITCHBURG PLUTONS HIGH ON MOUNT WACHUSETT** (15 MINUTES) The purpose of this stop is to note the character of the Wachusett Tonalite, to observe the pervasive E-W mineral lineation believed to be characteristic of the backfold stage in this region, and to look at the view to the west and south.

The Fitchburg plutons (Tucker, 1978; Maczuga, 1981) are a series of rather gently dipping sheets intimately involved with a series of tight isoclinal folds in Silurian Paxton Formation and Devonian Littleton Formation. The isoclinal folds are now east-directed (Figure 7). They may originally have been of the nappe stage, and possibly originally west-directed. The sills were certainly intruded before the (next) backfold stage, if not earlier, because they contain the characteristic backfold stage lineation and because the Wachusett Mountain sill is isoclinally folded by a fold of this stage. Northeast-trending dome-stage folds are also present in this region, as for example at Church Rock, but rock fabrics tend to be dominated by the backfold stage.

The rocks of the Fitchburg plutons are exposed in a broad north-trending syncline. The Wachusett Tonalite is the westernmost and lowest major intrusive unit. A similar granodiorite sill is exposed at South Monoosnoc Hill on the east limb of the syncline (Maczuga, 1981). The next major unit is muscovite-biotite granite gneiss. This is cut further east in the complex by more massive muscovite-biotite granite with uncertain tectonic relations. The latter has yielded a U-Pb zircon age of 390±15 Ma (Zartman and Naylor, 1984).

Proceed on road (up section) into schist septum above sill.
46.2 Intersection. Turn left toward summit.
46.4 Summit. Park near high point or to south if space is available.

**STOP 15. GRAY SCHIST AND FOLIATED GRANITE AT SUMMIT OF MOUNT WACHUSETT (2006')** (25 MINUTES) The actual summit of the mountain is interpreted as a small outlier of the base of the overlying sill (Dfgrg) of foliated biotite-muscovite granite-granodiorite gneiss. This unit was described petrographically and major- and trace-element analyses were presented by Maczuga (1981) based on two samples in cuts on Mountain Road to the southeast and five from the cliffs of Crow Hills to the northeast. The major mineral content ranges from 22-44% plagioclase An 23-29, 6-32% potassium feldspar, 23-28% quartz, 6-20% biotite, and 3-9% muscovite with accessory apatite, zircon, allanite, sphene, clinozoisite and ilmenite. Silica content ranges from 63-71% and $K_2O$ from 2.7-5.4%. Five of the samples classify as granite, two as granodiorite Note the pervasive E-W fabric in the granite and also the late cross-cutting tourmaline veins. Descend a few feet southwest to outcrops of the underlying Littleton Formation. These are dominated by tight, gently plunging, northeast-trending folds, probably of the dome stage, and are cut by pegmatite dikes. There are local obscure examples of the older E-W fabric, generally better preserved in the intrusive rocks.

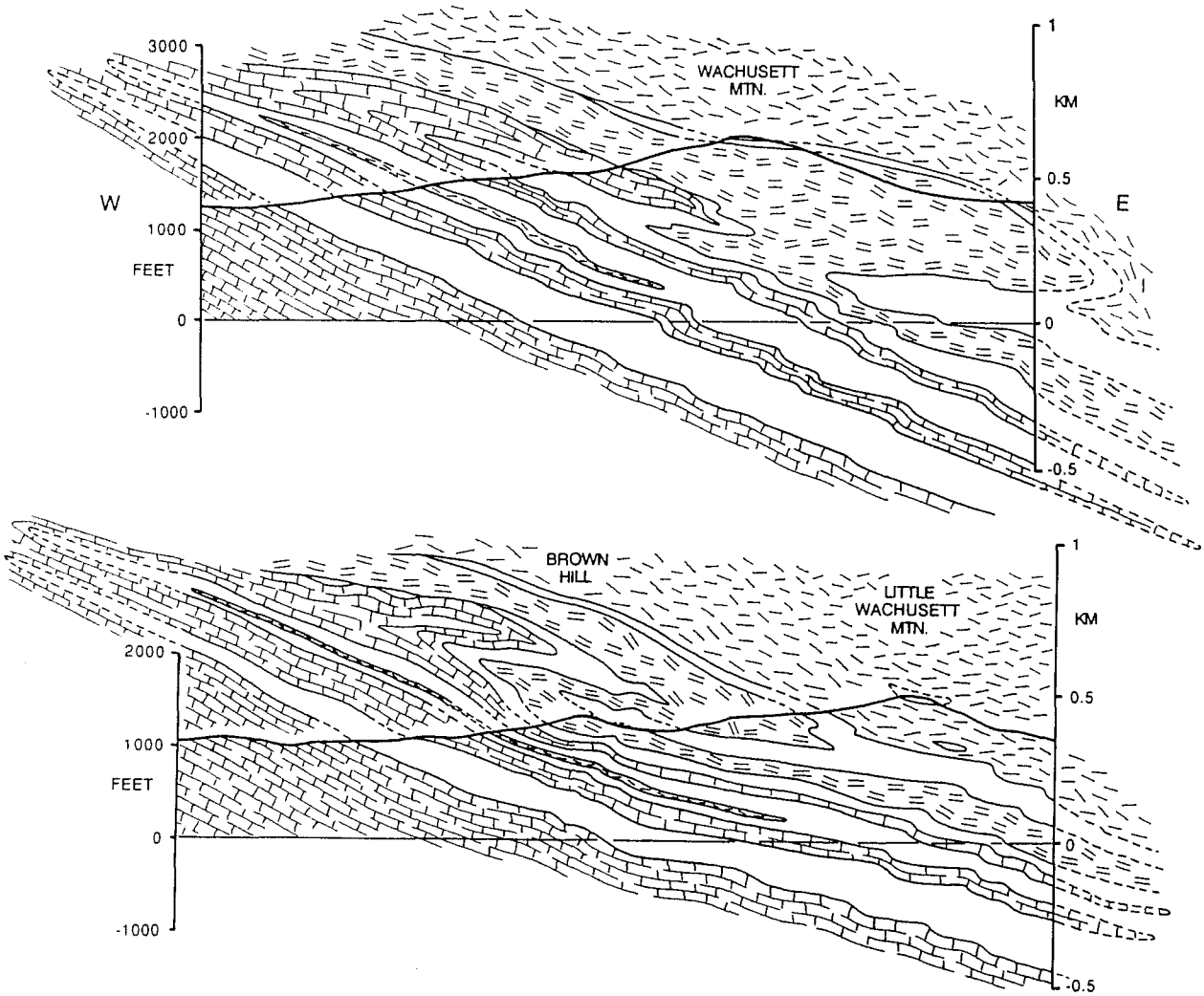



Figure 7. Detailed cross sections through Wachusett Mountain and through Brown Hill and Little Wachusett Mountain (Figure 6), showing relations between stratified rocks and lower parts of the Fitchburg Complex. Prepared for a paper presented in Boston by Tucker (1978).

On a clear day the view includes the Prudential Center and other buildings in Boston, firmly rooted in Avalon, and Mount Greylock in the Taconics in western Massachusetts, overlying North American Grenville basement. In the Swiss Alps this would be equivalent in distance (110 miles or 178 km) to a point at which you could simultaneously view Como in Italy at the north edge of the Po Plain and Zurich in northern Switzerland in the middle of the Swiss Plain! Such observations might justify the statement that the Massachusetts Appalachians are similar in many ways to the Swiss Alps but on a grander scale, being subordinate only in the vertical scale due to the ravages of time. Numerous hills and mountains in Connecticut, Rhode Island, New Hampshire and Vermont are also visible.

END OF TRIP: For those returning to Route 2, turn left (north) at park entrance and drive north on Mountain Road to junction with Route 140. Turn left (northwest) on 140 and follow back to Route 2. For those wishing to see a large exposure of the foliated granite gneiss of the Fitchburg plutons with gray schist xenoliths, follow same instructions but turn right (southeast) on Route 140 to locality indicated as Stop 16 on Figure 5. Park on right (south) at Redemption Rock and follow trail north of Route 140 to cliffs. For those proceeding south or southwest, turn right (south) at park entrance and follow Mountain Road to village of Princeton. Here take Route 62 east to Route 140 and 140 southeast to interchange on Interstate 190, which leads south to Interstate 290 in Worcester and connections to Mass. Turnpike and other Interstate Routes.

## REFERENCES CITED

Berry, H.N., IV, 1989, A new stratigraphic and structural interpretation of granulite-facies metamorphic rocks, Brimfield-Sturbridge area, Massachusetts and Connecticut: Unpublished Ph.D. Thesis, Department of Geology and Geography, University of Massachusetts, Amherst, 330 p., 4 black and white plates, 2 color plates (bulletin in preparation).

Clark, R. G. and Lyons, J. B., 1986, Petrogenesis of the Kinsman intrusive suite: Peraluminous granitoids of western New Hampshire: Journal of Petrology, v. 27, p. 1365-1393.

Emerson, B.K., 1917, Geology of Massachusetts and Rhode Island: U.S. Geological Survey Bulletin 597, 289 p.

Duke, E. F., 1978, Petrology of the Spaulding Group Tonalites, Penacook Quadrangle, New Hampshire: M. A. thesis, Dartmouth College, Hanover, N. H., 117p.

Field, M.T., 1975, Bedrock geology of the Ware area, central Massachusetts: Contribution No. 22 (Ph.D. Thesis), Department of Geology and Geography, University of Massachusetts, Amherst, 186 p.

Finkelstein, D. B., 1987, Sense of shear in a major mid-Acadian mylonite zone in a granulite-facies metamorphosed tonalite and Rangeley Formation of the Merrimack Belt, northeastern Connecticut: Geological Society of America Abstracts with Programs, v. 19, p. 14.

Hatch, N.L., Jr., Moench, R.H., and Lyons, J.B., 1983, Silurian - Lower Devonian stratigraphy of eastern and south-central New Hampshire: Extensions from western Maine: American Journal of Science, v. 283, p. 739-761.

Hatch, N. L., Jr., Robinson, Peter, and Stanley, R. S., 1988, Stratigraphy of the Connecticut Valley belt: **In** Hatch, Norman L. Jr., Editor, The Bedrock Geology of Massachusetts, U.S. Geological Survey Professional Paper 1366-B, p. B1-B34.

Lasaga, A. C., Richardson, S. M., and Holland, H. C., 1977, Mathematics of cation diffusion and exchange between silicate minerals during retrograde metamorphism: In Saxena, S. K., and Bhattacharj, S., Editors, Energetics of Geological Processes, Springer-Verlag, Berlin, p. 353-388.

Ludman, Allan, and Griffin, J. R., 1974, Stratigraphy and structure of central Maine: **In** Guidebook for field trips in east-central and north-central Maine, New England Intercollegiate Geological Conference, 66th Annual Meeting, Orono, Maine, p.154-179.

Lyons, J. B., 1988, Geology of the Penacook and Mount Kearsarge Quadrangles, New Hampshire: **In** Bothner, W.A., Editor, Guidebook, New England Intercollegiate Geological Conference, 80th Annual Meeting, Keene, N. H., p. 60-69

Maczuga, D. E., 1981, Petrology and geochemistry of the Devonian Fitchburg plutonic complex, central Massachusetts: Contribution No. 36 (M.S. Thesis), Department of Geology and Geography, University of Massachusetts, Amherst, 127 p.

Mook, A. L., 1967, Petrography of the Athol Quadrangle, Massachusetts: U. S. Geological Survey open file report.

Morton, P. S., 1985, Tectonic breccia of metamorphosed intrusive igneous rocks in an Acadian shear zone, Brooks Village, north-central Massachusetts: Contrib. No. 54 (M. S. thesis), Dept. of Geology and Geography, University of Massachusetts, Amherst, 43 p., 1 plate.

Osberg, P.H., 1980, Stratigraphic and structural relations in the turbidite sequence of south-central Maine: **In** Roy, D.C., and Naylor, R.S., Editors, A guidebook to the geology of northeastern Maine and neighboring New Brunswick. New England Intercollegiate Geological Conference, 72nd Annual Meeting, University of Maine, Presque Isle, p. 278-289.

Peterson, V.L., 1984, Structure and stratigraphy of the bedrock geology in the Ashburnham-Ashby area, north-central Massachusetts: Contribution No. 47 (M. S. Thesis), Department of Geology and Geography, University of Massachusetts, Amherst, 182 p., 6 plates.

Peterson, V. L., 1992, Structure, petrology, and tectonic implications of highly strained rocks along the west margin of the Acadian granulite-facies high, central Massachusetts: Ph. D. Thesis, Department of Geology and Geography, University of Massachusetts, Amherst, 283 p., 3 plates printed (bulletin in preparation will cover mainly local aspects).

Richardson, S. M., 1978, Fe-Mg exchange among garnet, cordierite and biotite during retrograde metamorphism: Ph. D. thesis, Harvard University, Cambridge, Massachusetts.

Robinson, Peter ,1979, Bronson Hill anticlinorium and Merrimack synclinorium in central Massachusetts: In Skehan, J.W., S.J., and Osberg, P.H., Editors., The Caledonides in the U.S.A.: geological excursions in the northeast Appalachians. IGCP Project 27, Caledonide Orogen, Weston Observatory, Weston, Massachusetts, p. 126-150.

Robinson, Peter, and Goldsmith, Richard, 1991, Stratigraphy of the Merrimack belt, central Massachusetts: **In** Hatch, N. L. Jr., Editor, The Bedrock Geology of Massachusetts. U.S. Geological Survey Professional Paper 1366-E-J, p.G1-G37.

Robinson, Peter, Field, M.T. and Tucker, R.D., 1982, Stratigraphy and structure of the Ware-Barre area, central Massachusetts: Guidebook, New England Intercollegiate Geological Conference, 74th Annual Meeting, Storrs, Connecticut, p. 341-373.

Robinson, Peter, Huntington, J.C., McEnroe, S.A., and Springston, G.E., 1988, Root zone of the Bernardston nappe and the Brennan Hill thrust involuted by backfolds and gneiss domes in the Mount Grace area, north-central Massachusetts: **In** Bothner, W.A., Editor, Guidebook, New England Intercollegiate Geological Conference, 80th Annual Meeting, Keene, N. H., p. 293-334.

Robinson, Peter, Tracy, Robert J., Hollocher, Kurt, T, Schumacher, John C., and Berry, Henry, N., IV, 1986, The central Massachusetts metamorphic high: **In** Robinson, Peter and Elbert, David C., Editors, Field Trip Guidebook: Regional metamorphism and metamorphic phase relations in northwestern and central New England, Contribution No. 59, Department of Geology and Geography, University of Massachusetts, Amherst, p. 195-288.

Shearer, C. K., 1983, Petrography, mineral chemistry, and geochemistry of the Hardwick Tonalite and associated igneous rocks, central Massachusetts: Contribution. No. 51 (Ph.D. Thesis), Department of Geology and Geography, University of Massachusetts, Amherst, 210 p.

Shearer, C.K. and Robinson, Peter, 1980, Constant temperature metamorphism of an $H_2O$ activity gradient in a pre-existing contact aureole, Petersham, Massachusetts: EOS, 61, p. 389.

Shearer, C. K., and Robinson, Peter, 1988, Petrogenesis of metaluminous and peraluminous tonalites within the Merrimack synclinorium: Hardwick Tonalite, central Massachusetts. American Journal of Science, v. 288-A, p.148-196.

Springston, George E., 1990, Stratigraphy and structural geology of the Royalston-Richmond area, Massachusetts-New Hampshire: Unpublished M.S. Thesis, Department of Geology and Geography, University of Massachusetts, Amherst, 132p., 3 plates printed (bulletin in preparation).

Thompson, P. J., 1985, Stratigraphy, structure, and metamorphism in the Monadnock quadrangle, New Hampshire: Contribution No. 58 (Ph.D. Thesis), Department of Geology and Geography, University of Massachusetts, Amherst, 255 p., 6 black and white plates, 2 color plates.

Tracy, R.J., 1975, High grade metamorphic reactions and partial melting in pelitic schist, Quabbin Reservoir area, Massachusetts: Contribution No. 20 (Ph.D. Thesis), Department of Geology and Geography, University of Massachusetts, Amherst, 127 p.

Tracy, R.J., and Robinson, Peter, 1988, Silicate-sulfide-oxide-fluid reactions in granulite-grade pelitic rocks, central Massachusetts: American Journal of Science, v. 288-A, p. 45-74.

Tucker, R. D., 1977, Bedrock geology of the Barre area, central Massachusetts: Contribution No. 30 (M.S. Thesis), Department of Geology and Geography, University of Massachusetts, Amherst, 132 p., 6 plates.

Tucker, R. D., 1978, Contact relations and deformation of the west margin of the "Fitchburg pluton", Wachusett Mountain area, central Massachusetts: Geological Society of America Abstracts with Programs, v. 10, p. 89-90.

Zartman, R.E. and Naylor, R. S., 1984, Structural implications of some radiometric ages of igneous rocks in southeastern New England: Geological Society of America Bulletin, v. 95, p. 522-539.

Zen, E-an, 1989, Tectonostratigraphic terranes in the Northern Appalachians: Their distribution, origin, and age; evidence for their existence: Field Trip Guidebook T 359, 28th International Geological Congress, American Geophysical Union, Washington, D. C., 68p.

Zen, E-an, Goldsmith, Richard, Ratcliffe, N.L., Robinson, Peter, and Stanley R.S., 1983, Bedrock geological map of Massachusetts, Scale 1/250,000 (two sheets in color, with cross sections, tectonic map, metamorphic map, explanation, black and white reference sheet): U.S. Geological Survey, Washington, D.C.

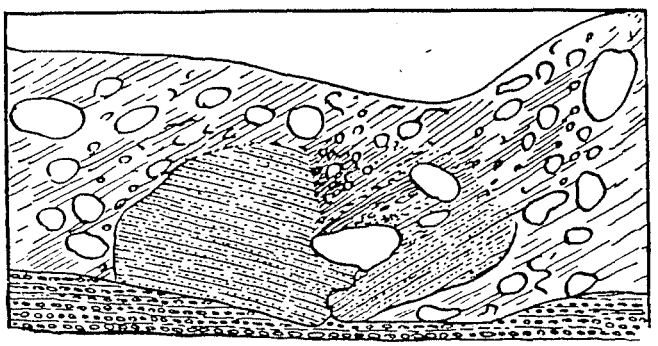


FIG. 34.—Sand bowlders crushed by the ice while frozen, from just south of the telegraph pole seen at the left in fig. 33.

# INVERTED METAMORPHISM, P-T PATHS AND COOLING HISTORY OF WEST-CENTRAL NEW HAMPSHIRE: IMPLICATIONS FOR THE TECTONIC EVOLUTION OF CENTRAL NEW ENGLAND

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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). Thompson et al. (1968) described the occurrence as a "metamorphic overhang" because high-grade sillimanite + K-feldspar bearing rocks on Fall Mountain, N.H. were observed to overlie structurally lower grade rocks. This 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. Subsequent mapping around Fall Mountain (Thompson and Rosenfeld, 1979; Allen, 1984, 1985) and in the root zone in the Merrimack synclinorium (Chamberlain, 1984; P. J. Thompson, 1984, 1985) has confirmed the existence of the Fall Mountain nappe based on stratigraphic and structural considerations.

Metamorphic studies of the Fall Mountain nappe in both the root zone and the structural outlier at Fall Mountain indicate that metamorphism in this area is the result of a complex interplay among pre-orogenic thermal setting, thermal perturbation owing to tectonic transport (thrusting and folding) and subsequent thermal relaxation. As a result, metamorphic grade and P-T histories vary in both space and time and are strongly dependent on structural level. The metamorphism of the upper and lower plates of the Fall Mountain nappe outlier have been documented by (Spear et al., 1983; Spear, 1986; Spear and Chamberlain, 1986; Spear et al, 1990). These papers document the dramatically different P-T paths of the different structural levels. Metamorphic studies in the root zone (Chamberlain, 1985, 1986; unpublished work of the author) have documented the similarity in early metamorphic history between rocks of the root zone and the upper plate of the Fall Mountain nappe, as well as a correlation between late-stage cross folding and metamorphism.

The purpose of this field trip is to examine the interrelationship between metamorphism and structural level in a traverse across the Bronson Hill anticlinorium. The trip begins in the Connecticut Valley metamorphic low along the Chicken Yard line and proceeds eastward up metamorphic grade and up structural level.

Two themes will be returned to repeatedly: (1) What is the relationship between metamorphic grade and structural level and (2) What is the P-T path of the rocks in each structural level and what does it tell us about the evolution of the terrane? Figure 1 shows a geologic map of the field trip area and Figure 2 shows a schematic cross section across the area showing the different structural levels illustrated by the field trip stops.

One of the main questions we will address on this trip is: "What is the cause of this metamorphic inversion observed across the Bronson Hill anticlinorium?". Inverted metamorphism is common in many metamorphic belts. In order to understand the significance of the metamorphic inversion it is crucial to determine whether the isotherms were ever inverted or are we simply looking at post-metamorphic stacking. It turns out that this is a very difficult question to answer and most, if not all, occurrences of inverted metamorphic gradients are at least consistent with the tectonic stacking occurring after the metamorphism. We will examine this question as we proceed on the field trip.

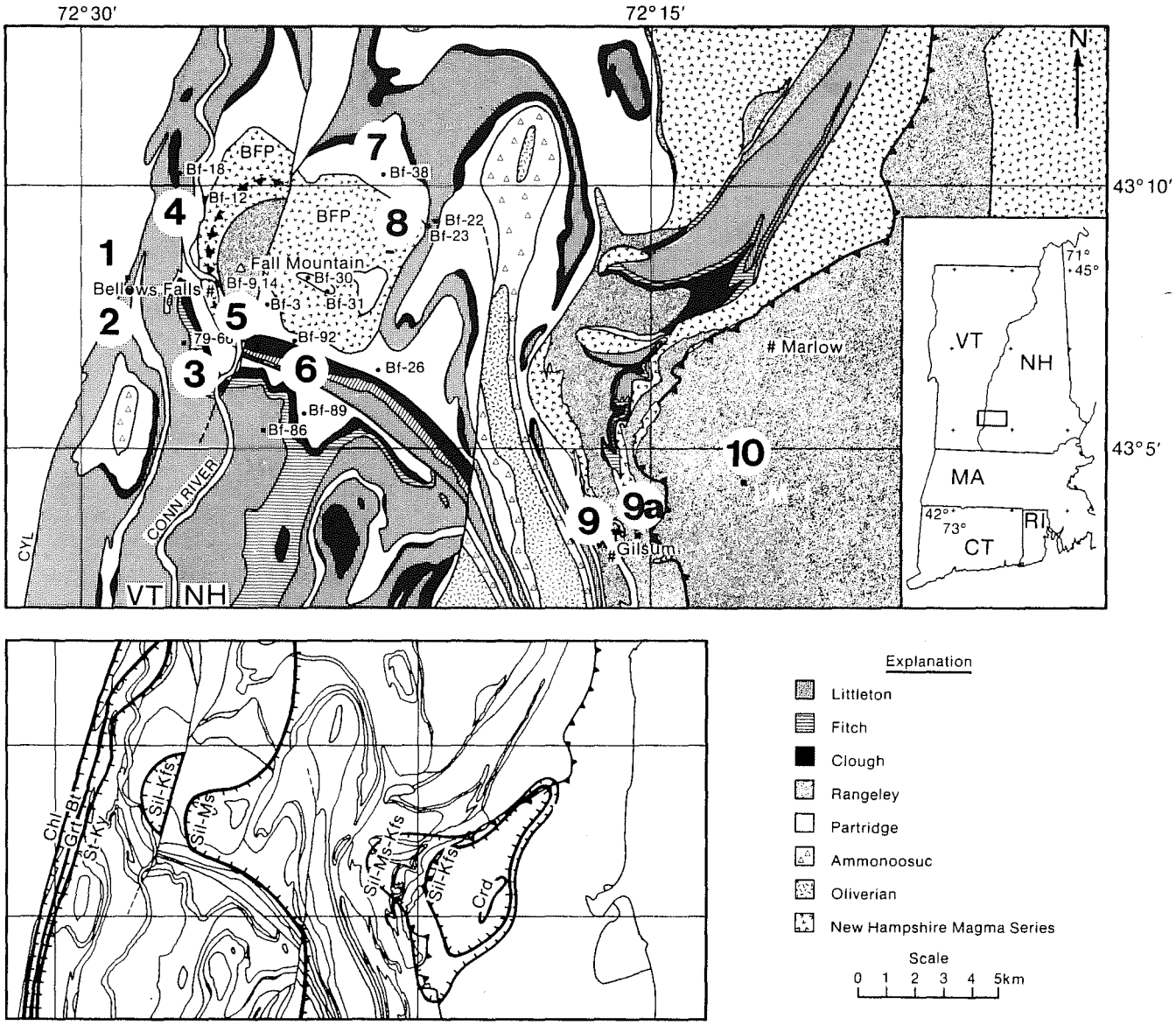


**Figure 1.** Geologic map of a part of west-central New Hampshire and adjacent Vermont depicting general geologic relations and sample locations in the vicinity of the Fall Mountain nappe and adjacent Merrimack synclinorium. BFP = Bellows Falls pluton; CYL = "chicken yard line". The Alstead dome (AD) lies along the axis of the Bronson Hill anticlinorium in the center of the map, and is cored by the Oliverian gneiss. Barbed line near Marlow is the Chesham pond thrust (CPT). Dashed barbed line within the BFP is the approximate location of the Fall Mountain thrust. Lower map shows peak metamorphic isograds. Geology after Kruger (1946), Thompson et al. (1968), Thompson and Rosenfeld (1979), Chamberlain (1985), Allen (1984) and the authors' mapping. Numbered dots show locations of field trip stops.

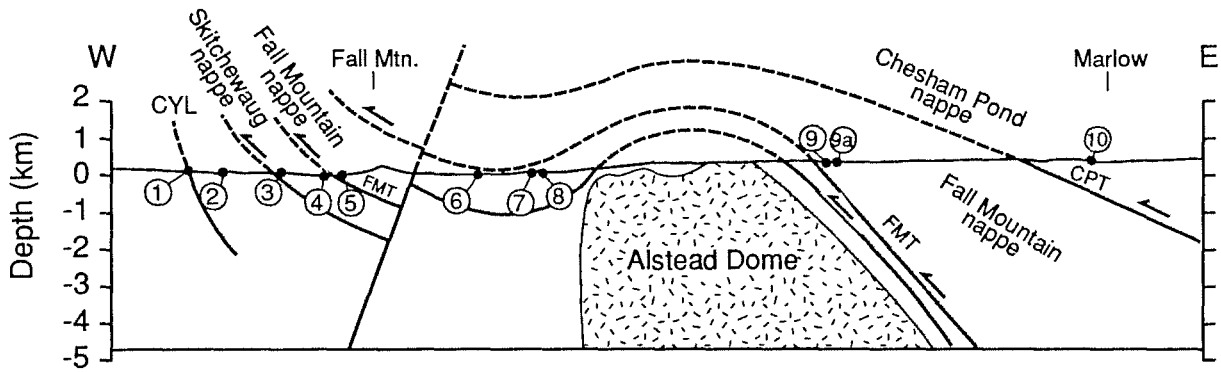


**Figure 2.** Schematic cross section (adapted and modified from Thompson et al, 1968) from the Chicken Yard Line (CYL) eastward along the field trip route. CPT = Chesham Pond thrust; FMT = Fall Mountain thrust. Numbered dots show structural position of field trip stops.

## GENERAL GEOLOGIC SETTING

(This section from Spear and Chamberlain, 1986)

### Stratigraphy

The Bronson Hill anticlinorium (BHA) and Merrimack synclinorium (MS) consist of a thick sequence of Ordovician, Silurian and Devonian clastic and carbonate metasediments. The stratigraphies of these two structural terranes are, however, somewhat different. Rocks in the Bronson Hill anticlinorium, for example, consist of Ordovician granitoid gneisses overlain by Ordovician volcanics (Ammonoosuc) and rusty, sulfidic schists (Partridge). These are, in turn, overlain by a Silurian conglomerate (Clough), and calc-silicate (Fitch) sequence, and Devonian, cyclically bedded schists (Littleton).

The occurrence of a thick sequence (greater than 2 km) of Silurian clastic sediments in the Merrimack synclinorium (Rangeley), not observed in the BHA, is consistent with the idea of an eastward thickening basin in Silurian times (Boone et al., 1970). According to this hypothesis, there was a "tectonic hinge" between a shallow-water shelf or island arc (BHA) and a deep water basin (MS during the Silurian. Thus, the thin Silurian sediments, present in the anticlinorium, thicken into coeval deep water sediments in the synclinorium. This hypothesis is supported by stratigraphic observations in higher structural levels present in the Bronson Hill anticlinorium (Chamberlain, 1985; J. B. Thompson, personal communication). J. B. Thompson and Chamberlain were able to map metasedimentary units that appear to be shelf-basin transitional facies equivalent to the Clough (shelf) quartzite and Rangeley (basin) Formation (Figure 3).

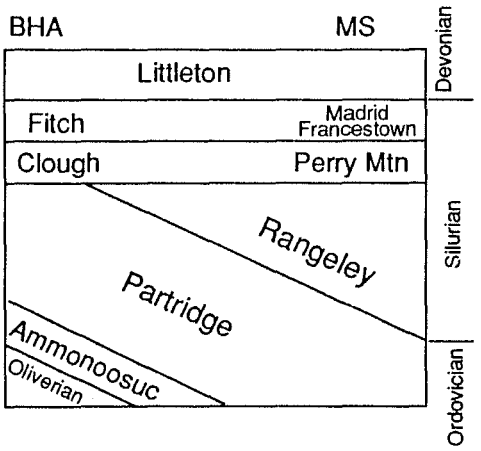


**Figure 3.** Stratigraphic column for the Bronson Hill anticlinorium (BHA) and Merrimack synclinorium (MS)

Westward transport of nappes during the Acadian Orogeny brought more basin-like sediments over the shelf sediments of the Bronson Hill anticlinorium. Because of this thrusting, higher structural levels in the anticlinorium have stratigraphic sequences more similar to the stratigraphy observed to the east in the Merrimack synclinorium. For example, two nappes discussed in this study (the Fall Mountain and Skitchewaug nappes) show many stratigraphic similarities to rocks in the synclinorium. The lower limb of the Skitchewaug nappe exposed at Bellows Falls, VT contains gray schists (Silurian?) below the Silurian Clough conglomerates and above the Ordovician Partridge schists. These gray schists thicken to the east, and in the upright limb of this nappe exposed near Gilsum, NH there are thicker sequences of gray schists between the Clough conglomerates and Ordovician metasediments (Partridge). The next highest structural level exposed in these regions, the Fall Mountain nappe, contains massive pelites (Rangeley) more similar to rocks present in the Merrimack Synclinorium. These stratigraphic complexities are the result of westward thrusting of an original east-thickening stratigraphic sequence.

**Structural Geology**

The structural history of this region is complex and involves several stages of deformation. The earliest period of deformation recognized was a period of nappe and thrust development. Napping was followed by gneiss dome formation and related folding about north-south axes (Thompson et al., 1968; Chamberlain, 1985). The latest period of folding recognized produced open folds with east-west trending axes.

Structural features resulting from all three phases of deformation can be observed both in the outlier and root zone of the Fall Mountain nappe. In the Bellows Falls area there are two large nappe structures exposed. The lower plate belongs to the overturned limb of the Skitchewaug nappe, and the upper plate is part of the Fall Mountain nappe. In this area, the nappes have been refolded about north-south axes (dome-stage folds?) and later folds with east-west trending axes. The late-stage open folds in this region are recognized both on gross outcrop pattern and by structural analysis (Allen, 1984, 1985). Allen (1984, 1985) has shown that the Fall Mountain outlier is folded into a major syncline associated with the latest Acadian folding.

East of the Alstead dome, in the root zone of the Fall Mountain nappe, similar deformation features are present. Here, the upper limb of the Skitchewaug nappe is overlain by rocks belonging to the Fall Mountain nappe. Nappe deformation in this region produced the dominant foliation observed. Later folding about north-south axes produced a spaced crenulation cleavage and abundant minor folds. The latest period of folding was about east-west trending axes, and produced no recognizable minor structural features (Chamberlain, 1985). However, the late Acadian folds are large; one synclinal axis can be traced from the Fall mountain outlier at Bellows Falls VT east into the root zone of the Fall Mountain nappe (Chamberlain, 1985).

**Plutonic Rocks**

The intrusive rocks that crop out in this region belong to the Acadian New Hampshire plutonic series (Thompson et al., 1968). Three representations of the New Hampshire plutonic series (Kinsman, Bethlehem, and Concord) are present in this region.

The Kinsman and Bethlehem members of the series are early (395-415 Ma) syntectonic intrusives (Lyons and Livingston, 1977; Aleinikoff and Moench, 1987; Moench, 1989; Moench and Aleinikoff, 1991). These rocks are slightly foliated and folded, and are thought to have been intruded along the base of the nappes during emplacement (Thompson, et al., 1968; Chamberlain, 1985; P. J. Thompson, 1985). These plutons are sheet-like bodies, no greater than 2.5 km thick, and may have once covered much of southern New Hampshire (Nielson et al., 1976). They range in composition from granite to tonalite and may represent lower crustal partial melts (Clark and Lyons, 1984). There is petrologic evidence in the Merrimack synclinorium suggesting that the Kinsman plutons have been metamorphosed (Thompson et al., 1968; Chamberlain and Lyons, 1983). Good exposures of Bethlehem pluton are exposed at the base of the Fall Mountain outlier in Bellows Falls VT and in the root zone near Marlow NH.

The Concord suite is considerably younger than the other members, with an age ranging from 380 to 275 Ma (Lyons et al., 1982). The Concord plutons typically consist of two-mica granites and appear to be post-tectonic. These granites are more abundant in the Merrimack synclinorium.

## ACKNOWLEDGEMENTS

Much of the work described in this field trip guide is the result of many years of work by the author and his students. Those to be especially acknowledged include Jane Selverstone, Don Hickmott, Matt Kohn, Frank Florence, Tom Menard, Stephan Paetzold and Danny Orange. Page Chamberlain is also to be acknowledged for his excellent mapping in the eastern part of this field trip area and showing the author several key outcrops. Special thanks go to Jack Cheney for driving around in the field and listening to the author's ideas (with which he does not necessarily agree).

## 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 | Westminister 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" as it is sometimes called.

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 in 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 counter clockwise.

One model for the evolution of the Connecticut Valley metamorphic low is that it is a synformal structure that folds high level, low grade rocks down so they are now at the same erosional level as higher grade rocks. This model predicts a symmetry of metamorphic pressures and P-T paths about the axis of the Connecticut Valley metamorphic low, which is not observed.

The model proposed here interprets the Connecticut Valley metamorphic low as a low-grade sandwich of rocks between the higher grade, inverted metamorphism seen in the high level nappes of the Bronson Hill anticlinorium, and the non-inverted ("normal") metamorphism seen in the Vermont sequence to the west. Figure 4 shows a cartoon of this idea.

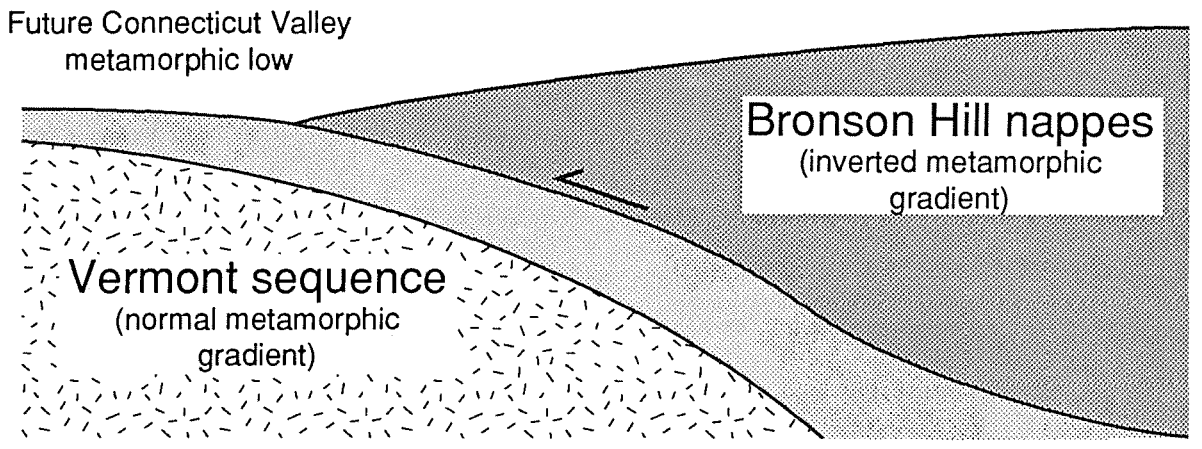


**Figure 4.** Cartoon showing the evolution of the Connecticut Valley metamorphic low. The nappes of the Bronson Hill anticlinorium are thrust over unmetamorphosed rocks that will become the Connecticut Valley metamorphic low. The highly sheared rocks along the Chicken Yard line are probably one of several surfaces along which movement has occurred. Subsequent dome formation rotates the fabric into vertical orientation and causes sufficient unroofing so the rocks do not heat beyond biotite grade.

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. Note, however, that the appearance of garnet in these rocks may be due to differences in bulk composition (e.g. higher MnO content) and not real differences in temperature. This point will be discussed on the outcrop with new microprobe data. Note that if the grade is increasing and if the model I propose is correct, then these higher grade rocks are structurally above the biotite zone rocks of the Chicken Yard line.

3.9 Continue east on Route 121.
4.6 Old Westminister Road on right (south). Continue on 121.
4.8 Pass under I-91.

Look to the left (north) and note the rocks along the far bank of Saxtons River. These rocks are in the staurolite zone and contain large staurolites with the assemblage staurolite + garnet + biotite + muscovite + quartz. There are two penetrative foliations in the rock plus a late kink banding. Staurolite overgrows both foliations. Retrogression is common with chlorite replacing staurolite. It is difficult to be sure, but it looks like the kink banding is coeval with the chloritization of the staurolite.

Note that we are only about 1 km to the east of the Chicken Yard Line and the chlorite zone rocks. I believe there is a major structural discontinuity between the garnet zone rocks of stop 2 and these staurolite grade rocks in the Saxtons river. The chlorite alteration may well have been produced during emplacement of the staurolite grade rocks onto low grade rocks and the fluids required for the alteration may have been derived from the dewatering of the lower grade rocks.

5.1 Cross Saxtons River.
5.6 Turn right at circle with flagpole
5.7 Bear left (do not cross river)
5.9 Pull off onto side of road on left (west) and park. Walk down path to east to river. **Warning–Poison Ivy.**

**Stop 3. 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 and to discuss the reaction history and P-T path of these rocks. (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 toping 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 pools 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

staurolite + biotite + $H_2O$ = garnet + chlorite + muscovite

which is a discontinuous reaction in the KFMASH system. The replacement of staurolite by a potassic phase such as muscovite is common in replacement textures (e.g. Carmichael, 1969; Foster, 1981, 1983) and attests to the higher mobility of K relative to Al. The alteration of staurolite here is completely from that observed in the rocks along the bank of the Saxtons river that we passed on our way here in that here staurolite is replaced mostly by muscovite whereas there staurolite is replaced by chlorite.

No P-T paths have been calculated from rocks at this locality, but a staurolite zone rock from along strike to the north (location BF-18) has yielded the P-T path shown in Figure 5. The path shows simultaneous heating and loading, as would be expected from a rock that was being loaded by the emplacement of higher level nappes.

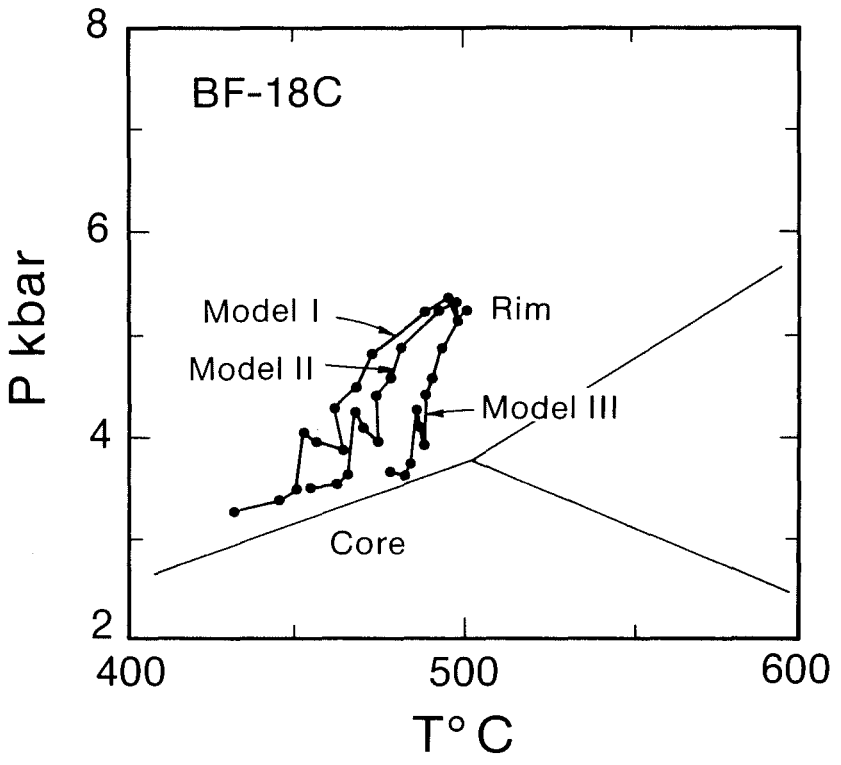


**Figure 5.** P-T paths computed for lower plate sample BF-18C. The three paths correspond to the three model assemblages: I = garnet + staurolite + biotite + chlorite; II = garnet + staurolite + biotite; and III = garnet + chlorite + biotite. From Spear et al. (1990).

5.9 Continue straight (north) on dirt road
6.55 Route 121. Turn right (east)
7.0 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 Bellows Falls center. Continue north on Route 5.
7.55 Miss Bellows Falls Diner on the left
7.7 Bridge across Connecticut River (to Rt. 12 and Walpole). Turn right and cross river to New Hampshire.
7.75 Stop light at Route 12. Turn left (north) on Route 12.
9.1 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 County line. Sullivan County
10.4 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 Skitchewaug nappe immediately below the Bellows Falls pluton and the Fall Mountain nappe.

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 ± 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 + $H_2O$ = garnet + chlorite + muscovite

(see Figure 6). Staurolite-bearing assemblages were probably overstepped as a result of the rapid change in pressure conditions.

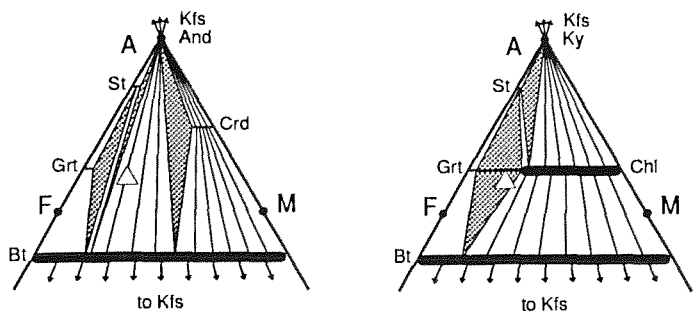


**Figure 6.** Inferred AFM diagrams for early (andalusite + biotite) and late (garnet + chlorite) parageneses.

The best collecting is along the railroad cut. Almost all samples contain andalusite pseudomorphs and, locally, relict andalusite. At the far southern end of the railroad cut is a calcareous, epidote-bearing rock, which also contains pseudomorphs and which have not been identified.

10.4 Turn the vans around and return south on Route 12.
13.0 Stoplight at bridge to Vermont. Turn left and continue south on Route 12 towards Walpole.
13.6 Stoplight at bridge to Vermont. Note excellent outcrops in river. Proceed straight (south) through stoplight.
13.7 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. If the timing is right we will eat lunch on the outcrops in the river so bring your lunch. Climb down bank to outcrops in river below.

**Stop 5. Sillimanite + K-feldspar zone rocks of the Fall Mountain nappe.** (60 minutes including lunch). (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. The contact between these units is well exposed. The upper most occurrence of pluton is a sharp contact. Proceeding down into the pluton are lenses and pods of schists, which locally have been nearly completely assimilated. Large garnets (1-2 cm radius) can be observed in the pluton in places. These are not igneous garnets but rather the relicts of assimilated schist. Dikes of pegmatite cut the pluton but nowhere have we observed plutonic rock intruding the schist. Therefore, we interpret the contact to be a shear zone with pieces of upper plate schist incorporated into the pluton.

The pluton contains plagioclase + K-feldspar + biotite + quartz. A zircon from this locality was analyzed by T. M. Harrison (of UCLA) on the SHRIMP in Canberra, Australia and he obtained an age of 407 ± 5 Ma, which is interpreted to be the crystallization age of the pluton (in Kohn et al, 1992).

The Rangeley formation is dominantly gray schist but locally there are boudins of amphibolite and calc-silicate. The large porphyroblasts in the gray schist are sillimanite pseudomorphs after andalusite ("turkey track rocks"). In a few places these sillimanites are folded with the foliation.

Assemblages observed in the gray schist include quartz + plagioclase + garnet + biotite + sillimanite ± spinel (only as inclusions within sillimanite) ± chlorite (retrograde) ± staurolite (retrograde) ± 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 ± staurolite which was produced during retrogression of the upper plate. Retrogression is common in the rocks here. The first stage occurred by the reaction

garnet + K-feldspar + $H_2O$ = sillimanite + biotite

and is responsible for the consumption of garnet in many samples. Then the reaction

sillimanite + K-feldspar + $H_2O$ = muscovite + quartz

is crossed, which converts all K-feldspar to muscovite and produces the selvages of muscovite around sillimanite. Once muscovite replaces K-feldspar, the reaction

sillimanite + biotite = garnet + muscovite

proceeds, which results in a second generation of small, idioblastic garnet. In addition, the reaction

sillimanite + biotite + garnet + $H_2O$ = staurolite

has also occurred locally, producing small staurolites in selvages around sillimanite and around some early garnets. The fluids responsible for the retrogression presumably came from dewatering of structurally lower rocks following nappe emplacement or perhaps may have come from dewatering of the pluton during crystallization. The retrogression was not pervasive, however, as some samples contain pristine sillimanites.

The P-T path of the rocks from this locality was illustrated by Spear et al (1990) and is reproduced here in Figure 7. Early metamorphism was low pressure, high temperature into the sillimanite + K-feldspar zone at a pressure of 3-4 kbar. The rocks experienced loading following peak metamorphism up to pressures of 5-6 kbar and cooling to 525 °C. The loading is interpreted to have resulted from the emplacement of another nappe structurally above the Fall Mountain nappe (now eroded away) and the cooling is interpreted to have resulted from the emplacement of the Fall Mountain nappe onto the cooler rocks of the Skitchewaug nappe.

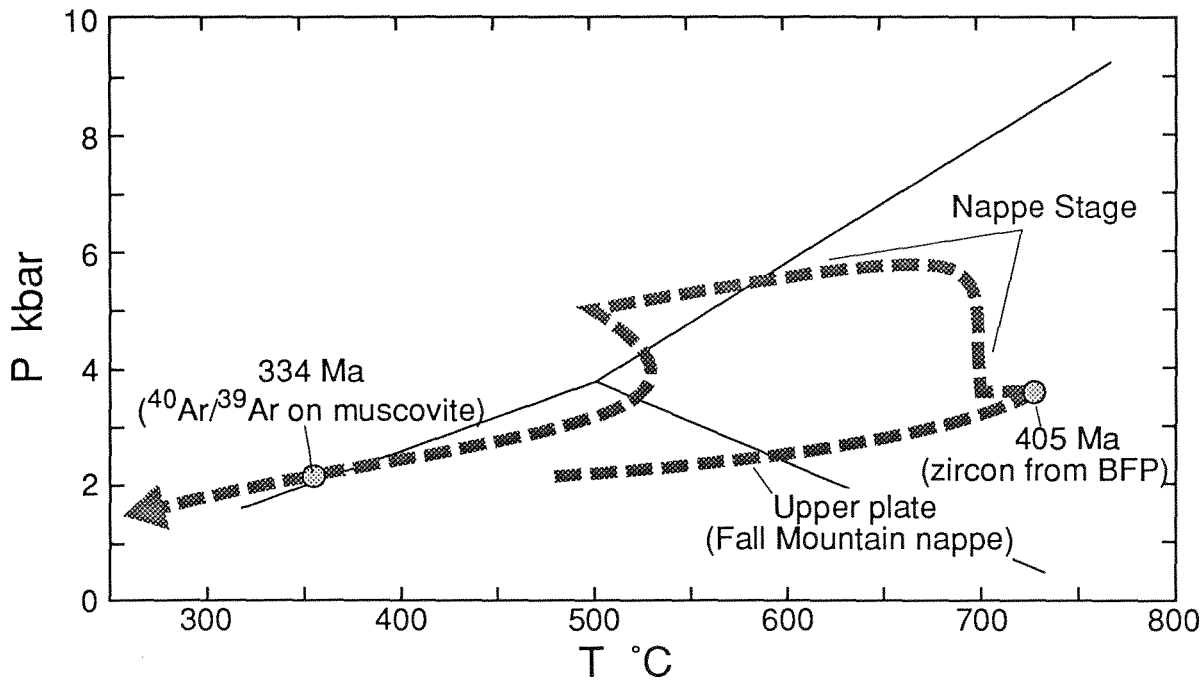


**Figure 7.** P-T path for rocks from the Fall Mountain nappe from Spear et al (1990).

The identity of the nappe that is higher than the Fall Mountain nappe is in question, because it does not crop out anywhere. However, we will present evidence later in the trip that the nappe is the Chesham Pond nappe (named here) that is comprised of still higher grade rocks (garnet + cordierite zone).

One further point to make about timing of metamorphism and thrust emplacement. Rocks of the Fall Mountain nappe were at 700-750 °C at their peak temperature whereas rocks of the Skitchewaug nappe were at 550 °C at their thermal peak. Therefore, the Skitchewaug nappe could not have been immediately below this outcrop throughout its history (or else it would be at the same temperature as this rock). My interpretation is that we have a series of in sequence thrusts with the Fall Mountain nappe being first loaded from above by another nappe, then the pair riding to the west along on the Bellows Falls pluton and then picking up the Skitchewaug nappe when the Fall Mountain nappe had cooled somewhat. We will see later in stop 8 that the lower contact of the Bellows Falls pluton also assimilates part of the Skitchewaug nappe, so the shortening must have occurred within the pluton itself.

13.7 Continue south on N.H. Route 12.
14.1 Historic marker turnoff
14.65 Turn left (east) onto Route 123 south
15.1 Turn left following Route 123 towards Alstead
15.4 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 Intersection at Old Drewsville Road. Go straight.
16.3 Pull off on right side of road. Outcrop is on left side of road.

**Stop 6. Staurolite-kyanite (sillimanite?) zone, Skitchewaug 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 Skitchewaug 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.

The interpretation of this rock is that it experienced initial high-T, low-P metamorphism during emplacement of the Bellows Falls pluton and the initial assemblage was andalusite + biotite, similar to that observed along the railroad tracks at stop 4. Then, when the Skitchewaug nappe was loaded during emplacement of the higher Fall Mountain nappe (and possibly the Chesham Pond nappe) the andalusite + biotite tie lines were replaced by garnet + chlorite tie lines by the reaction

andalusite + biotite + $H_2O$ = garnet + chlorite + muscovite

similar to stop 4. In contrast to stop 4, however, the rocks here continued to heat subsequent to loading, and the garnet + chlorite association reacted to produce staurolite + biotite via the reaction

garnet + chlorite + muscovite = staurolite + biotite + $H_2O$

and then kyanite appeared via the reaction

staurolite + chlorite + muscovite = kyanite + biotite + $H_2O$

Sillimanite then appears within muscovite presumably because we have crossed the kyanite = sillimanite boundary. The peak metamorphic assemblage is therefore interpreted to be garnet + chlorite + biotite + staurolite + sillimanite + muscovite + quartz (kyanite is metastable in the sillimanite field). This low variance assemblage requires that garnet is present because of extra components (garnet composition is $X_{alm} = 0.66$, $X_{prp} = 0.14$, $X_{sps} = 0.17$, $X_{grs} = 0.03$) and that $\mu_{H_2O}$ is buffered by the mineral assemblage. Alternatively, the chlorite could all be retrograde, but it does not appear to be on textural criteria.

The higher peak grade here than in the west (e.g. stop 4 at the railroad tracks) reflects the regional post-nappe

emplacement metamorphic gradient, as seen on the isograd map.

Other rocks from the area display similar pseudomorph textures as those shown here (see Figure 9b of Spear et al., 1990). In many rocks the early pseudomorph is readily identified, but in some only a ghostly clot of white mica is observed. The nearly ubiquitous presence of these types of pseudomorphs in the Skitchewaug nappe suggests that the entire structural level experienced a similar early high temperature, low pressure metamorphism.

P-T paths from this and nearby rocks have been discussed by Spear et al. (1990). Results from sample BF-86 are shown in Figure 8. The early path is within the andalusite field and shows a period of isobaric heating, presumably caused by intrusion of the Bellows Falls pluton, followed by a period of nearly isothermal loading, presumably caused by emplacement of the Fall Mountain or Chesham Pond nappe.

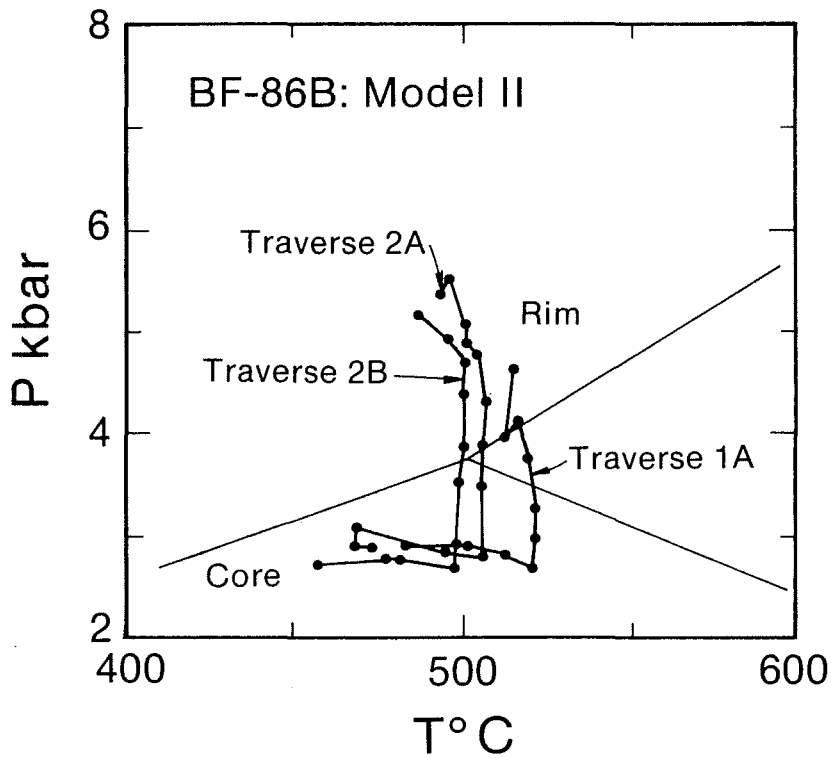


**Figure 8.** Preferred P-T paths from lower plate sample BF-86B computed from the method of Spear and Selverstone (1983). (Model II from Figure 13b of Spear et al., 1990). Paths were computed assuming the assemblage garnet + chlorite + biotite near the rim of the garnet and garnet + andalusite + biotite near the core.

16.3 Continue in the same direction (to the east)
17.3 Turn left (north) on Old Cheshire Turnpike. The road turns to dirt after a bit.
18.8 Drewsville. Intersection with route 123. Proceed to stop sign at Route 123 (the main road) and turn right towards Alstead (do not turn right onto the local village road).
19.1 After bridge at light turn right (follow 123) to Alstead.
19.3 County line
19.7 Beware of the moose!

20.25 Town line
21.0 Stop sign. Turn left, then immediately right up the hill. Road turns to dirt after a short distance.
22.5 "T" junction. Turn right.
22.52 Turn off onto path on left between stone walls. Park.

**Stop 7. Staurolite zone, Skitchewaug nappe (lower plate to the Fall Mountain nappe)** (20 minutes). (Bellows Falls 15' quadrangle; Bellows Falls 7 1/2 minute quadrangle. Sample location BF-38). The purpose of this stop is to examine the beautiful large staurolite rocks. Please respect the builder of the stone wall and do not collect from the wall! Ample samples to collect in small quarry 50 meters across field to the north.

The rocks here are in the Littleton Formation in the Skitchewaug nappe and the most obvious feature is the large staurolite crystals that are seen on every surface. The staurolites are late in the paragenetic sequence and overgrow early foliations as well as include garnets. It is quite clear from detailed petrology on these and similar samples that the staurolite producing reaction is

garnet + chlorite + muscovite = staurolite + biotite + $H_2O$

In addition to staurolite, this reaction produces a substantial quantity of biotite which in these rocks can readily be seen cross cutting the fabric.

The growth of large crystals such as these requires small amounts of overstepping of reaction, which can be achieved through slow heating. The staurolite producing reaction is clearly post nappe fabric, so the existence of these large staurolites leads us to conclude that the rocks heated slowly for a period of time following nappe emplacement, probably on the order of a few tens of degrees. Pressures during staurolite growth based on plagioclase - garnet barometry are 5-6 kbar and there is no measurable change in the depth of these samples during staurolite growth.

Perhaps the most startling feature of these large staurolite rocks is the extent of their exposure. You can find rocks identical to these in appearance, fabric, assemblage, paragenesis and texture in a belt that begins near Littleton, New Hampshire and extends at least as far south as Keene, New Hampshire, for a strike distance of 200 km. Whatever process affected these rocks causing them to attain their distinctive appearance was reasonably homogeneous over at least this distance.

22.52 Turn around and retrace path towards Alstead. Turn right on dirt road.
22.7 Turn left and go down hill towards Alstead
22.8 Walker Hill Road on right — go straight.
23.7 Blacktop begins
24.1 Church — bear left
24.2 Intersection. Turn left onto 123 and cross river.
24.25 Bear left (follow 123)
24.3 Pass Alstead village store
24.75 Pull out on left. Park vans and climb down to river.

**Stop 8. Sillimanite zone, Skitchewaug nappe** (lower plate to the Fall Mountain nappe) (20 minutes). (Bellows Falls 15' quadrangle; Bellows Falls 7 1/2 minute quadrangle. Sample location BF-23). This stop is to examine the Skitchewaug nappe in the sillimanite zone and to observe the intrusive relations between the Bellows Falls pluton and the Littleton Formation of the Skitchewaug nappe.

Rocks exposed in the river are the Bellows Falls pluton and the Littleton Formation of the Skitchewaug nappe. Pieces of partly assimilated schist are observed within the pluton similar to the relations observed between the pluton and the upper plate at stop 5. Here we have evidence that the contact between the lower plate and the pluton is part intrusive and part a shear zone. Our interpretation is that this is a sheared intrusive contact.

The assemblage observed in the schists at this outcrop is garnet + staurolite + biotite + muscovite + quartz. Up

stream a few tens of meters are exposed outcrops of schist with large staurolite and, locally, large sillimanites. The regional grade is therefore sillimanite + muscovite (+ biotite). The large staurolites here overgrow all fabrics, similar to the last stop.

24.75 Exit pull out and turn left onto Route 123.
25.6 Follow Route 123 towards Marlow (do not take 123A)
27.0 Enter valley. This is the Alstead dome. The Oliverian magma series weathers readily and so the domes often form valleys.
28.65 Old mill. Cross bridge.
28.7 Bear right onto dirt road that follows the western side of Lake Warren.
30.0 Intersection. Turn right (south).
30.6 Outcrops on left (east) side of road are the middle member of the Silurian Rangeley Formation in the sillimanite zone.
33.7 Small outcrop on left. Pull vans off on right and stop. Be careful of traffic.

**Stop 9. Sillimanite-muscovite zone, Root zone of Fall Mountain nappe** (20 minutes). (Bellows Falls 15' quadrangle; Walpole 7 1/2 minute quadrangle. Sample location 89-1). This stop is to examine rocks of the root zone of the Fall Mountain nappe.

The rocks exposed at this stop are gray schists of the Silurian Rangeley formation (middle member). The peak metamorphic assemblage is garnet + biotite + sillimanite + muscovite + quartz. Two generations of garnet are observed, both of which are small (on the order of 1 mm diameter or smaller). The early generation garnet is found in the matrix of the rock and is xenoblastic. The late generation is found within clots of fibrolitic sillimanite and is idioblastic. Zoning in both generations is identical. (Core: $X_{prp} = 0.10$; $X_{alm} = 0.71$; $X_{sps} = 0.16$; $X_{grs} = 0.03$; Fe/(Fe+Mg) = 0.88. Rim: $X_{prp} = 0.09$; $X_{alm} = 0.78$; $X_{sps} = 0.09$; $X_{grs} = 0.04$; Fe/(Fe+Mg) = 90.

The reaction(s) responsible for growth of early garnet is not known, but is likely to have been the breakdown of chlorite:

chlorite + quartz = garnet + $H_2O$

One of the curious features of the sillimanite zone rocks in the vicinity of the Alstead dome is that they show no evidence for ever having had staurolite in the assemblage. That is, the rocks appear to have jumped directly from garnet grade into the sillimanite grade. The sillimanite producing reaction is therefore probably

garnet + chlorite + muscovite = sillimanite + biotite + $H_2O$

which would result in the corrosion of the early generation garnet. Skipping staurolite parageneses is expected in rocks with sufficient MnO and the cores of garnets from this rock contain $X_{sps} = 0.14$-$0.16$. Some of the sillimanite produced by this reaction occurs replacing muscovite that has been folded in the last fabric producing episode, so this reaction proceeds to the right after most of the deformation had occurred. This reaction has a relatively steep, positive dP/dT slope.

Production of secondary garnet occurred by the reaction

sillimanite + biotite = garnet + muscovite

This reaction is water-absent and has relatively flat dP/dT slopes in the sillimanite field. Production of garnet occurs as the reaction is crossed towards increasing pressure first as temperature is increasing and then along the cooling part of the path. Our interpretation is that the pressure increase was caused by nappe emplacement.

The zoning in the garnet shows an increase in Fe/Mg towards the rim. From Figure 9 it can be seen that crossing the reaction isopleths towards increasing pressure (in order to grow garnet) would produce a garnet that

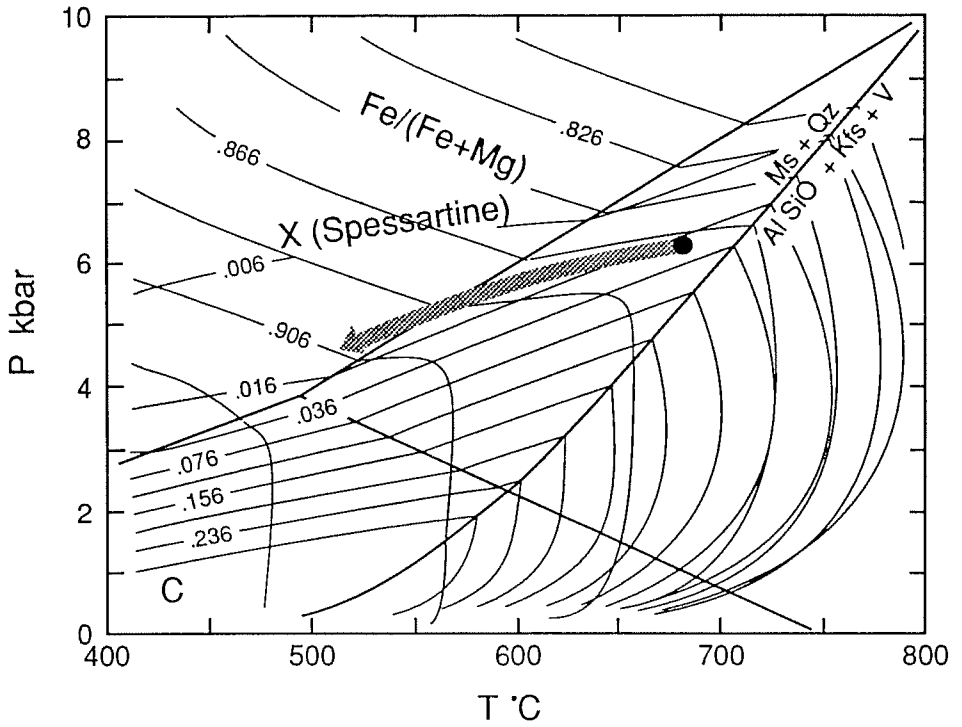


**Figure 9.** P-T diagram showing contours of $X_{sps}$ and Fe/(Fe+Mg) for the assemblage garnet + $Al_2SiO_5$ + biotite + quartz + muscovite or K-feldspar. The zoning in garnet from this location, and the production of secondary garnet, is consistent with a late P-T path of nearly isobaric cooling.

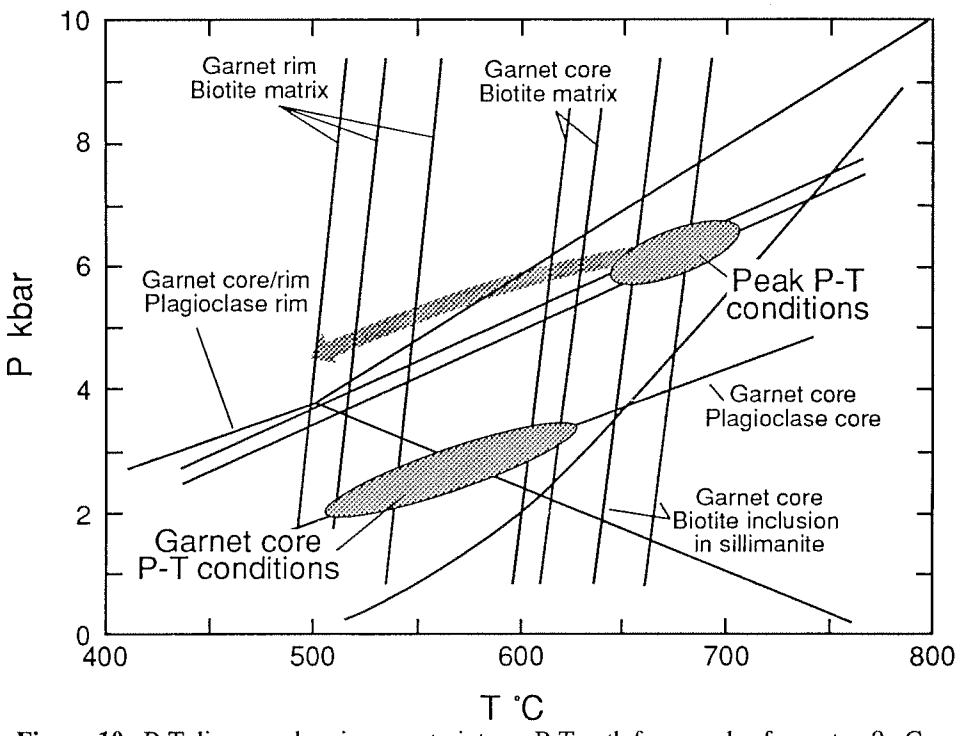


**Figure 10.** P-T diagram showing constraints on P-T path for samples from stop 9. Garnet-plagioclase barometry using inclusions of plagioclase in garnet and zoning in plagioclase are consistent with a counter clockwise P-T path. Garnet zoning in secondary garnets constrains the nearly isobaric cooling path (see Fig. 9).

zones towards lower Fe/Mg at the rim - just the opposite of what is observed. At a peak temperature of 675 °C, however, a garnet with a radius of 1 mm will homogenize at cooling rates of 10 °C/Ma or slower. Therefore, we interpret all of the zoning in these garnets to be the product of diffusion during cooling from the metamorphic peak. Cooling along the P-T path shown in Figure 9 will produce the observed zoning (increasing Fe/Mg, decreasing $X_{sps}$, increasing $X_{alm}$, and relatively constant $X_{prp}$ and $X_{grs}$.

The P-T path here is similar to the one derived for the Fall Mountain nappe at stop 5 (see Fig. 10). The early low pressure, high temperature part of the P-T path is documented by sillimanite pseudomorphs after andalusite, which are found in nearby rocks (see stop 9A) and by garnet-plagioclase geobarometry using plagioclase core compositions and inclusions in garnet. Peak metamorphic conditions are documented by garnet-biotite thermometry and garnet-plagioclase barometry using garnet cores, biotite inclusions in large sillimanite and plagioclase rims. The nearly isobaric cooling path is constrained by the zoning in garnet.

33.7 Continue straight (south)
33.85 Rusty outcrops on right are Partridge formation
34.4 Side road on left. This is where optional stop 9A begins.

Optional stop 9A

0.0 Make very sharp left turn (140 degrees) onto side road.
0.25 Road turns to dirt
0.4 Pull of and park on left so that you do not block road. To find the outcrops you must cross the road (to the east), enter the woods, cross the small stream and walk along contour for approximately 100 meters. There you will find ledges approximately 2 meters high of Rangeley Formation with large prismatic sillimanites pseudomorphed after andalusite.

**Stop 9A. Sillimanite-muscovite zone, Root zone to Fall Mountain nappe** (20 minutes). (Bellows Falls 15' quadrangle; Walpole 7 1/2 minute quadrangle. Sample location BF-78). This stop is to examine more rocks of the root zone of the Fall Mountain nappe.

The conspicuous feature of the rocks from this outcrop are the large prismatic sillimanite crystals that are pseudomorphed after andalusite. Many examples of relict chiastolitic crosses can be found in samples from this locality. Included in sillimanite are biotite and hercynite spinel, the same as are found in the Connecticut River below Fall Mountain. There is little doubt that these are the same rocks and they have undergone a very similar metamorphic history. One significant difference is the absence of any retrogression on the rims of the sillimanite crystals, which was so prevalent at Fall Mountain. The reason, it is believed, is that these rocks were not emplaced on schists that underwent significant dewatering following thrust emplacement. In the absence of significant fluid influx, retrogression is not possible.

The assemblage here is garnet + biotite + sillimanite + muscovite + quartz + plagioclase. This is one of the outcrops discussed by Chamberlain (1986; see also Spear and Chamberlain, 1986, stop F4). Chamberlain mapped several generations of folds in this area and this outcrop is located where the F2 and F3 anticlines intersect. The P-T path as presented by Chamberlain (1986) and as deduced by our work is one of isobaric cooling identical to that shown in Figures 9 and 10.

Return to vans. Turn around in triangle and retrace steps to main road. Pick up road log at the stop sign at the main road at 34.4 miles

34.4 Main road. Turn left towards Gilsum.
34.45 Gilsum center. Turn left at monument in center of town. Note minerals in monument. Gilsum is the site of a large mineral show every year.
34.7 Stop sign. Turn left onto NH Route 10.
37.5 Pull off road onto shoulder on right and park.

**Stop 10. Garnet-cordierite zone, Chesham Pond nappe** (20 minutes). (Lovewell Mountain 15' quadrangle; Stoddard 7 1/2 minute quadrangle. Sample location LM-1; See Spear and Chamberlain, 1986, Stop F5). This stop is to examine the metamorphism in the structurally highest rocks. Inasmuch as this locality sits above the Chesham Pond thrust, we have called this structural unit the Chesham Pond nappe.

This outcrop is Silurian Rangeley formation (middle member) and contains the highest-grade pelitic assemblages found in the area. The pelitic rocks contain the prograde assemblage garnet + cordierite + biotite + alkali feldspar + sillimanite + plagioclase. Alkali feldspars are large (several cm across) and may not be present in a thin section sized sample. They are very apparent in outcrop, however. Cordierite is usually observed in greenish clots, which are green because of chlorite alteration. Fresh cordierite is not easy to find.

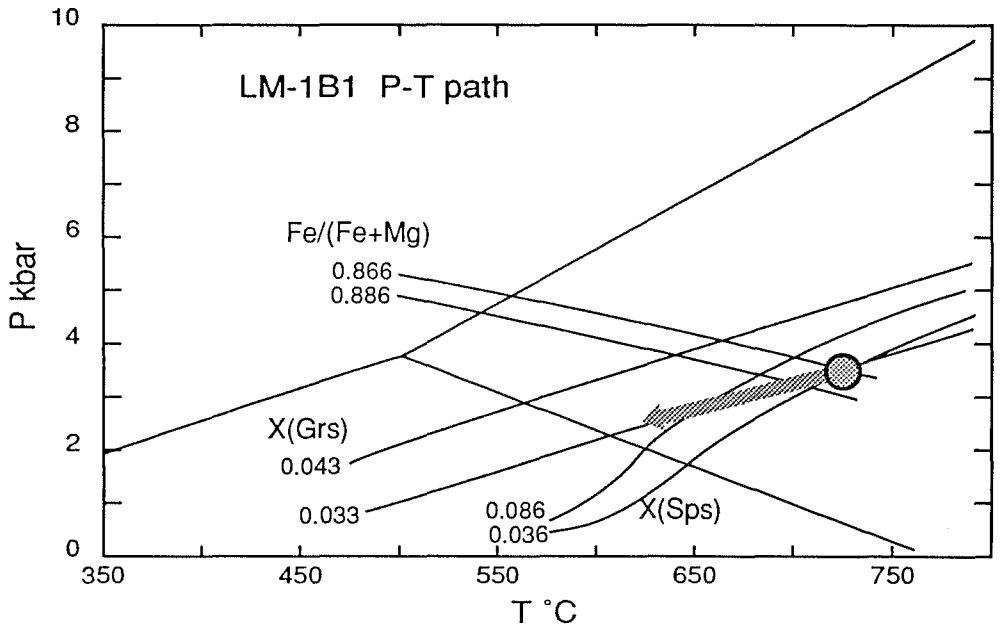


**Figure 11.** P-T diagram showing contours of $X_{sps}$, $X_{grs}$, and Fe/(Fe+Mg) for the assemblage garnet + cordierite + biotite + sillimanite + K-feldspar + quartz + plagioclase. The only P-T path consistent with the observed zoning and reaction progress is one of nearly isobaric cooling.

The P-T data obtained from garnet-biotite thermometry and garnet-plagioclase barometry indicates peak P-T conditions of 740 °C, 3.5 kbar (see Fig. 11). There is abundant textural evidence for the replacement of garnet and cordierite by sillimanite and biotite, which we interpret as the reaction

garnet + cordierite + K-feldspar + $H_2O$ = sillimanite + biotite

This reaction, plus the garnet zoning produced by this reaction suggests a P-T path of nearly isobaric cooling, as shown in Figure 11. (See Chamberlain, 1986, Spear and Chamberlain, 1986, stop F5, and Selverstone and Chamberlain, 1990 for an alternative interpretation of the reaction texture, P-T path and tectonic significance. This outcrop is located at the intersection of two large F2 and F3 Acadian synclines and Chamberlain interpreted the metamorphism at this outcrop to have resulted from thermal effects that occurred during downfolding.)

We interpret the P-T path to be the result of cooling following thrust emplacement. It is highly significant that the peak pressure experienced by these rocks is only on the order of 3.5 kbar, as contrasted with 6 kbar experienced by rocks at the previous two stops. Stop 10 is only 3-4 km east of stops 9 and 9A, so the difference in pressure requires some amount of tectonic thinning following attainment of peak pressures. We suggest that backsliding

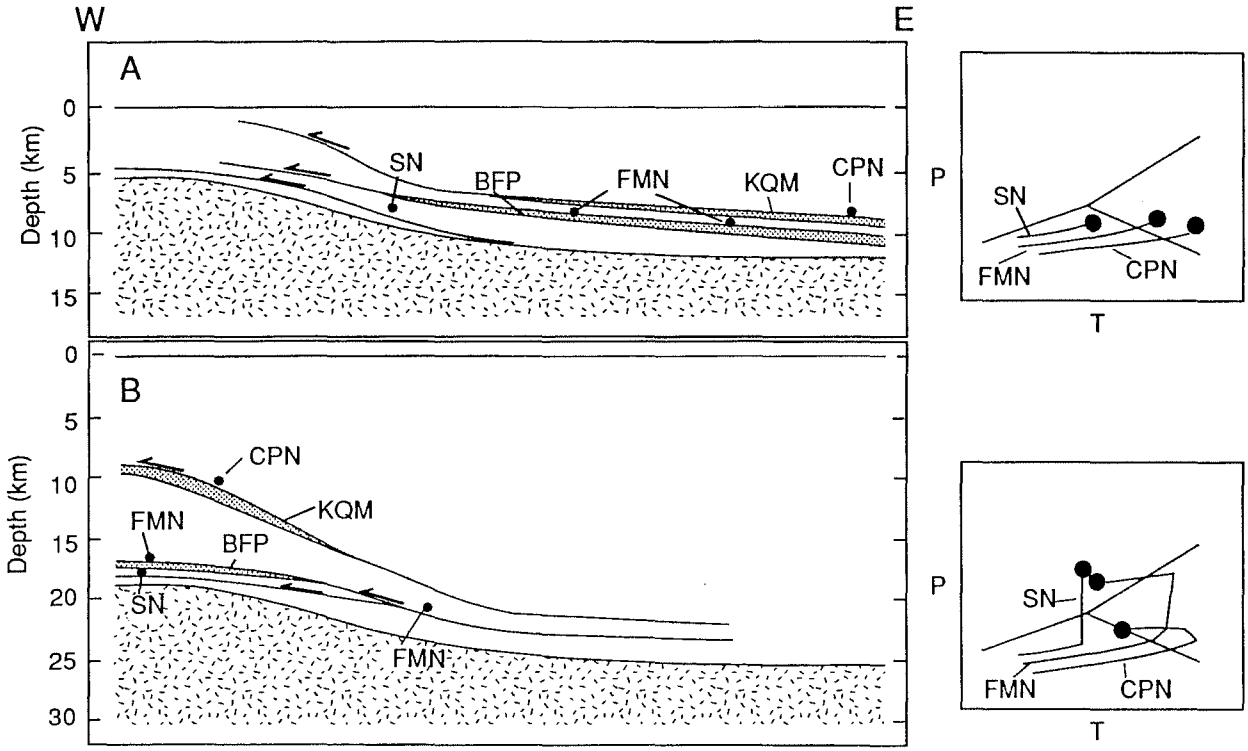


**Figure 12.** "Tectoon" showing the evolution of the Acadian orogeny during the pre and early nappe stages. (a) shows the pre-thrusting environment and the intrusion of the Bethlehem gneiss and Kinsman quartz monzonite. (b) shows the relative juxtaposition of thrusts. Note that in this model the low 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.

along the Chesham pond thrust may be responsible for the tectonic thinning.

**End of trip**

## CONCLUDING REMARKS AND "TECTOON"

It is clear from the traverse we have taken on this field trip, that the metamorphic grade increases from west to east across the Bronson Hill anticlinorium and into the Merrimack synclinorium. It is also apparent that the structural level increases as one traverses from west to east so that the highest grade rocks are found in the highest structural level. This requires structural readjustment of the metamorphic zones along major thrusts. Some of the metamorphism is early (e.g. the pervasive pseudomorphs in the Skitchewaug and Fall Mountain nappes) and some is post-thrusting (most of the final isograds). The metamorphic pattern is quite difficult to sort out unless metamorphic paths within individual structural units are examined.

Although it is always dangerous to do so, it is important to try and illustrate our thoughts about the evolution of this metamorphic belt, as shown in the "tectoon" in Figure 12. The pre-tectonic setting of eastern North America was one of a marginal basin (presumably 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. It is most likely that the entire basin experienced high T/P metamor-

phism as a result of high heat flow into the basin, high heat production from radioactive decay (Chamberlain and Sonder, 1990) and advection of heat by plutons (Fig. 12a).

Thrusting must have begun at approximately this time. The large sillimanite pseudomorphs after andalusite are lineated and folded by nappe stage fabrics so at least some deformation must have occurred during or following the emplacement of the plutons. A particularly attractive hypothesis is that the thrusts moved, in part, along the plutons themselves, in a mechanism that Hollister and Crawford (1986) have described as "melt enhanced deformation". Field evidence for this is the shearing at both the upper and lower contacts of the Bellows Falls pluton and the incorporation of high grade mineral assemblages from the schist into the pluton itself.

The most reasonable explanation for the inverted metamorphism across the Bronson Hill anticlinorium is a series of in-sequence thrust duplexes. The Chesham Pond thrust is the most easterly and is therefore the first to move. The 3.5 kbar maximum pressure experienced by these rocks requires that they were structurally high and the absence of any tectonic loading in the P-T paths suggests that this was the highest level nappe. The Fall Mountain nappe was transported along the Bethlehem gneiss (Bellows Falls Pluton) and emplaced upon the Skitchewaug nappe (Fig. 12b).

At some point the deformation zone must have intersected the basement gneisses of the Taconian island arc (the Oliverian magma series) and we believe that the most reasonable interpretation of the gneiss domes is as a series of ramp anticlines where basement has been caught up in the zone of deformation (not illustrated in Fig. 12). This, of course, requires a basal decollement beneath the gneiss domes along which transport can occur. This decollement must break the surface and the most likely spot is between trip stops 2 and 3 (between the garnet and staurolite zones). This is in the vicinity of the Bernardston nappe.

This thrust probably broke the surface and transported the entire package westward over the then unmetamorphosed sediments which now comprise the Chicken Yard line. Additional movement apparently occurred still lower down along the Chicken Yard line itself.

It is not clear how far west the Acadian deformation zone proceeded, but it is very reasonable that the Chester dome is another ramp anticline formed as Grenville basement entered the deformation zone. Deformation finally stopped as the Chester dome ramped up onto the Green Mountains.

## REFERENCES CITED

Aleinikoff, J. N. and Moench, R. H., 1987, U-Pb geochronology and Pb isotopic systematics of plutonic rocks in northern New Hampshire - ensimatic vs. ensialic sources: Geological Society of America Abstracts with Programs, v. 19, p. 1.

Allen, T., 1984, The Fall Mountain outlier: a piece of the Fall Mountain nappe [B. A. thesis]: Harvard University.

Allen, T., 1985, A reinterpretation of the Fall Mountain nappe, as seen from Fall Mountain, North Walpole, New Hampshire: Geological Society of America Abstracts with Programs, v. 17, p. 2.

Boone, G. M., Boudette, E. L. and Moench, R. H., 1970, Bedrock geology of the rangeley Lakes-Dead River region, western Maine: 24 p.

Carmichael, D. M., 1969, On the mechanism of prograde metamorphic reactions in quartz-bearing pelitic rocks: Contributions to Mineralogy and Petrology, v. 20, p. 244-267.

Chamberlain, C. P., 1985, Tectonic and metamorphic history of a high-grade terrane, southwestern New Hampshire [Ph.D. thesis]: Harvard University, Cambridge, Massachusetts, 207 p.

Chamberlain, C. P., 1986, Evidence for the repeated folding of isotherms during regional metamorphism: Journal of Petrology, v. 27, p. 63-89.

Chamberlain, C. P. and Lyons, J. B., 1983, Pressure, temperature and metamorphic zonation studies of pelitic schists in the Merrimack Synclinorum, south-central New Hampshire: American Mineralogist, v. 68, p. 530-540.

Chamberlain, C. P. and Sonder, L. J., 1990, Heat-producing elements and the thermal and baric patterns of metamorphic belts: Science, v. 250, p. 763-769.

Clark, R. G. and Lyons, J. B., 1984, Mixing models for the Kinsman quartz monzonite: Geological Society of America Abstracts with Programs, v. 16, p. 9.

Foster, C. T., Jr., 1981, A thermodynamic model of mineral segregations in the lower sillimanite zone near Rangeley, Maine: American Mineralogist, v. 66, p. 260-277.

Foster, C. T., Jr., 1983, Thermodynamic models of biotite pseudomorphs after staurolite: American Mineralogist, v. 68, p. 389-397.

Hollister, L. S. and Crawford, M. L., 1986, Melt-enhanced deformation - A major tectonic process: Geology, v. 14, p. 558-561.

Kohn, M. J., Spear, F. S., Orange, D. L., Rumble, D., III and Harrison, T. M., 1992, Pressure, temperature, and structural evolution of west-central New Hampshire: Hot thrusts over cold basement: Journal of Petrology, v. 33, p. 521-556.

Kruger, F. C., 1946, Structure and Metamorphism of the Bellows Falls Quadrangle of New Hampshire and Vermont: Bulletin of the Geological Society of America, v. 57, p. 161-206.

Lyons, J. B., Boudette, E. L. and Aleinikoff, J. N., 1982, The Avalonian and Gander zones in central eastern New England: in J.S.J.a. Beland, ed., Major Structural Zones and Faults of the Northern Appalachians: Geological Association of Canada, p. 43-66.

Lyons, J. B. and Livingston, D. E., 1977, Rb-Sr age of the New Hampshire plutonic series: Geological Society of America Bulletin, v. 88, p. 1808-1812.

Moench, R. H., 1989, Metamorphic stratigraphy and structure of the Connecticut Valley area, Littleton to Piermont, New Hampshire: in ed., A transect through the New England Appalachians: Washington, D. C., American Geophysical Union, p. 45-53.

Moench, R. H. and Aleinikoff, J. N., 1991, Geologic map of the Littleton-Moosilauke-Piermont area, NH0VT: type area of the Piermont allochthon: Geological Society of America Abstracts with Programs, v. 23, p. 106.

Nielson, D. L., Clark, R. G., Lyons, J. B., Englund, E. J. and Borns, D. J., 1976, Gravity models and mode of emplacement of the New Hampshire Plutonic Series: in P.C. Lyons and Brownlow A.H., ed., Studies in New England Geology: Geological Society of America Memoir 146, p. 301-318.

Selverstone, J. and Chamberlain, C. P., 1990, Apparent isobaric cooling paths from granulites: two counter examples from British Columbia and New Hampshire: Geology, v. 18, p. 307-310.

Spear, F. S., 1986, P-T paths from central New England: The evolution of a paired metamorphic belt: Geological Society of America Abstracts with Programs, v. 18, p. 68.

Spear, F. S. and Chamberlain, C. P., 1986, Metamorphic and tectonic evolution of the Fall Mountain nappe complex and adjacent Merrimack synclinorium: in P. Robinson, ed., Regional metamorphism and metamorphic phase relations in northwestern and central New England: Amherst, Massachusetts, International Mineralogical Association, University of Massachusetts, p. 121-143.

Spear, F. S., Hickmott, D. D. and Selverstone, J., 1983, Petrologic evidence for a metamorphic discontinuity-Bellows Falls area, New Hampshire and Vermont: EOS, v. 64, p. 878.

Spear, F. S., Hickmott, D. D. and Selverstone, J., 1990, Metamorphic consequences of thrust emplacement, Fall Mountain, New Hampshire: Geological Society of America Bulletin, v. 102, p. 1344-1360.

Spear, F. S. and Selverstone, J., 1983, Quantitative P-T paths from zoned minerals: Theory and tectonic applications: Contributions to Mineralogy and Petrology, v. 83, p. 348-357.

Thompson, J. B. J., Robinson, P., Clifford, T. N. and Trask, N. J., 1968, Nappes and gneiss domes in west-central New England: in E.-a. Zen, White W.S., Hadley J.B. and Thompson J.B.J., ed., Studies of Appalachian Geology, Northern and Maritime: New York, John Wiley, p. 203-218.

Thompson, J. B. J. and Rosenfeld, J. L. R., 1979, Reinterpretation of nappes in the Bellows Falls-Brattleboro area, New Hampshire - Vermont: in J.W.S.a.P.H. Osberg, ed., The Caladonides in the USA - Geological Excursions in the Northeast Appalachians,: Weston, Massachusetts, Weston Observatory, p. 117-121.

Thompson, P. J., 1984, Stratigraphy and structure of the Monadnock quadrangle, New Hampshire: refolded folds and associated fault zones: Geological Society of America Abstracts with Programs, v. 16, p. 67.

Thompson, P. J., 1985, Stratigraphy and structure of the Monadnock quadrangle, New Hampshire [Ph. D. thesis]: University of Massachusetts, Amherst Massachusetts.

# GLACIATION OF THE WORCESTER PLATEAU, WARE-BARRE AREA, AND EVIDENCE FOR THE SUCCEEDING LATE WOODFORDIAN PERIGLACIAL CLIMATE

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## INTRODUCTION

This trip examines the erosional and depositional effects of multiple glaciations on the rock surface of the uplands of central Massachusetts. Erosional effects are: 1) removal of the pre- and interglacial soils and most of the saprolite in the granitic and the deeply oxidized, sulfidic, schistose rocks, 2) glacial erosion of smooth, U-shaped, valley-bottom surfaces, and 3) glacial scouring of closed basins in the relatively less resistant rock units. Eroded rock materials accumulated in sulfidic Illinoian tills in drumlins, in the thin, late Wisconsinan surface till sheet, and, in the valleys, in the coarse-grained glacial terrace deposits and fine-grained lake-bottom deposits (Mulholland, 1974). Because of the sulfidic composition of the terrace sand and gravel deposits and its deleterious effect on concrete (J.A. Sinnott, pers. comm., 1992), these deposits have not been excavated extensively, and are preserved for studies of their classical glacial morphology and their sedimentology. This trip visits deep exposures in a variety of glacial landforms, some of which reveal the glaciodeltaic origin of the thick ice-contact deposits. The thick sequence of sediments in deltaic morphosequences contrast with the relatively thin fluvial sediments in the Ware outwash deposit. Grain-size analysis of maximum gravel clast sizes aids in identifying individual heads of morphosequences.

The distribution and altitudes of the glaciodeltaic and lake-bottom sediments in the valleys in the Ware-Barre area are related to the extent of glacial lakes that trapped the meltwater sediments. Recognition of sedimentary facies, individual deltaic morphosequences, gradational transitions between sedimentary facies, and overlapping contact relations between deposits in new subsurface data (Lapham and Maevsky, 1990) are used to infer the extent of the glacial lakes, including glacial Lake Winimussett. Recessional ice-margin positions required to hold in various lakes (the "requisite ice barriers" of Alden, 1924) are correlated. This allows us to deduce the regional trend of successive, topographically controlled ice-margin positions of valley ice lobes and larger lobes of the Laurentide ice sheet (Stone and Peper, 1982). This allostratigraphic method contrasts with previous analyses of morphologic features and attendant speculation on the size of stagnant ice blocks and regional stagnation (Flint, 1930). This latter method has been proven inadequate at quadrangle and regional scales (see examples in Larson and Stone, 1982; Cadwell, 1986; Stone and others, 1982, 1992). The trip further compares the deposits and the regime of the stagnant-ice zone in the narrow, south-draining valley of Muddy Brook with deposits in the wider valley of the Ware River and the north-draining valley of Winimusset Brook.

Previous geologic mapping studies of the Ware-Barre area emphasized morphologic features of the deposits. W.C. Alden (1924) concluded that till of a single glaciation formed few drumlins and the thin till sheet on top of bedrock. Alden mapped extensive deposits, composed chiefly of sand and gravel, which he considered to be ice-marginal, terminal moraine deposits that indicated the local trend of the receding ice margin. Alden also mapped valley-side morainal or kame terraces, glacial delta or sand plain deposits east of Gilbertville, sand and gravel deposits of temporary glacial lakes in the valley of Winimusset Brook, eskers in four segments in Muddy Brook valley, and outwash sand and gravel in terrace segments in the Ware and Wheelwright areas. Alden did not identify outflow channels for glacial lakes in the area, and he showed only one ice barrier symbol that marked the probable position of the front of the glacier behind upland deposits south of Barre Plains. Alden's map and report clearly showed his understanding of the regional significance of recessional ice-margin positions defined on the basis of deposits of local temporary lakes. His lack of details in the Ware-Barre area indicates the difficulties of interpreting these upland areas.

J.W. Mulholland (1974) compared the two-till stratigraphy in drumlins and other areas of thick till with similar stratigraphies described by Pessl and Schafer (1968). Like Pessl and Schafer, he concluded that the lower till was the product of a separate pre-late Wisconsin glaciation. He also subdivided meltwater deposits in the Ware quadrangle into kames, kame

deltas, kame terraces, kame moraines, ice-channel deposits, and valley-train deposits. Mulholland concluded that stagnation-zone retreat was the mode of glacial retreat in the area, but he also envisioned that isolated, stagnant-ice tongues as long as five to seven km remained during sedimentation of glaciofluvial deposits in the valleys south of the retreating ice margin (Mulholland, 1982, fig. 1). Mulholland identified two major ice-margin positions that trend northeasterly in the lower Ware valley on the basis of upland kame-terrace deposits. Mulholland correlated deposits in the northern part of the area along two major ice-margin positions that trend east-west.

Previously in the Barre area, ice-wedge cast structures were observed in two exposures (Larsen, 1979). One of these structures is described here, but it is not exposed at present. Another sand pit that exposed three ice-wedge casts was found in the East Brookfield area in August, 1992; this locality is the last stop of the fieldtrip. The ice-wedge casts in many localities support the current understanding of a late, harsh, periglacial climate that lingered in the region subsequent to deglaciation, as emphasized by J.R. Stone and G.M. Ashley (1991; 1992 NEIGC Trip A-7).

## PHYSIOGRAPHY AND BEDROCK GEOLOGY OF THE WARE-BARRE AREA

The Ware-Barre area is in the central uplands of central Massachusetts (fig. 1), locally known as the Worcester County Plateau (Emerson, 1917), which is within the New England Upland section of the New England physiographic province of Fenneman (1938). Accordant hilltops in the uplands range in altitudes from 900 to 1300 ft. Valleys in the area are narrow, from 1 km wide across the valley floor (Muddy Brook valley) to 2 km wide (Ware River valley). Relief from the floors of the valleys to adjacent hilltops is about 400 ft. Upland areas contain numerous drumlins and scattered areas of rock outcrop. Valleys are filled with stratified glacial deposits, and postglacial floodplain and swamp deposits. The Ware River and its tributaries make up the principal drainages of the area. The Ware River flows southwest from the Barre area to Ware, thence further south and west to the Chicopee River which flows into the basin of glacial Lake Hitchcock in the Connecticut River valley (Fig. 1). Muddy Brook is a major south-flowing tributary of the Ware River in the western part of the Ware quadrangle. Winimusset Brook flows north to the Ware River in the eastern part of the quadrangle.

The bedrock geology in the area includes the distinctive rusty weathering, sulfidic schists of central Massachusetts (including the Brimfield Schist of Emerson, 1917), which impart deep iron-manganese staining (Stop 1) to the upper and permeable zones of the surficial deposits. Bedrock units include intrusive igneous rocks, and sedimentary rocks and their metamorphic equivalents of the Merrimack belt of central Massachusetts (Zen and others, 1983). The area lies on a west-dipping, overturned limb of a large antiformal nappe (as shown in section DD' north of Barre Plains, Zen and others, 1983). Rock units strike generally north beneath north-trending strike ridges and valleys. On the west side of the area, resistant Monson Gneiss in the core of the nappe underlies the uplands. The intrusive Hardwick Tonalite underlies the uplands between Muddy Brook valley and the valley of the Ware River. The similarly resistant Coys Hill Porphyritic Granite Gneiss is a quasi-concordant intrusion within the Littleton Formation (Zen and others, 1983), and underlies the distinctive ridge crest northwest of Wheelwright (fig. 2, 5). Schists of the Paxton Formation of Zen and others (1983) are flat-lying in the area east of Barre where the Paxton underlies the eroded uplands. Relatively nonresistant metamorphosed sedimentary rocks, which dip 20° to 25° west, underlie the valleys of Muddy Brook, the Ware River, and Winimussett Brook. These rocks include sulfidic schists of the Partridge Formation (which includes the Brimfield Schist of Emerson, 1917), and schists and minor gneiss of the Littleton Formation. The sulfidic schists of the Partridge Formation contain pyrrhotite, which weathers readily, and pyrite, and some units contain hydrogen sulfide in fluid inclusions, as well as finely disseminated carbon. The sulfidic assemblage probably accumulated in black muddy sediments that formed in anoxic conditions in the Early to Middle Ordovician Iapetus (proto Atlantic) Ocean (P. Robinson, pers. comm., 1992).

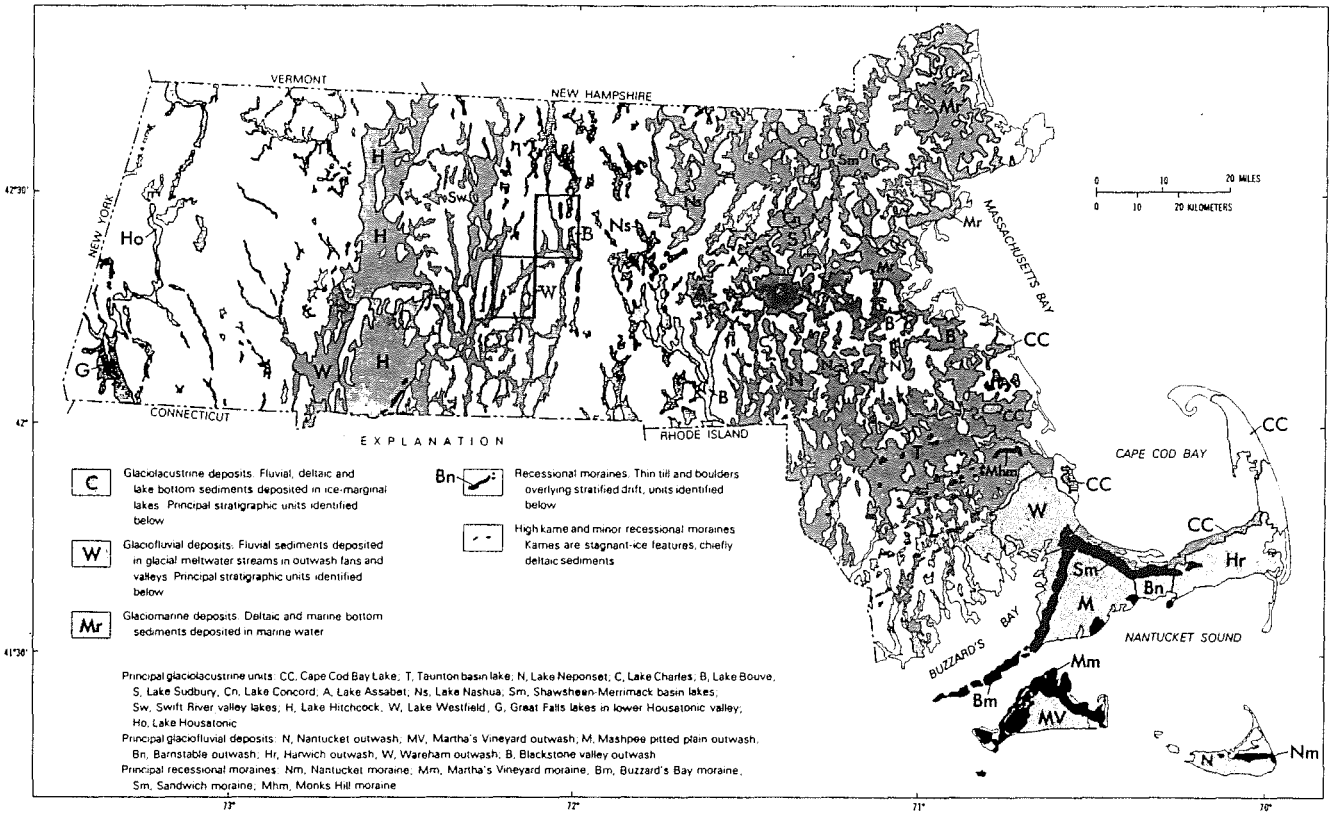



Figure 1. Location map of the Ware-Barre area, central Massachusetts, in relation to the distribution of glacial stratified meltwater deposits (Stone, 1981); **W**, Ware 7.5' quadrangle, **B**, Barre 7.5' quadrangle.

## EROSIONAL EFFECTS OF GLACIATION IN THE WARE-BARRE AREA

Multiple glacial erosion has removed the pre- and interglacial soils and most of the preglacial saprolite (Schafer and Hartshorn, 1965) from the bedrock surface. In some deep road cuts, weathered rock materials are preserved in rock zones characterized by differences in grain size, compositional layering, or fractures, notably in deeply oxidized zones within sulfidic schists. Differential glacial erosion has accentuated the physiographic distinctions between resistant and nonresistant rock units. Erosion of resistant upland rocks has produced aligned ridge crests and tributary valleys controlled by the trend of foliation or local fractures in the concordant intrusive rocks.

The trend of the Coys Hill Porphyritic Granite Gneiss is an excellent example of litho/structural control of topography in the area. Figures 2 and 3 show the relationship of the Coys Hill to the rock-controlled ridge crest topography northwest of Wheelwright. Figure 3 shows about 100 ft of relief along the glacially plucked eastern side of the erosionally resistant intrusive body. Topographic evidence for the trace of a similarly resistant extension of the Coys Hill in the Ware River valley is less clear. Figure 2 shows a buried saddle in the bedrock surface between glacially scoured depressions near the trace of the Coys Hill that is depicted on the bedrock geologic maps (Field, 1976, Zen and others, 1983). Results of shallow seismic investigations do not conclusively confirm the trace of the Coys Hill. Seismic velocities for the shallow upper zone of the presumed schists that underlie the valley-bottom depressions (13,065

to 14,155 ft/sec, Lapham and Maevsky, 1990) are similar to velocities of rock from three seismic refraction lines (115, 103A, 103B, fig. 3) that cross the trace of the Coys Hill (12,964 to 13,770 ft/sec). Line 40, which is above the projected Coys Hill trace, has a shallow rock velocity of only 11,087 ft/sec. The seismic velocities indicate that the upper part of the bedrock beneath the lines does not contain a detectably denser, continuous body of Coys Hill. The Coys Hill may be fractured and partially weathered, and thus may have seismic characteristics similar to those of the enveloping schists. J.D. Peper (pers. comm., 1992) pointed out that Emerson (1917) referred cryptically to a large fault east of Ware, but did not show the fault on his geologic map. Subsequent work (fig. 2; Field, 1976, Lapham and Maevsky, 1990) did not confirm evidence for such a fault. If the west-dipping Coys Hill body and the bedrock saddle are coincident in the subsurface beneath the Ware River valley (fig. 2), a northeast-striking fault in the short basin extension defined by the 500-ft contour could offset the Coys Hill body with an apparent left-lateral displacement through the central part of the saddle. Such a pattern is consistent with subsurface data in figure 2. Alternatively, coincidence of the Coys Hill trace and the bedrock saddle could be related to discontinuity of the Coys Hill, nonconcordant boundaries of the granite gneiss, or a fold offset of the body in the limb of the nappe. Parasitic folds displace the Coys Hill with a sinistral rotation sense in the fold limb (section DD' of Zen and others, 1983), and such a fold in the 300 ft of relief from the ridge crest to the valley bottom could offset the trace of the unit in an easterly direction.

Glacial erosion has produced smooth, U-shaped, bedrock valley cross sections that cut deeply into the strike belts of the gently dipping, nonresistant schists. Glacial scouring and overdeepening of the bedrock surface has produced a deep basin beneath Winimussett Brook valley (fig. 2). The basin coincides with the trace of the Partridge and Littleton schists, and has a minimum elevation of the bedrock surface of 364 ft along seismic line 11 (fig. 3). The basin has more than 180 ft of closure below the 550-ft bedrock-surface contour, as shown by well and seismic data to the south (fig. 3). Similarly, bedrock-surface contours in the 450-ft to 500-ft interval in the Ware River valley west of the Coys Hill (fig. 2) may outline another strike-controlled, scoured basin in the schists. This basin may extend to the narrow valley at Gilbertville where bedrock may be relatively shallow beneath the bouldery alluvium.

Glacial erosion features indicate a unimodal ice-movement direction toward the south-southeast in the Ware-Barre area. Striations in uplands and lowlands trend south-southeasterly (Mulholland, 1974, F.D. Larsen, unpub. data). Steep, glacially plucked rock-outcrop slopes are on the south and southeasterly sides of hills. Glacially smoothed surfaces are on the northwesterly sides of some outcrops.

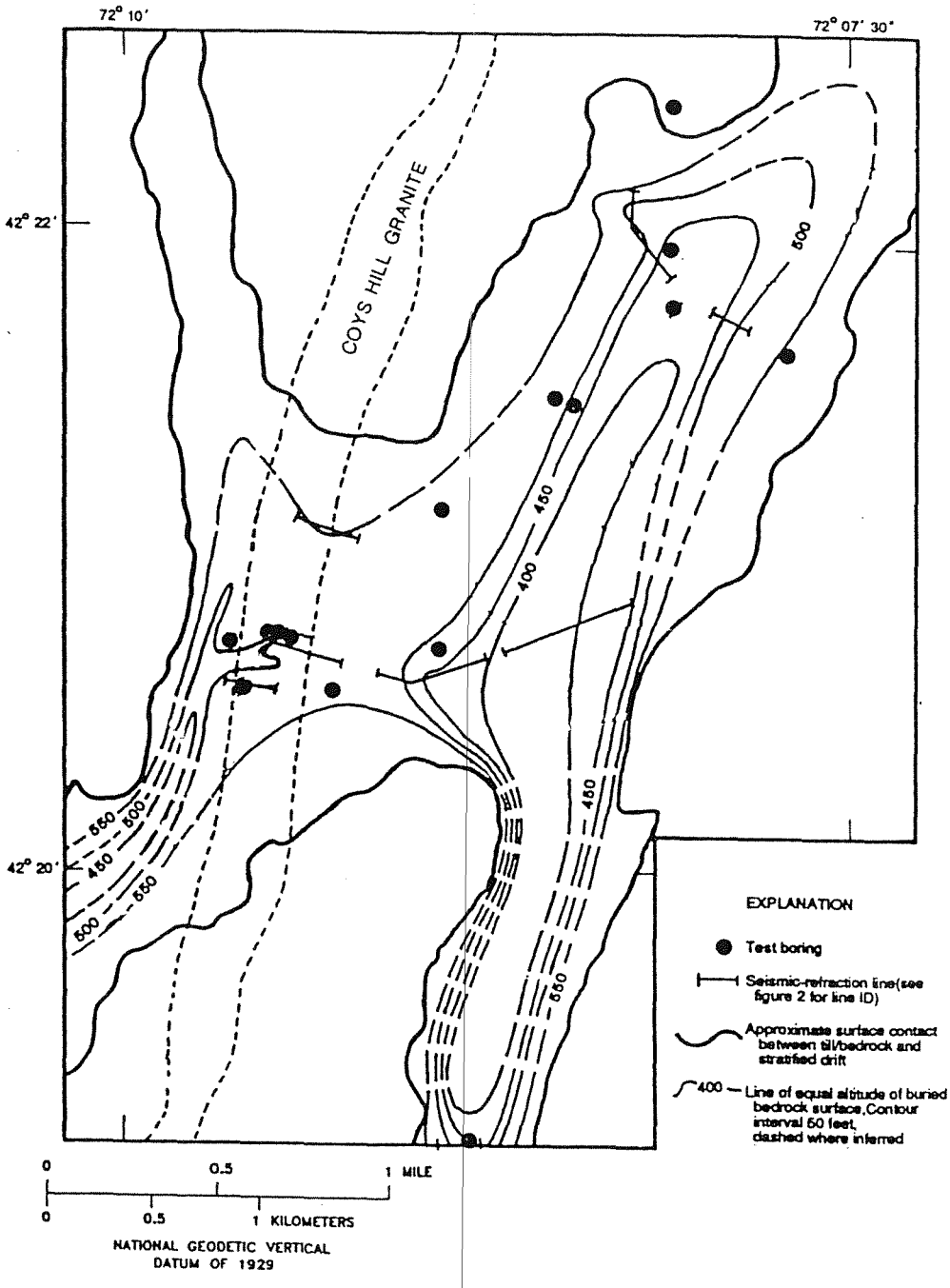



Figure 2. Map of the altitude of the bedrock surface, northeastern quarter of the Ware quadrangle (modified from Lapham and Maevsky, 1990).

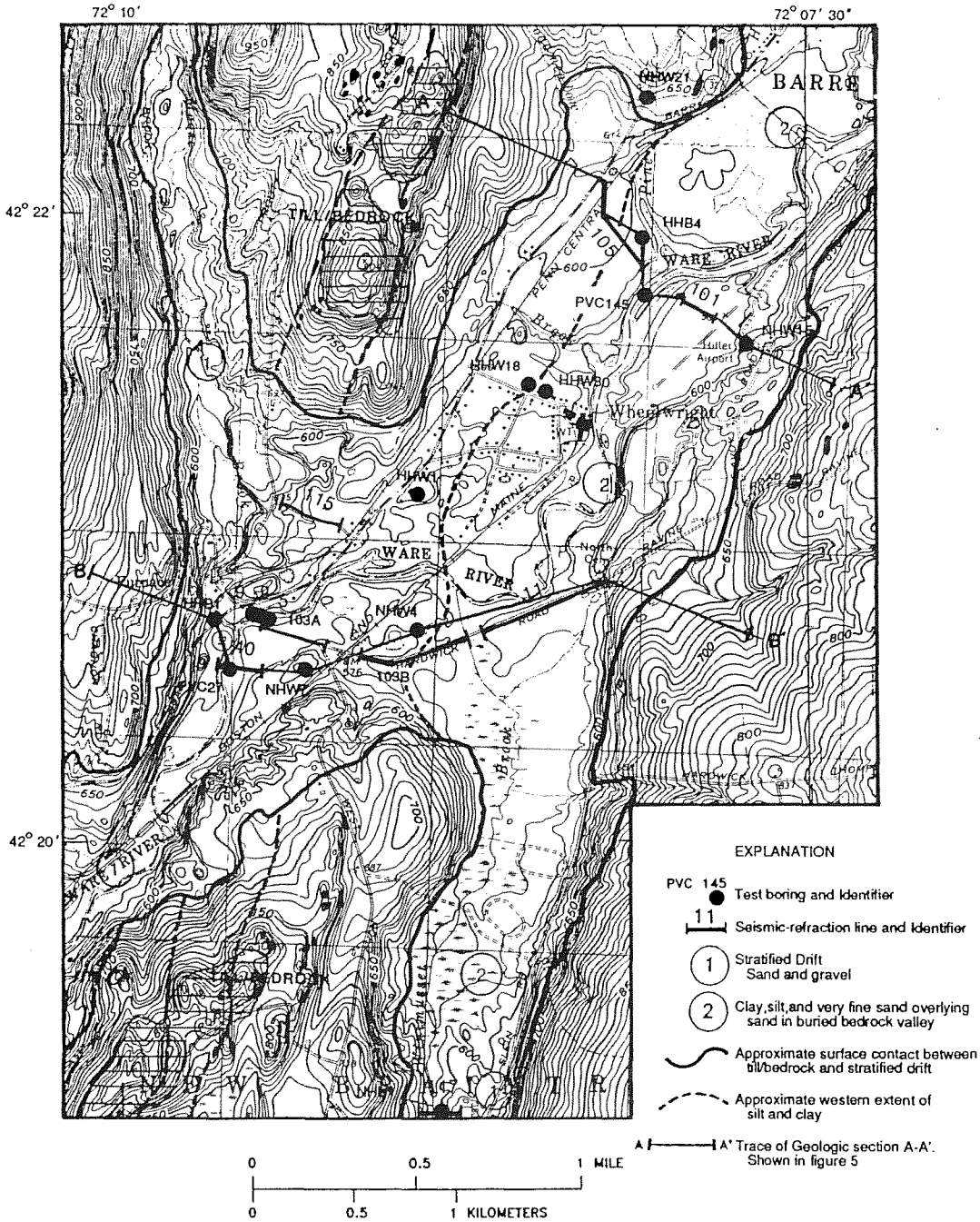



Figure 3. Map showing control of bedrock and subsurface geologic data (figs. 2, 6) in the northeastern quarter of the Ware quadrangle (modified from Field, 1976, Mulholland, 1974, and Lapham and Maevsky, 1990).

## DEPOSITIONAL EFFECTS OF GLACIATION IN THE WARE-BARRE AREA

### Till Deposits

The lower or older till of New England's two tills (Schafer and Hartshorn, 1965, Stone and Borns, 1986, Weddle and others, 1989) underlies drumlins (fig. 4) and other areas of thick till in the Worcester County Plateau (Mulholland, 1974, Stone, 1980, F.D. Larsen, unpub. data). No deep exposures of the older till in drumlins are available presently in the Ware-Barre area, so a general discussion of till stratigraphy is based on limited samples and morphologic relationships. In the Ware-Barre area, the older till is a compact, silty sand variety, derived from the weathered and eroded fine-grained schists (Mulholland, 1974). The till has the characteristic deep weathering zone at the top, and probably is as much as 150 ft to 200 ft thick in some drumlins. Drumlins are located chiefly on uplands in the area, but some thick till deposits that include the older till are present along the sides of valleys. Drumlins occur as individuals or in small groups; drumlin long axes trend exclusively south-southeast in the area (F.D. Larsen and B.D. Stone, unpub. data). The older till presently is correlated with late Illinoian glaciation, about 165 to 150 ka (Stone, in Weddle and others, 1989). The upper till of late Wisconsinan age forms the typically thin mantle over the bedrock surface. This surface till is sandy, locally deeply iron-stained, and was derived from relatively fresh bedrock, the older till in drumlins, and local areas of saprolite. The position of an upland drumlin and generalized stratigraphy of the two late Pleistocene till deposits are shown schematically in figure 4. The schematic relationships indicate that the drumlins and the lower till are integral elements of the present glacially streamlined landscape developed in the upland bedrock surface. We may infer that the land surface on drumlins and on adjacent upland areas of shallow bedrock have been little modified by glacial and nonglacial erosion during the last 150 ka. Likewise, the locations of consequent, litho/structurally controlled, pre-Illinoian river valleys and erosional features of the present landscape have persisted and have been locally overdeepened or accentuated by two episodes of glacial erosion in the last 200,000 yrs.

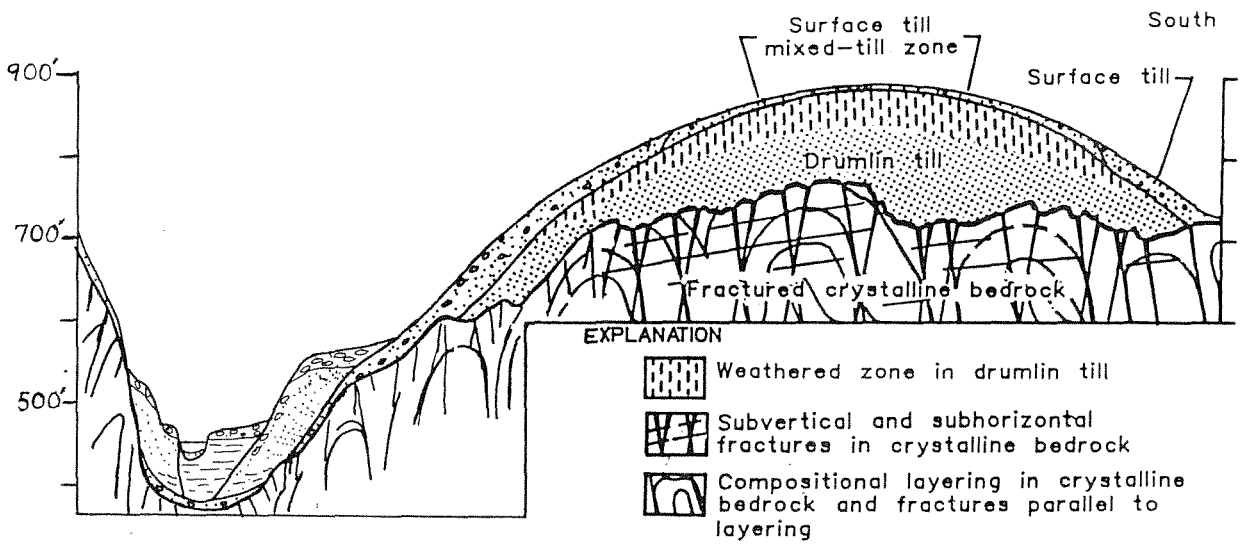



Figure 4. Schematic diagram of till stratigraphy in an upland drumlin in central Massachusetts (based on geologic map data of Mulholland, 1974, Stone, 1980, F.D. Larsen, unpub. data); stratified deposits in valley have symbols explained in Figure 6.

**Meltwater Sediment Deposits**

Meltwater sediments of late Wisconsinan age fill the valleys of the Ware-Barre area, and are scattered in small upland basins north of local drainage divides (fig. 5). The deposits include coarse-grained, sand and gravel deposits, commonly 50-100 ft thick, that underlie various ice-contact morphologic features, and fine-grained glaciolacustrine deposits, as much as 200 ft thick, that fill local valley basins. Classical morphologic units, distinguished by Mulholland (1974), are ideally preserved and include: 1) kames, 2) kame deltas, which are flat-topped and consist of fine sand, silt, and clay bottomset strata, overlain by fine-to-coarse sand and gravelly foreset strata, which in turn are overlain by poorly sorted cobble gravel and sand, 3) kame terraces, which are flat-topped and were inferred by Mulholland to have been formed of stream deposits, 4) kame moraine, which consists of very poorly sorted and collapsed glacial-stream deposits containing boulders and thin layers of diamict sediments, and which grade southward into finer grained stratified sediments, 5) ice-channel deposits, chiefly sand and gravel in sharp-crested ridges as long as 1.9 mi, and 6) valley train deposits, which are chiefly pebble-cobble gravel and sand in terraces along the Ware River.

Deep exposures of various sedimentary facies, morphologic and grain size characteristics, and new subsurface data and basin analyses are employed to contrast 1) ice-marginal meltwater deposition in glacial Lake Winimusset in the Ware/Winimusset valley, 2) deposition in a series of sediment-dammed glacial ponds in the Muddy Brook valley, and 3) deposition of thin glaciofluvial outwash following ice-margin retreat and draining of glacial lakes in the Ware River valley. Meltwater sedimentary facies are distinguished on the basis of grain size, sedimentary structures, lateral and vertical gradational changes or boundaries, and comparison with modern glacial-stream and glacial-lake sediments. Glaciofluvial sand and gravel facies are exposed in the upper deposits of Stops 1-7. These sediments are horizontally bedded, alternating strata of gravel and sand, exhibiting a range of maximum lengths of gravel clasts of 8 to 85 cm. At the ice-contact heads of deposits, the largest gravel clasts have lengths of more than 60 cm. In the area, glaciofluvial facies commonly are less than 15 ft thick, but may be as much as 20-30 ft thick at the ice-contact heads of some deltaic deposits. Glaciodeltaic foreset facies are exposed in the lower sections at Stops 1-4, and at Stop 8. These sediments are inclined from 10° to 35°, and consist of interbedded pebble-cobble gravel and coarse sand, or chiefly sand and minor silt. The glaciofluvial facies overlies foreset facies in deltaic deposits; the base of the glaciofluvial facies is the most accurate estimate of the paleowater surface of the glacial lake. Glacial lake-bottom facies are not exposed, but well data show that these sediments are sand, silt, and clay that stratigraphically overlie slightly older ice-marginal foreset strata (sections A-A' and B-B', fig. 6).

Ice-contact and depositional-slope morphologic features, pit exposures, areal distribution of deposits of different grain size, and well data indicate that the kames, kame terraces, and kame deltas in the Ware-Barre area are mostly lacustrine ice-contact morphosequences (Koteff and Pessl, 1981). These morphosequences are also known as ice-marginal deltas. In contrast to these ice-marginal deposits, the dipping planar surfaces of the wide valley-bottom terrace in the Wheelwright area, and the grain size and sedimentary structures in the underlying sand and gravel (Stops 5-8) support the interpretation of the terrace as a fluvial, non ice-contact morphosequence. This morphosequence is also known as a valley-train deposit (Mulholland, 1974), and is herein named the Ware outwash deposit.

Exposures in deposits in the Muddy Brook valley reveal that flat-topped landforms contain deltaic sediments (Stops 1, 3). The surface altitudes of these flat-topped deposits rise from 450 ft to 580 ft from south to north in 6.6 mi, yielding a northerly rising gradient of nearly 20 ft/mi. This slope exceeds by a factor of 4 the regional isostatic gradient of 4.7 ft/mi (Koteff and others, 1988) determined from nearby Lake Hitchcock. Such a slope of surface altitudes of deltaic deposits indicates damming of the narrow valley by successive deposits during ice-margin recession, which resulted in a series of sediment-dammed basins rising to the north. The deep exposure in the ice-channel deposit at Stop 3 reveals interbedded pebble-cobble gravel strata with anticlinal structure, collapsed on the sides of the ridge. Comparison of modal and maximum sizes of gravel clasts and the absence of interbedded sand strata or channel-form gravel units indicate that the deposit is not a glaciofluvial facies. The ice-channel deposit is interpreted to be composed of sand and gravel foreset strata that prograded into a narrow glacial pond within the ice channel.

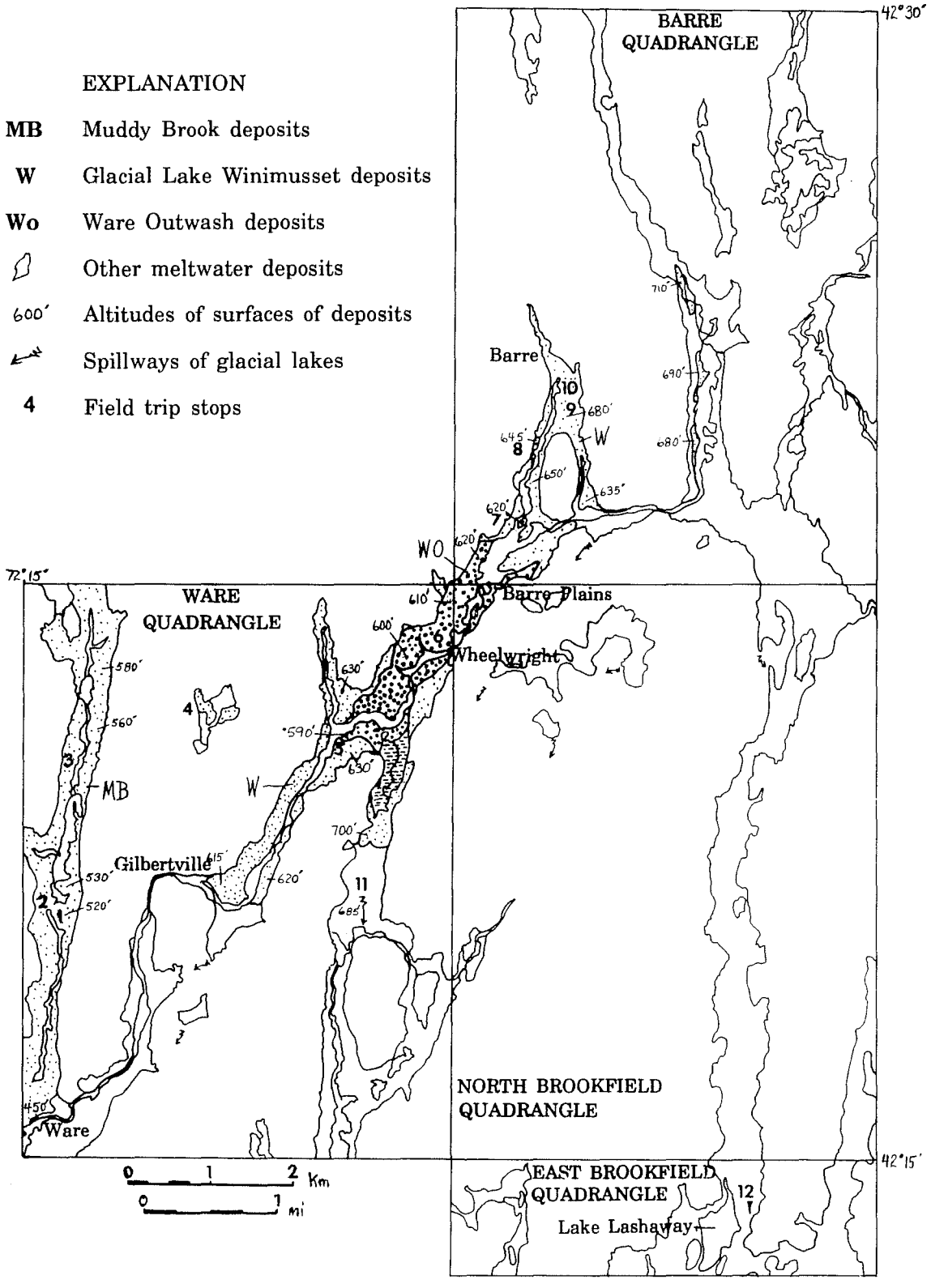



Figure 5. Surficial geologic sketch map of the Ware-Barre area (map data from Mulholland, 1974, F.D. Larsen, unpub. data).

Exposures of foreset strata in the flat-topped kame deltas and kame terraces of the Ware/Winimusset valley (Stops 4, 6) and the stratigraphic relationships of the subsurface lake-bottom sediments reveal that a single level of glacial Lake Winimusset filled the valley to north of the Barre Plains area. Surface altitudes of deltaic deposits rise from 615 ft east of Gilbertville, to 650 ft north of Barre Plains, yielding a gradient of 5-6 ft/mi along the 339° azimuth of isostatic tilt. The spillway for this stage of Lake Winimusset was in the Gilbertville area, over till or rock on the side of valley, or over ice and drift in valley. Limited exposures have not permitted better resolution of the lake water plane; the large delta at Stop 5 may have dammed the valley to the north to a slightly higher lake level. A slightly older, higher stage of Lake Winimusset was controlled by the spillway eroded into older stratified deposits along West Road, New Braintree (Stop 7). The duration of both stages of Lake Winimusset must be consistent with sedimentation rates of reported clay (60 ft thick) and silt (120 ft thick) units that underlie the Ware outwash deposit in the valley south of Wheelwright. The duration remains conjectural, but several decades is not unreasonable.

The Ware outwash deposit is a single unit which is inset into deltaic deposits and overlies deltaic and lake-bottom sediments of Lake Winimusset (Stops 4,5, Fig. 6a,b). The outwash deposit is similar to others that postdate glacial-lake sedimentation in large glacial lake basins (c.f. the Townsend outwash of Koteff and Stone, 1991; the meltwater terrace of Koteff, 1968; the Farmington valley train of Stone and others, 1985). All such terraces have ice-contact or non ice-contact heads, and form relatively thin fluvial deposits that extend across the valley bottoms. The terraces locally are collapsed in kettles. These terraces were deposited by streams that ultimately flowed into lower glacial lakes, but the terrace sediments are mappable only in local reaches of their respective basins.

Ice-marginal morphosequences in the Muddy Brook and Ware-Winimusset valleys and in small upland basins indicate successive ice-margin positions and stagnation-zone retreat across the area. The position of the inferred spillway and drift dam of Lake Winimusset, extensive deltaic deposits in the northeastern part of the lake basin, and the evidence for meltwater deposition of the Ware outwash deposit supports the inferred northeast trend of the receding ice margin. Such a trend would lead to early, rapid sedimentation in the northeastern part of the basin behind a drift dam deposited in the stagnant ice zone. The argument is strengthened by the similar trend of ice-margins in the uplands, and in better constrained basins (fig. 7). The Ware-Barre area was deglaciated by ice that flowed from the Connecticut Valley lobe; the area is west of the interlobate zone along the eastern margin of the Worcester County Plateau (fig. 7; Stone, 1980, Stone and Borns, 1986).

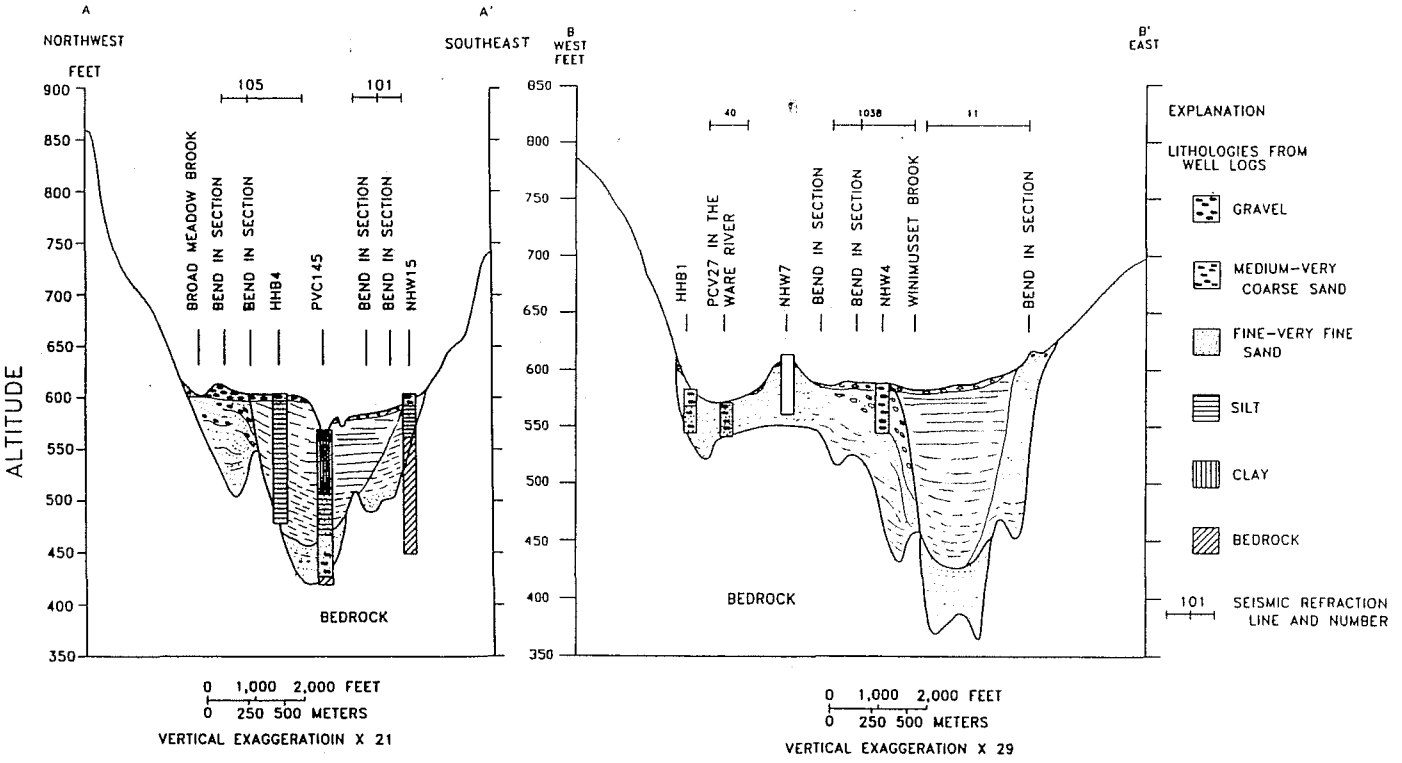



Figure 6. Cross-sections of meltwater sediments and postglacial alluvial and swamp deposits (modified from Lapham and Maevsky, 1990); locations of sections shown in figure 3.

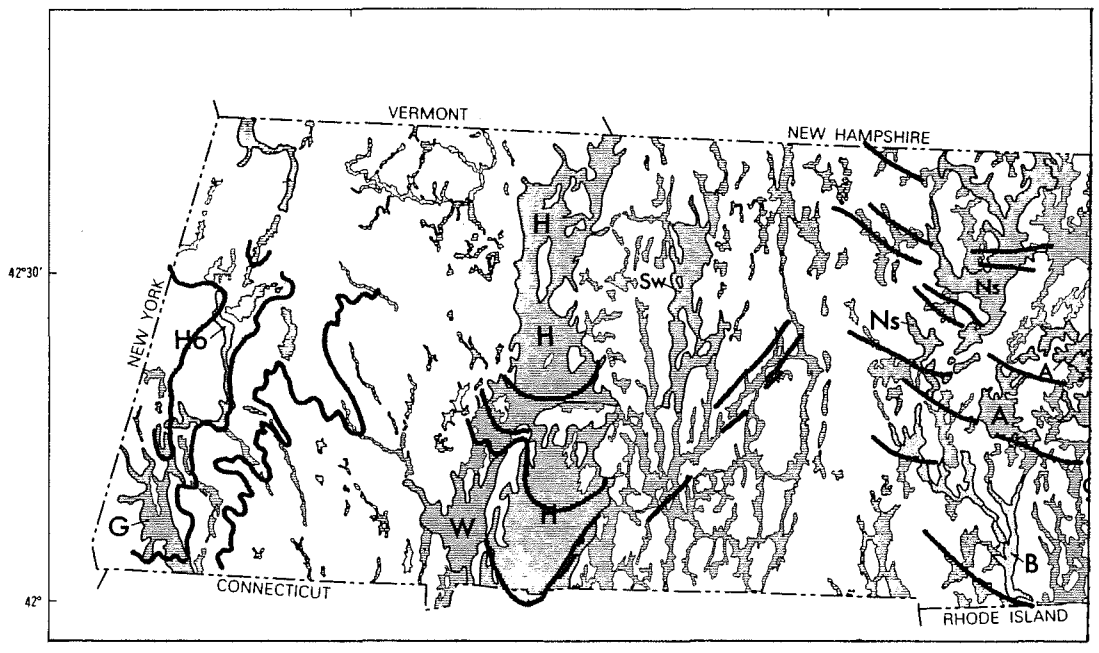



Figure 7. Map of correlated, retreatal ice-margin positions in the Ware-Barre area, in relation to selected ice-margin positions of Peper (1979), Warren and Stone (1986), Larson (1982), and Stone and Peper, 1982)

## ICE-WEDGE CASTS IN THE BARRE AREA

Larsen (1979) reported deep ice-wedge cast structures in stratified sediments in two exposures in the Barre area. These structures are not exposed at present, but the details of the upper part of one of the wedges developed in sand and gravel is shown in figure 8. The wedge is similar to wedge structures described by J.P. Schafer and J.R. Stone (in Stone and Ashley, 1992 NEIGC trip A-7), all of which are interpreted as ice-wedge casts on the basis of criteria of Stone and Ashley (1992) and R.F. Black (1976). The Barre wedge features collapsed and dissaggregated gravel wall beds, and a surface-depression filling of mixed eolian fine sand and coarser sand and gravel at the top and as a plug in the upper meter of the wedge. Another sand pit that exposed three ice-wedge casts was found in the East Brookfield area in August, 1992. This locality (STOP 12) displays the lower parts of three wedges developed in sandy foreset strata. The ice-wedge casts in multiple localities in Massachusetts support the current understanding of a harsh periglacial climate that existed in the region during and after deglaciation.

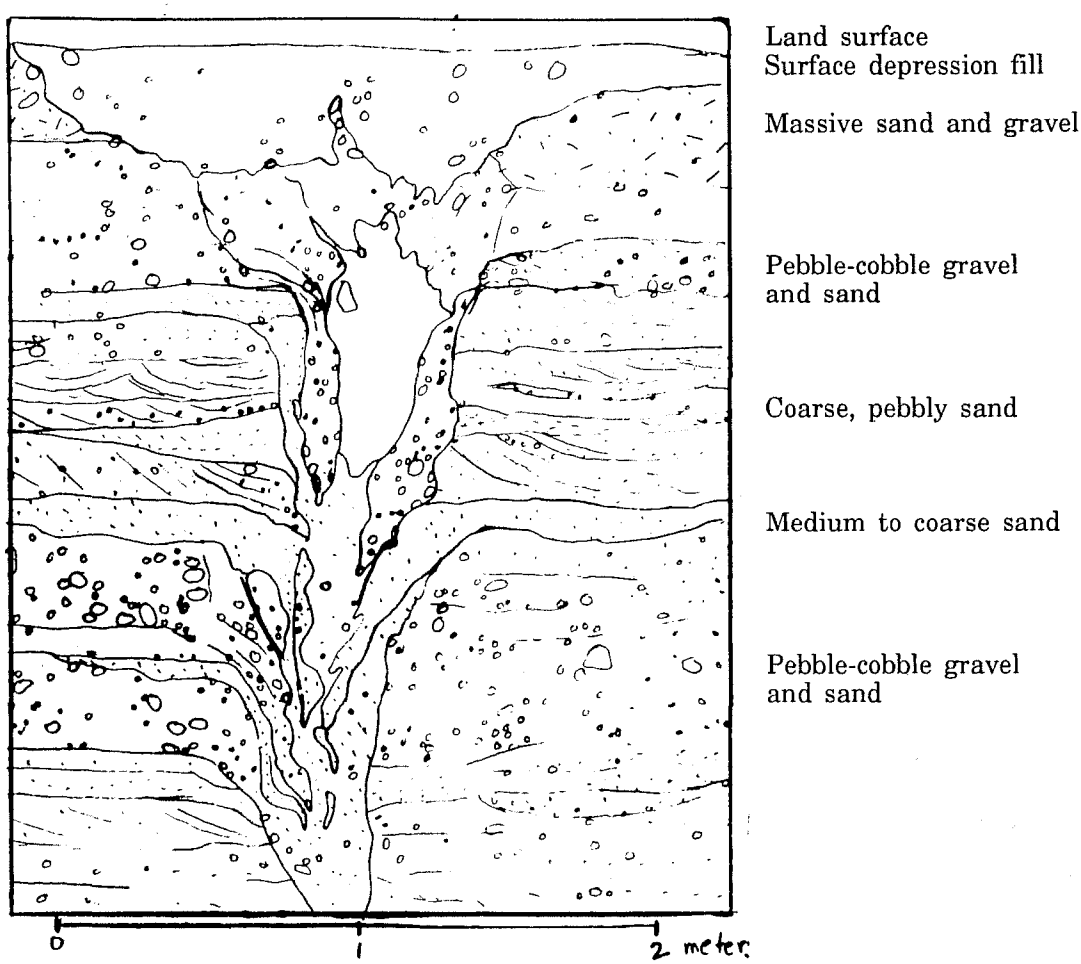



Figure 8. Ice-wedge cast, Barre locality, drawn from a photograph. Vertical scale is 9.5 ft (2.9 m) from land surface to bottom of figure.

## SUMMARY AND HISTORY OF GLACIAL EROSION AND DEPOSITION IN THE WARE-BARRE AREA

The depositional record of early Pleistocene multiple glaciations is not preserved in the Ware-Barre area or elsewhere on the mainland of southern New England. Deposits from perhaps as many as three early Pleistocene glaciations are preserved on Long Island, Georges Bank, and in New Jersey (Stone and Borns, 1986, Stone and others, 1989). These older glacial deposits strongly support the conclusion that the region was not merely eroded by slope and fluvial processes in early Pleistocene periglacial climates, but rather was deeply incised by glacial erosion which removed pre- and interglacial saprolite and perhaps much fresh bedrock.

The erosional effects of the Illinoian and late Wisconsin glaciations are seen in the present landscape. Differential glacial erosion of the bedrock surface by grinding and comminution of nonresistant schists, and plucking of jointed, cleaved, or faulted blocks of all rock units have produced the characteristic steep rock-outcrop areas on the plucked south sides of hills. This is shown by the erosion of the Coys Hill, and the closed basins in nonresistant rock units, such as the schists of central Massachusetts. The coincidence of upland terrains and ridge-crests underlain directly by resistant rock units, and overdeepened valley basins underlain by strike belts of nonresistant units conclusively demonstrate rock lithologic and structural control of the features of the eroded bedrock surface. It therefore seems reasonable to assume that this same lithologic and structural control of erosional features was effective in fluvial and slope erosion episodes in pre- and interglacial intervals. The net effect of late Cenozoic erosion is the overdeepened and U-shaped valleys whose rock floors must be 10's to 100's of meters below their preglacial altitudes. Illinoian drumlins perched on glacially smoothed, fresh-bedrock uplands and in valleys of the present landscape indicate that two episodes of glacial erosion in the last 165 ka have removed significant volumes of weathered and fresh rock but have maintained a streamlined till-bedrock landscape that resisted deep scour of the uplands. In the Ware-Barre area, both glaciations were related to the Hudson-Champlain lobe, and glacial movements were exclusively south-southeasterly.

The late Wisconsinan Laurentide ice sheet entered the central Massachusetts area about 24 ka, flowing from the Hudson-Champlain trough and fanning out toward the southeast. At full glaciation ice related to the eastern side of the Hudson-Champlain lobe extended to eastern Connecticut and western Rhode Island, where it formed an interlobate reentrant near Block Island. The late Wisconsin ice sheet eroded and reshaped drumlins along the interlobate zone in the eastern part of the Worcester County Plateau (Stone, 1980, Stone and Peper, 1982), but striations and drumlins in the Ware-Barre area have a unimodal south-southeast orientation. During deglaciation, the ice margin trended northeasterly, and interacted with the topography to deposit sediments indicative of stagnant-zone retreat. The ice margin dammed small glacial lakes in all northerly draining upland basins. In the largest valley in the area, ice-marginal sedimentation dammed the valley in the narrows near the Gilbertville area, creating the lower stage of glacial Lake Winimusset. Ice-marginal deltaic deposits nearly filled some reaches of the lake basin; lake-bottom sediments accumulated in the widest and deepest basin of the lake, which remained as a wide body of open water. In contrast, successive ice-margin positions in narrow Muddy Brook valley left ice blocks and coarse sediments in deltaic morphosequences that filled and dammed that valley, creating a series of higher glacial pond basins from south to north.

The Laurentide ice sheet retreated across the Ware-Barre area during an interval of perhaps only 200 years, about 16 ka (Stone and Borns, 1986). The periglacial climate that accompanied ice-margin recession produced voluminous meltwater and meltwater sediments which were trapped in glacial lakes and ponds in front of the ice margin (fig. 1). The climate was severe enough to exclude arboreal vegetation during deglaciation of the region (Davis, 1965, Leopold, 1955, Sirkin, 1982, Godreau and Webb, 1989, Davis and Jacobson, 1985, Jacobson and others, 1987). The climate also produced extensive ground ice and true permafrost, the evidence for which includes ice-wedge casts and pingo scars, and extensive eolian deposits. We conclude that a strong seasonality characterized the periglacial climate of the region from 16 ka to about 13 ka. Such a climate effected glacial melting, rapid ice-margin recession, and deposition of voluminous meltwater sediments, while maintaining an average annual temperature below 0° C.

## ACKNOWLEDGEMENTS

Field studies for this trip were conducted in the cooperative geologic mapping program of the U.S. Geological Survey and the Massachusetts Department of Public Works, J.A. Sinnott, Massachusetts State Geologist. Christopher Parisan, University of Lowell, assisted in grain size analyses and Stop descriptions. Mary DiGiacomo-Cohen, Long Island Sound Resource Center, University of Connecticut at Avery Point, Groton, Ct, assisted in preparation of the manuscript. Reviews by J.D. Peper and R.N. Oldale improved the manuscript. Discussions with J.R. Stone and G.M. Ashley contributed to the conclusions of this manuscript.

## ROAD LOG

This trip begins at the University of Massachusetts football stadium parking lot. From the stadium, follow Massachusetts Rt. 116 south to its intersection with Massachusetts Rt. 9 (fig. 9). Turn left and follow Rt. 9 east, through Amherst, Belchertown, past Quabbin Reservoir, and through Ware Center, to the intersection of Rt. 9 and Rt. 32 in Ware. Turn right (south) on Rt. 32, and continue 0.4 mi to McDonald's restaurant on the right. The fieldtrip group will reassemble in McDonald's parking lot.

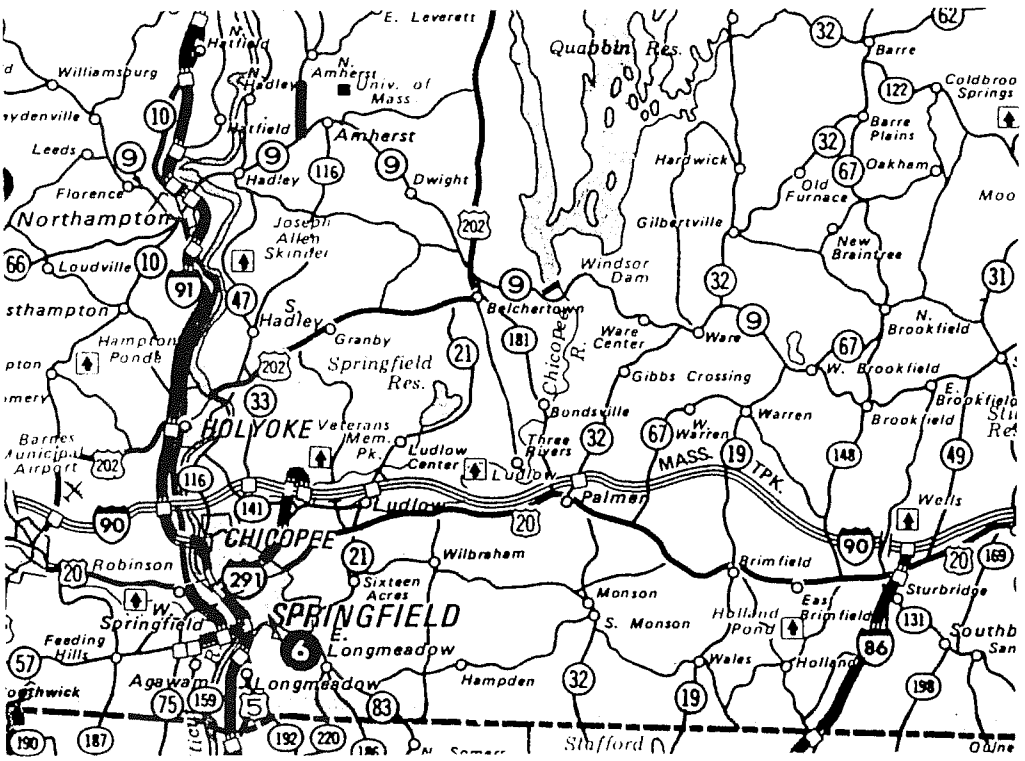



Figure 9. Map of fieldtrip route.

Mileage

0.0 Road log begins at McDonald's Restaurant on Rt. 32, Ware, which is 0.4 mi south of the intersection of routes 9 and 32. Turn left (N) on Rt. 32
0.4 Turn right (E) on Main St.
0.2 Turn left on Pleasant St.
0.6 Bear right (N) on Greenwich Rd.
2.2 Turn right on King St.
0.4 Turn right on Greenwich Rd.
0.1 Turn right on Hardwick Pond Rd.
0.2 Cross Muddy Brook and note relief of gravelly floodplain alluvium.
0.1 Park on side of road.

**STOP 1. GRAVEL PIT, HARDWICK POND ROAD**, on left (Ware quadrangle). This pit was excavated into an ice-marginal deltaic deposit with a flat fluvial plane at altitude of 520 ft (fig. 10). The upper part of the pit exposes sand and gravel glaciofluvial facies, with total thickness of interbedded pebble-cobble gravel and minor sand 1.5-2.0 m thick. The distribution of the 25 largest gravel clasts from the gravel beds has a maximum clast length of 99 cm, and a mean of 44.4 cm. The sand and gravel delta topset strata overlie pebble-cobble gravel and coarse sand foreset beds, which dip southerly to southeasterly. Gravel clasts include scattered rotten schists. Beds of gravel and sand are iron- and manganese-stained. Some beds are lightly cemented by iron, however pores remain open.

Leave pit and retrace route right (W) on Hardwick Pond Rd.

0.3 turn right (N) on Greenwich Rd.
0.1 Park along road.

**STOP 2. GRAVEL PIT, GREENWICH ROAD**, on left (Ware quadrangle). This pit was excavated into an ice-marginal (kame) terrace remnant with upper surface at 560 ft altitude. The upper beds in the pit are coarse gravel glaciofluvial facies. Of 25 boulders, the 17 largest rounded clasts from the upper part of the section, exclusive of 8 angular boulders > 1 m in length, have a maximum length of 94 cm, with a mean of 80 cm. The upper beds were originally horizontally bedded, but are folded and collapsed in different parts of the pit. The upper coarse gravel beds overlie with an erosional angular disconformity dipping gravel and coarse sandy beds. These lower beds appear to be similar to the massive gravel beds above, and could be fluvial beds that are collapsed and eroded syndepositionally by succeeding flows. Closer inspection of associated sandy beds, and pebbly sand beds reveals lateral continuity of these lower strata, which are interpreted as sand and gravel foreset strata.

Leave pit and proceed north on Greenwich Rd.

0.9 Bear right at Y intersection.
0.8 Bear right (E) at 4-way intersection.
1.0 Turn right into landfill.

**STOP 3. HARDWICK LANDFILL, INC.** (Ware quadrangle). The upper pit (fig. 10) was excavated into an ice-marginal ridge with upper surface at 580 ft altitude. Upper beds in the pit are coarse gravel facies with maximum clast length of 48 cm and mean 37 cm. In a small pit north of the road, the maximum clast size is 46 cm, and the mean is 30 cm. The lower pit was cut into an ice-channel ridge, with flat-topped beads at 520 ft to 530 ft altitude. Compare the size of maximum clasts at the top and sides of the ridge with sizes of clasts in the upper pit. Clasts derived from the central part of the ridge have maximum lengths of 41 cm, and average 27.6 cm. Note that the largest clasts are at the base of the ridge on either side. Also note that the thick gravel section is relatively well sorted, is homogeneous texturally, and shows lateral continuity of bedding. It does not contain interbedded coarse gravel or sand beds, and is notably lacking in channel bedforms. The gravel in the ridge is interpreted as sand and gravel foreset strata, deposited subaqueously in a pond environment between large blocks of grounded, stagnant ice. Close inspection reveals that these beds dip south, parallel to the trend of the ridge. The ridge thus is not a tunnel feeder filled with coarse glaciofluvial gravel-to-sand sedimentary sequences deposited under hydrostatic pressure.

Return to entrance to landfill, and turn right on Greenwich Rd.

0.5 Cross Muddy Brook.
1.8 Turn right on Greenwich Rd.
0.3 Hardwick Village, cross Rt. 32.
0.3 Drive slowly while observing landforms.

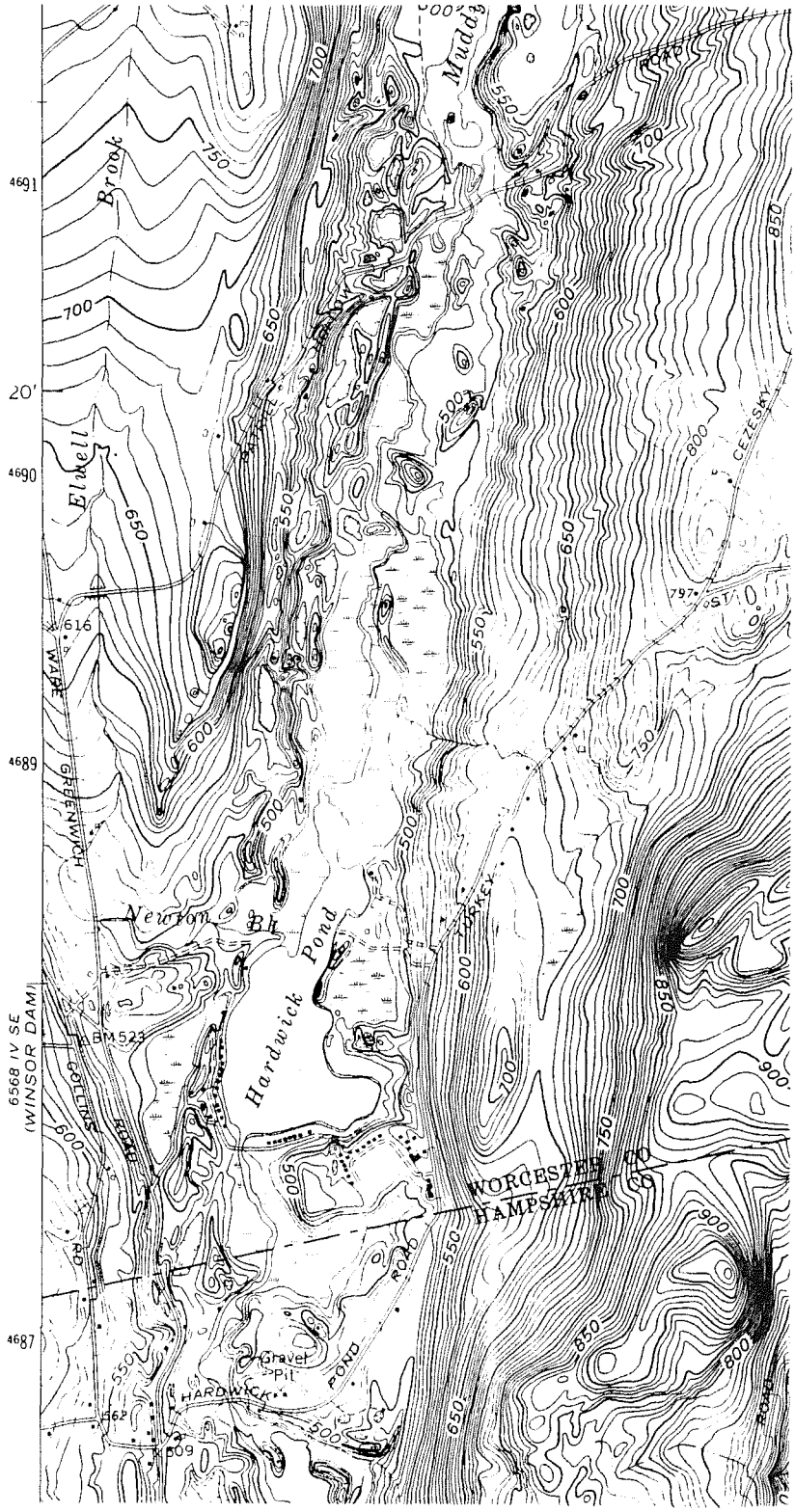



Figure 10. Location of Stops 1,2, and 3; base from Ware 7.5' quadrangle.

**STOP 4. HARDWICK MORAINE** (Ware quadrangle). Moraine deposits of Mulholland (1974); note surface boulders. The section in a small pit on the north side of the road revealed poorly sorted but not compact diamict sediment overlying stratified sand and gravel. The deposits clearly are not flat-topped stratified sediments; the moraine may be a dump moraine, composed of some stratified sediments and flowtills or other sediment-flow deposits.

0.3 Pass Church St.
2.2 Cross Rt. 32.
0.3 Turn right into pit entrance.

**STOP 5. HARDWICK ROAD PIT** (Ware quadrangle). This pit was cut into a flat-topped terrace remnant, with surface altitudes of 590 ft and 580 ft. Fluvial sand and gravel deposits are at the top of the main pit and also at the north end of the pit. In the main pit area, these fluvial beds total less than one meter thick. The maximum gravel clast length is 20 cm, and the mean is 13.2 cm. Outsized angular boulders are scattered in the sand and gravel. In the pit area to the north, the fluvial sediments underlie a slightly lower terrace surface and are about 1.3 m thick. The glaciofluvial sediments are the distal-most deposits of the Ware outwash. Beneath the fluvial outwash cap are sets of sandy foreset strata that dip southerly, and which were deposited in Lake Winimusset. The outsized bouldery foreset beds contain a poorly sorted sand and gravel matrix, possibly deposited as sediment flows from stagnant ice blocks in contact with the deltaic pile.

Return to Hardwick Rd., turn left (N).

0.3 Turn right (NE) on Rt. 32.
1.2 Turn right into pit entrance.

**STOP 6. McDONALD, INC., WASHED STONE AND SAND, Rt. 32** (Ware quadrangle). This pit and the pit to the north expose the pebble-cobble gravel and sand facies of the Ware outwash deposit. Compare the grain size with the gravel at Stop 5. Here, the maximum gravel clast is 17 cm long, and the mean is 12 cm. Follow the pit road east into the Town of Hardwick gravel pit. Note grain size of terrace materials and the relatively good sorting. In this pit, the maximum clast length is 19 cm, and the mean is 14.6 cm.

1.0 Turn right, Rt. 32.
1.9 Cross Ware River.
0.4 Turn left at Village Market
1.0 Turn left into pit entrance.

**STOP 7. SYKES PIT, SOUTH BARRE ROAD.** (Barre quadrangle). This pit was cut into an erosional remnant of the Ware outwash deposit at altitude 620 ft in the center of the valley. The fluvial terrace sand and gravel beds are generally less than 1 m thick in the area of the pit. The underlying lower foreset strata of Lake Winimusset are sandy and contain depositional sequences of climbing ripples and draped laminations. The maximum surface altitudes of the ice-contact deltaic deposits in this part of the valley are over 640 ft to 650 ft, more than 20 ft above the altitude of the terrace sand and gravel.

Return to pit entrance, turn left (N) on Rt. 32.

1.3 Drive slowly, and observe excavation on left.

**STOP 8 (ALTERNATE). EXCAVATION INTO WARE OUTWASH DEPOSITS, RT.32** (Barre quadrangle). This pit is excavated into terrace sediments embanked against the till hillslope at an altitude of about 645 ft. The original surface of the terrace was about 15 ft to 20 ft below the surface of the deltaic deposit that underlies the cemetery to the north. In August, 1992, the terrace fluvial sand and gravel facies showed south-dipping trough crossbeds interbedded with layered pebble-cobble gravel. The maximum clast length is 21 cm, and the mean is 15.9 cm.

Proceed north on Rt. 32.
0.1 Straight at intersection.
0.4 Turn right on Worcester Rd.
0.2 Turn left on Walnut Hill Rd.
0.1 Park on side of road.

**STOP 9. WALNUT HILL ROAD PIT** (Barre quadrangle). This pit was cut into the 680 ft flat surface which underlies the cemetery and High School football field. The pit exposes a section of the Lake Winimusset deltaic deposit. Fluvial beds at the top of the pit have a maximum clast length of 35 cm, and a mean of 23.8 cm. The deposit is little collapsed. It originally extended up valley to altitudes of 710 ft, and probably derived from meltwater that issued from the ice margin in the upper reaches of the drainage basin. The terrace is thus an non-ice-contact morphosequence.

Return to road.

0.2 Park on side of road.

**STOP 10 (ALTERNATE). UPPER WALNUT HILL ROAD PIT** (Barre quadrangle). This pit exposes the mined-out remains of the head of the glacial Lake Winnemusset sediments, the northern-most ice-contact delta with surface altitude of 710 ft. The maximum gravel clast has a length of 35 cm, and a mean of 24.4, and thus indicates that the sediments derived from an outwash head up valley.

Return to cars.

Retrace route south on Rt. 32 to Hardwick Rd.

7.4 Turn left on Hardwick Rd.
0.4 Bear right on West Rd.
2.8 Park on side of road.

**STOP 11. SPILLWAY FOR HIGH STAGE LAKE WINEMUSSET, WEST ROAD** (Ware quadrangle). The spillway channel in the cleared field is 1400 ft long, and 100 ft wide at its narrowest part. It is cut into the head of an ice-marginal deltaic deposit (the New Braintree kame moraine of Mulholland, 1974), as revealed in a pit east of the spillway channel. The deposit is part of a system of such deposits in the Mill Brook valley which drains to the south. These deposits rise to altitudes of 710 ft at this locality, where Mulholland observed numerous boulders and lenses of diamicton or probable flowtills in the glaciofluvial facies at the top of the section.

Continue south on West Rd.

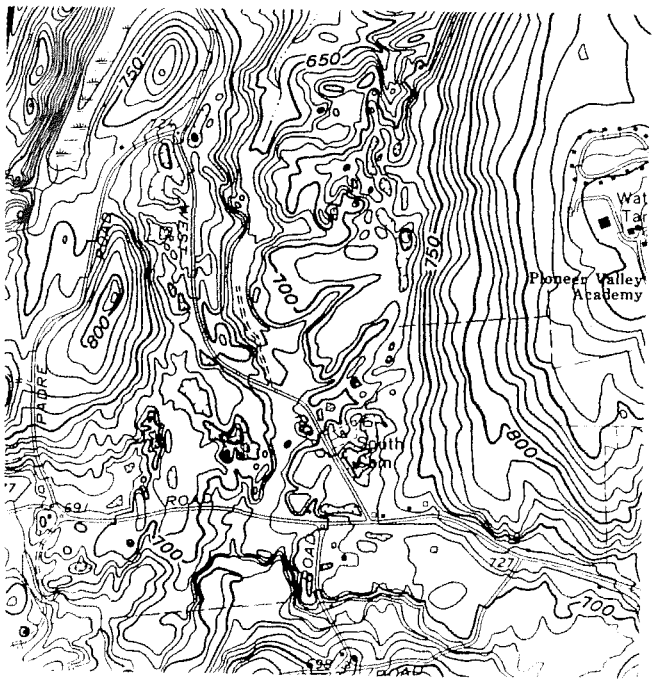



**Figure 11. Location of Stop 9 (alternate).**

0.4 Turn left (E) on Gilbertville Rd.
0.7 Turn right (S) on West Brookfield Rd.
0.6 Bear left (E) on Prouty Rd.
0.6 Bear right (S).
0.2 Turn left (E) on Waite Corner Rd.
1.3 Drive through a swarm of drumlins.
0.6 Turn left (E) on Bigelow Rd.
0.9 Turn right (S) on Main St., North Brookfield.
0.6 Turn left (E) on Ward St.
1.8 Turn right (S) on Green Rd.
1.4 Turn left (E) on Harrington St.
0.6 Enter pit on left.

**STOP 12. HARRINGTON STREET PIT, LAKE LASHAWAY,** (East Brookfield quadrangle). This pit was cut into deltaic deposits with surface altitudes of 640 ft to 650 ft. These deposits were part of the system of sediment-dammed deposits in the Brookfield River valley (fig. 4). The pit exposes coarse pebbly sand and pebble gravel foreset strata in steeply dipping sets, and fine-to-medium sand foreset strata in gently dipping sets. Handlevelling to the original land surface on the edge of the pit indicates that 1 to 2 m of topset sand and gravel deposits have been removed above the active pit face. Three wedge structures with divergent strikes deform the foreset strata in the pit face. Note that they are vertical structures, and compare the geometry of these wedges in sand with the geometry of wedges in near surface sand and gravel deposits (fig. 8).

## END OF TRIP

To return to Massachusetts Turnpike (fig. 9), drive south on Harrington St.; turn left (E) on Route 9. Turn right (S) on Route 49; turn right (W) on Rt. 20. Continue westbound on Rt. 20 to entrance to I-84. Go north to the Massachusetts Turnpike, or go south on I-84.

## REFERENCES CITED

Alden, W.C., 1924, The physical features of central Massachusetts, U.S Geological Survey Bulletin 760-B, 105 p.

Black, R.F., 1976, Periglacial features indicative of permafrost: Quaternary Research, v. 6, p. 3-26.

Cadwell, D.H., ed., 1986, The Wisconsinan stage of the First Geological District, eastern New York; New York State Museum Bulletin Number 455, 192 p.

Davis, M.B., 1965, Phytogeography and palynology of northeastern United States, in Wright H.E. Jr., and Frey, D.G. eds., The Quaternary of the United States: Princeton, N.J., Princeton University Press, p. 377-401.

Davis, R.B. and Jacobson, G.L., 1985, Late glacial and Early Holocene landscapes in northern New England and adjacent areas of Canada: Quaternary Research, v. 23, p. 341-368.

Emerson, B.K., 1917, Geology of Massachusetts and Rhode Island: U.S. Geological Survey Bulletin 597, 289 p.

Fenneman, N.M., 1938, Physiography of the Eastern United States, McGraw-Hill, New York, 534 p.

Field, M.T. 1976, Bedrock geologic map of the Ware area, central Massachusetts: U.S. Geological Survey Open-file Report 76-708, 2 plates.

Flint, R.F., 1930, The glacial geology of Connecticut; Connecticut State Geological and Natural History Survey Bulletin 47, 294 p.

Gaudreau, D.C. and Webb, T., 1985, Late-Quaternary pollen stratigraphy and isochrone maps for the northeastern United States: in Pollen Records of Late-Quaternary North American Sediments, A.A.S.P. Foundation, p. 247-279.

Jacobson, G.L., Jr., Webb, T. III, and Grimm, E.C., 1987, Patterns and rates of vegetation change during the deglaciation of eastern North America; in North America and adjacent oceans during the last deglaciation, Vol. K-3 Geological Society of America, p. 277-288.

Koteff, Carl, 1966, Surficial geologic map of the Clinton quadrangle, Worcester County, Massachusetts; U.S. Geological Survey Geologic Quadrangle Map GQ-567, scale 1:24,000.

Koteff, Carl, Stone, J.R., Larsen, F.D., Ashley, G.M., Boothroyd, J.C., and Dincauze, D.F., 1988, Glacial Lake Hitchcock, Postglacial uplift, and post-lake archaeology, in Brigham-Grette, J. ed., Fieldtrip Guidebook AMQUA 1988, p. 169-208.

Koteff, Carl, and Pessl, Fred, Jr., 1981, Systematic ice retreat in New England: U.S. Geological Survey Professional Paper 1179, 20 p.

Koteff, Carl, and Stone, B.D., 1991, Surficial Geologic Map of the Townsend quadrangle, Massachusetts: U.S. Geological Survey Geologic Quadrangle Map, scale 1:24,000.

Lapham, W.W., and Maevsky, Anthony, 1990, Buried bedrock topography, evidence of overdeepening of Winimusset Brook bedrock valley, and reinterpretation of the extent of glacial Lake Winimusset, Ware quadrangle, Massachusetts; Northeastern Geology, v. 13, no. 1, p. 64-73.

Larsen, F.D., 1979, Retreat of ice left eastern upland of Massachusetts subject to deep freeze; U.S. Geological Survey Prof. Paper 1150, p.58.

Larsen, F.D., and Hartshorn, J.H, 1982, Deglaciation of the southern portion of the Connecticut Valley of Massachusetts, in Larsen, G.J. and Stone, B.D., eds., Late Wisconsinan glaciation of New England: Dubuque, Iowa, Kendall/Hunt Publishing Company, p. 115-128.

Larson, G.J., and Stone, B.D., eds., 1982, Late Wisconsian glaciation of New England; Dubuque, Iowa, Kendall/Hunt Publishing Company, 242 p.

Leopold, E.B., 1955, Climate and vegetation changes during an interstadial period in southern New England: Unpublished PhD dissertation, Yale University, New Haven, Connecticut, 99 p.

Mulholland, J.W., 1974, Surficial geology of the Ware quadrangle, Worcester and Hampshire Counties, Massachusetts: Unpublished Ph.D. dissertation, University of Massachusetts, Amherst, Massachusetts, 239 p.

Mulholland, J.W., 1982, Glacial stagnation-zone retreat in New England: Bedrock control; Geology, v. 10, p. 567-571.

Peper, J.D., 1979, Surficial Geologic Map of the Palmer quadrangle, Massachusetts; U.S. Geological Survey Geologic Quadrangle Map, scale 1:24,000.

Pessl, Fred, Jr., and Schafer, J.P., 1968, Two-till problem in Naugatuck-Torrington area, western Connecticut, Trip B-1, in Orville, P.M, ed., Guidebook for field trips in Connecticut: New England Intercollegiate Geological Conference, 60th Annual Meeting, Connecticut State Geological and Natural History Survey, Guidebook 1, 25 p. (paged separately).

Schafer, J.P. and Hartshorn, J.H., 1965, The Quaternary of New England: in H.E. Wright, Jr. and Frey, D.G., eds. The Quaternary of the United States, Princeton University Press, p. 113-128.

Sirkin, L., 1982, Wisconsinan glaciation of Long Island, New York, to Block Island, in Larsen, G.J. and Stone, B.D., eds., Late Wisconsinan glaciation of New England: Dubuque, Iowa, Kendall/Hunt Publishing Company, p. 35-59.

Stone, B.D, 1980, Surficial Geologic map of the Worcester North and part of the Paxton Quadrangle, Massachusetts, U.S. Geological Survey Miscellaneous Investigations Map I-1434, scale 1:24,000.

Stone, B.D., 1981, The Massachusetts State surficial geologic map, in Farquahar, O.C., ed., Geotechnology in Massachusetts, University of Massachusetts, Amherst, Massachusetts.

Stone, B.D., and Peper, J.D., 1982, Topographic control of the deglaciation of eastern Massachusetts: ice lobation and the marine incursion: in Larsen, G.J. and Stone, B.D., eds., Late Wisconsinan glaciation of New England: Dubuque, Iowa, Kendall/Hunt Publishing Company, p. 145-166.

Stone, B.D., and Borns, H.W., 1986, Pleistocene glacial and interglacial stratigraphy of New England, Long Island, and adjacent Georges Bank and Gulf of Maine: in Sibrava, Vladimir, Bowen D.Q., and Richmond, G.M., eds., Quaternary Glaciations in the Northern Hemisphere: Quaternary Science Reviews, v. 5, p. 39-52.

Stone, B.D., Stanford, S.D., and Witte, R.S, 1989, The surficial geologic map of northern New Jersey [abs]; Geological Society of America Program with Abstracts, v. 21, no. 2, p. 69.

Stone, J.R., Ashley, G.M., 1992, Ice-wedge casts, pingo scars, and the drainage of glacial Lake Hitchcock, Trip A-7 in this volume.

Stone, J.R., Ashley, G.M. and Peteet, D.M., 1991, Cross section through a post lake Hitchcock surface depression: New $^{14}C$ dates and evidence for periglacial origin: Geological Society of America, Abstracts with Programs, v. 23, p. 135.

Stone, J.R., Schafer, J.P., and London, E.H., 1982, The surficial geologic maps of Connecticut illustrated by a field trip in central Connecticut; in Guidebook for fieldtrips, New England Intercollegiate Geological Conference 74th Annual Meeting, Storrs, CT, p. 5-30.

Stone, J.R., Stone, B.D., and Lewis, R.S., 1985, Late Quaternary deposits of the southern Quinnipiac-Farmington lowland and Long Island Sound Basin: their place in a regional stratigraphic framework; in Tracy, R.J., ed., Guidebook for fieldtrips in Connecticut and adjacent areas of New York and Rhode Island, New England Intercollegiate Geological Conference 77th Annual Meeting, New Haven; State Geological and Natural History Survey of Connecticut, Guidebook No. 6, p. 535-576.

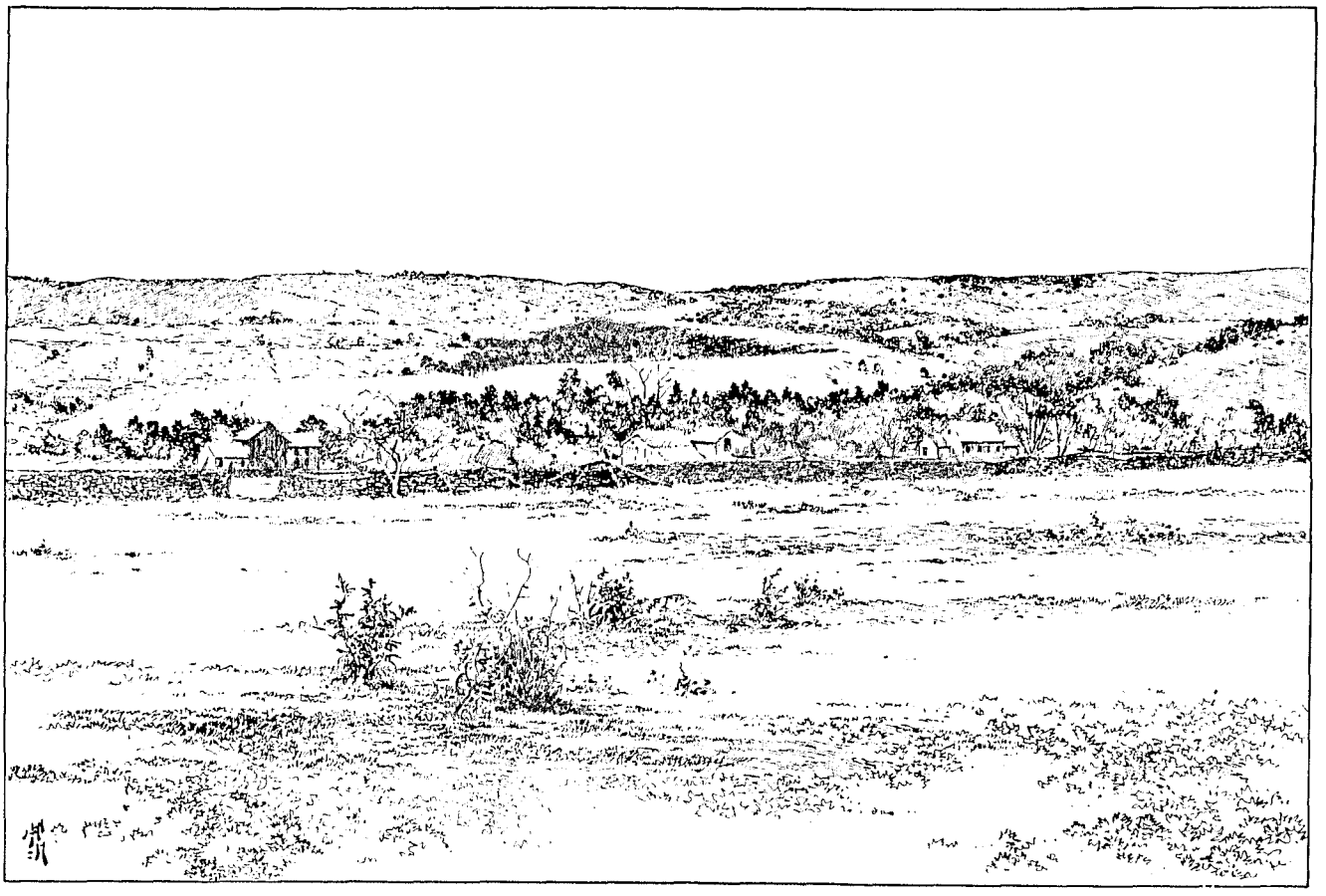
Stone, J.R., Schafer, J.P., London, E.H., and Thompson, W.B., 1992, Surficial Materials Map of Connecticut, U.S. Geological Survey Special Map; scale 1:125,000.

Warren, C.R., and Stone, B.D., 1986, Deglaciation stratigraphy, mode, and timing of the eastern flank of the Hudson-Champlain lobe in western Massachusetts, in Cadwell, D.H., ed., The Wisconsinan stage of the First Geological District, eastern New York; New York State Museum Bulletin Number 455, p. 168-192.

Weddle, T.K., Stone, B.D., Thompson, W.B., Ratelle, M.J., Caldwell, D.W., and Clinch, J.M., 1989, Illinoian and late Wisconsinan tills in eastern New England: a transect from northeastern Massachusetts to west-central Maine, with a section on till stratigraphy of southern New England by B.D. Stone, in Berry, A.W. Jr., ed., New England Intercollegiate Geological Conference, 81 st Annual Meeting, Guidebook for fieldtrips in southern and west-central Maine: Farmington, Maine, University of Maine, p. 25-85.

Zen, E-an, ed., Goldsmith, Richard, Ratcliffe, N.L., Robinson, Peter, and Stanley, R.S., 1983, Bedrock geologic map of Massachusetts, U.S. Geological Survey Special Map, scale 1:250,000.





THE GREAT SERPENT ESKER IN PELHAM; LOOKING NORTH.

# STRATIGRAPHY AND PALEOECOLOGY OF THE DEERFIELD RIFT BASIN (TRIASSIC-JURASSIC, NEWARK SUPERGROUP), MASSACHUSETTS

by
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## INTRODUCTION

Triassic and Jurassic strata of the Deerfield basin comprise a classic area for geological and paleontological studies. The purpose of this guide is to provide an overview and some new details on the stratigraphy and paleoecology of the basin.

## GEOLOGICAL SETTING

Early Mesozoic lithospheric extension led to the formation of a long series of rift basins (Figure 1) extending from Greenland and Spitzbergen to the Gulf of Mexico and Morocco. The Deerfield basin is the erosional remnant of one of these rifts, the exposed North American contingent of which are termed the Newark Supergroup (Figure 2). The Deerfield basin is connected to the Hartford basin of Connecticut and Massachusetts, and constitutes the northernmost exposures of the Newark Supergroup in the United States.

In cross section, the Deerfield basin has the half graben shape characteristic of most rifts (Schlische, 1990). There is a large west-dipping master fault zone on the east side of the basin towards which most of the basin strata tilt, and there is some evidence that there are a series of fault-bounded steps, or rider blocks in basement adjacent to the main border fault (Emerson, 1898; Bain, 1932). This fault zone parallels the older Paleozoic structural fabric and probably reactivated older structures (MacFadyen, *et al.*, 1981). Dips appear steeper in the north and in older strata in the basin suggesting that the half graben shape developed syndepositionally as the result of differential subsidence along the border fault zone, as in other Newark Supergroup basins (Schlische, 1989). In longitudinal section, the Deerfield basin is an asymmetrical syncline, with the northern limb being steeper. Therefore, the thickest part of the section is preserved along the axis of the syncline in the northern third of the basin.

Strata of the Deerfield basin (Figure3) consitute two major genetic sequences: a lower, Late-Triassic age fluvial and alluvial arkose; and an upper, Early Jurassic age lacustrine and alluvial siltstone to conglomerate, with an interbedded basalt low in the sequence. These are divided into five major units, four of which have been formally named as formations (Robinson and Luttrell, 1985).

The basal unit, the Sugarloaf Arkose (~2000 m), is primarily a red pebbly to conglomeratic fluvial and alluvial arkose. It is overlain by a relatively thin (~50 m) sequence of gray and red lacustrine sandstone, siltstone, and pebbly sandstone that we informally term the Fall River beds (Figures 3 and 4). The Triassic-Jurassic boundary presumably falls somewhere near the transition between the Fall River beds and the Sugarloaf Arkose. The 80 m Deerfield basalt which follows, is a tholeiitic lava flow of the high-titanium quartz-normative type (Tollo, in Olsen, *et al.*, 1989). The strata above the Deerfield Basalt comprise a cyclical sequence of red to black siltstone to conglomerate termed the Turners Falls Sandstone and Mt. Toby Conglomerate (or Turners Falls and Mt. Toby formations - Robinson and Lutrell, 1985). These strata, however, mostly represent penecontemporaneously deposited facies of fine-grained fluvial and fluvial clastics (Turners Falls Sandstone) that laterally interfinger with coarse alluvial arkose and conglomerate (Mount Toby Formation). The combined thickness of these formations is variable, depending on location within the basin, but locally may reach more than 2 km (Figure 3).

## SUMMARY OF GEOLOGICAL HISTORY

A succession of accretion events affected Eastern North America during the Paleozoic, ulimately culminating in the condensation of Pangaea. The zone between the adjacent cratons was highly structured by compression and transpression. Many brittle and ductile structures were reactivated as major normal and transtensional strike-slip faults during the onset of regional NW-SE extension (Swanson, 1986) somewhere near the beginning of the Late Triassic (~230 MA). Subsidence of the hanging wall and elevation of the footwall of these reactivated faults created a very large series of half graben in the rifting zone, one of which was the Deerfield basin.

There is presently no direct date for the onset of sedimentation in the Deerfield basin. However, based on lithological correlation with the New Haven Arkose of the Hartford basin, the older exposed strata of the Sugarloaf Arkose can be expected to approximate the Carnian-Norian boundary (~220 Ma) (Cornet, 1977a). Because of the

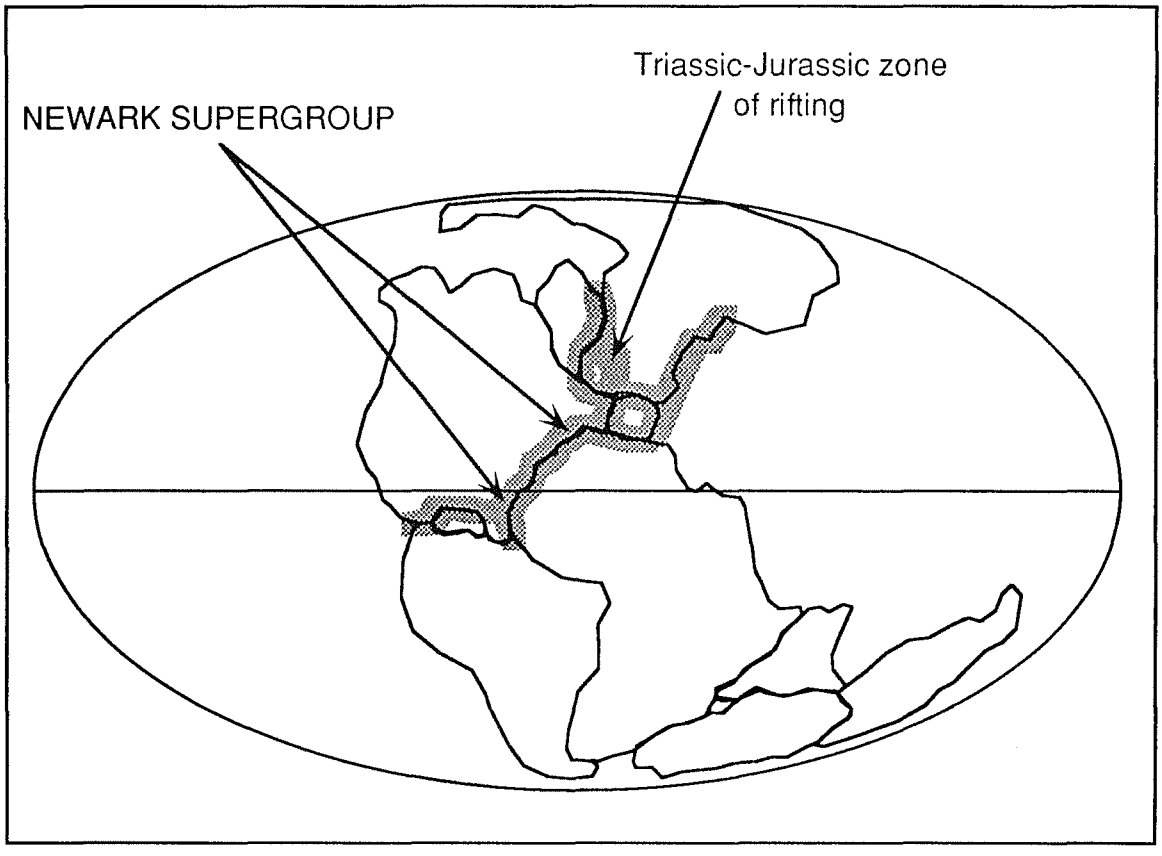



Figure 1. Pangaea during the Carnian Age (~225) of the Late Triassic. Continental positions are based on new paleomagnetic results from the Newark basin (Witte, et al., 1991; Olsen and Kent, 1990). Note that the equator lay at about the present position of Virginia during the Carnian. This is the time during which deposition probably began in the Deerfield basin.

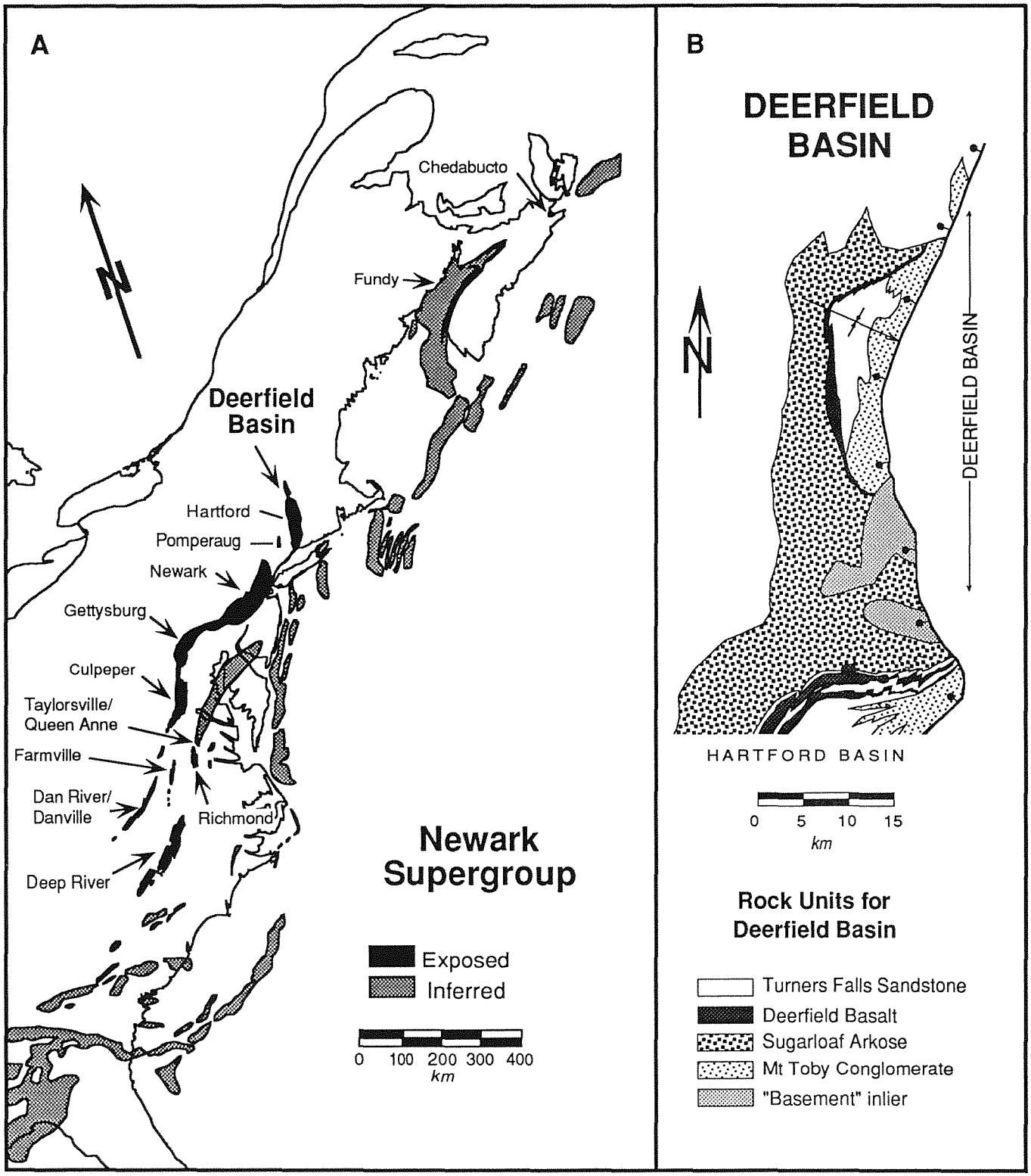



Figure 2. The Newark Supergroup of eastern North America (A) and the Deerfield Basin of northern Massachusetts (B). Figures modified from and courtesy of R. Schlische. Note that the names for units in (B) apply only to the Deerfield basin; the units for the Hartford basin have separate names.

pattern of hanging wall onlap commonly present in half graben (Schlische, 1992), yet older strata are probably present, deeply buried along the axis of the basin.

The Fall River beds are Early Jurassic in age (Cornet, 1977a), and the transition from fluvial (Sugarloaf Fm.) to lacustrine environments must have occurred very close to the Triassic-Jurassic boundary (202 Ma, Sutter, 1988; Dunning and Hodych, 1990; Hodych and Dunning, 1992), probabaly due to an increase in regional extension rate (Schlische and Olsen, 1989). Shortly thereafter, the basin was filled by the lava lake of the Deerfield basalt. At least 2 km of lacustrine to alluvial strata of the Turners Falls and Mt. Toby formations were deposited on top of the basalt, followed by at least 2 to 3 km of additional strata before deposition ceased (based on organic maturity and fission track data (Pratt, et al, 1988).

During the Carnian Age of the Late Triassic, the Deerfield basin lay at about 4° N latitude (Witte, *et al.*, 1991), near the center of the northern lobe of Pangaea. By the time the last preserved Early Jurassic age strata had been deposited, the basin had drifted north, along with the rest of the North American plate, to about 8° N (based on paleomagnetic results from contemporaneous strata of the Newark basin, D.V. Kent, pers comm, 1992). The basin clearly lay in the tropics, but because the sequence is entirely non-marine, the early Mesozoic altitude of the depositional surface is unknown, although presumably at some elevation above sea level. Thus, the climate, although tropical may have been ameliorated by altitude, much like the East African rifts that enclose lakes Tanganyika and Malawi (Manspiezer, 1988). According to the global climate models of Kutzback and Gallimore (1988), the Deerfield basin should have experienced a strong monsoonal climate, evidence for which is seen in the strongly cyclical lacustrine strata which was modulated by Milankovitch-type cyclical climate changes, controlled by celestial mechanics.

If the history of the Deerfield basin is similar to the Newark basin, most of the tilting of the strata, and hence most deposition, was over by the Middle Jurassic (~175 Ma), during which time eastern North America witnessed a major hydrothermal event. This event produced a strong magnetic overprint as well as resetting K-spars in the igneous rocks and in many areas annealing fission tracks in zircons, sphenes and apatites (Roden and Miller, 1991). This event may have been associated with the beginning of true sea-floor spreading and the production of the first Atlantic oceanic crust. The magnetic overprints of the sedimentary rocks of the Hartford and Deerfield basins appear considerably more complex than that of the Newark basin, however, suggesting additional and younger hydrothermal events, perhaps associated with the intrusion of the near by White Mountain plutons and Early Cretaceous dikes (McEnroe, 1989).

The transition from regional NW-SE extension to NW-SE compression from ridge-push probably occurred somewhere near this time, and there may have been an interval of NE-SW compression or shear (Wise, 1988; DeBoer and Clifton, 1988; Olsen, et al, in press). Because the basin was above sea level, erosion of the stratigraphic section probably began as basin subsidence slowed, at sometime after the Early Jurassic. Although uplift of eastern North America may have occurred during the initial phases of the production of oceanic crust, the post-rift unconformity would have been produced as erosion proceeded towards sea level. By the Early Cretaceous (~130 Ma) a combination of thermal (cooling) subsidence and erosion brought the basin to its present elevation near sea level. The thickness of any coastal plain strata that might have been deposited prior to Neogene sea level drop and erosion must have slight. From at least the Cretaceous to the present, the basin, as well as most of eastern North America, has been under mild regional NW-SE compression.

## PALEONTOLOGY AND PALEOECOLOGY

### Historical Notes

The earliest published observations of fossils from the Deerfield basin were made by Edward Hitchcock in 1818, when he recorded his discovery of "distinct impressions of fish", occurring in a "schistose rock" [gray, micaceous, silty shale] on the east bank of the Connecticut River, north of Sunderland. Hitchcock, a clergyman, geologist, educator, and later professor and president of Amherst College and State Geologist of Massachusetts, was an astute observer and an assiduous collector. Hitchcock's remarks on the Sunderland fossils were included in his preliminary description of the geology and mineralogy of the Deerfield basin, appearing in the inaugural volume of Benjamin Silliman's American Journal of Science (Hitchcock, 1818). In subsequent years, Hitchcock became one of the Journal's most frequent contributors, and he and Silliman established a lifelong correspondence and friendship. At Silliman's request, in 1821, Hitchcock employed two men to further excavate the Sunderland site, and in less than half a day they uncovered more than 50 fish and associate plant fragments. Hitchcock forwarded a box containing several Sunderland fishes to Silliman, who in turn sent examples to Alexandre Brongniart in Paris for identification. In an extended account of the geology, mineralogy and scenery of the Connecticut Valley Mesozoic in the sixth volume of the Journal (1823), Hitchcock provides a thorough description of the stratigraphy and fossil occurrences at the Sunderland locality, clearly noting the intimate association of the fossiliferous shale with the coarse conglomeratic units. The diagrams of Sunderland fishes accompanying Hitchcock's report are of significance because they are the earliest graphic renderings of Newark Supergroup fossils and the first illustrations of complete

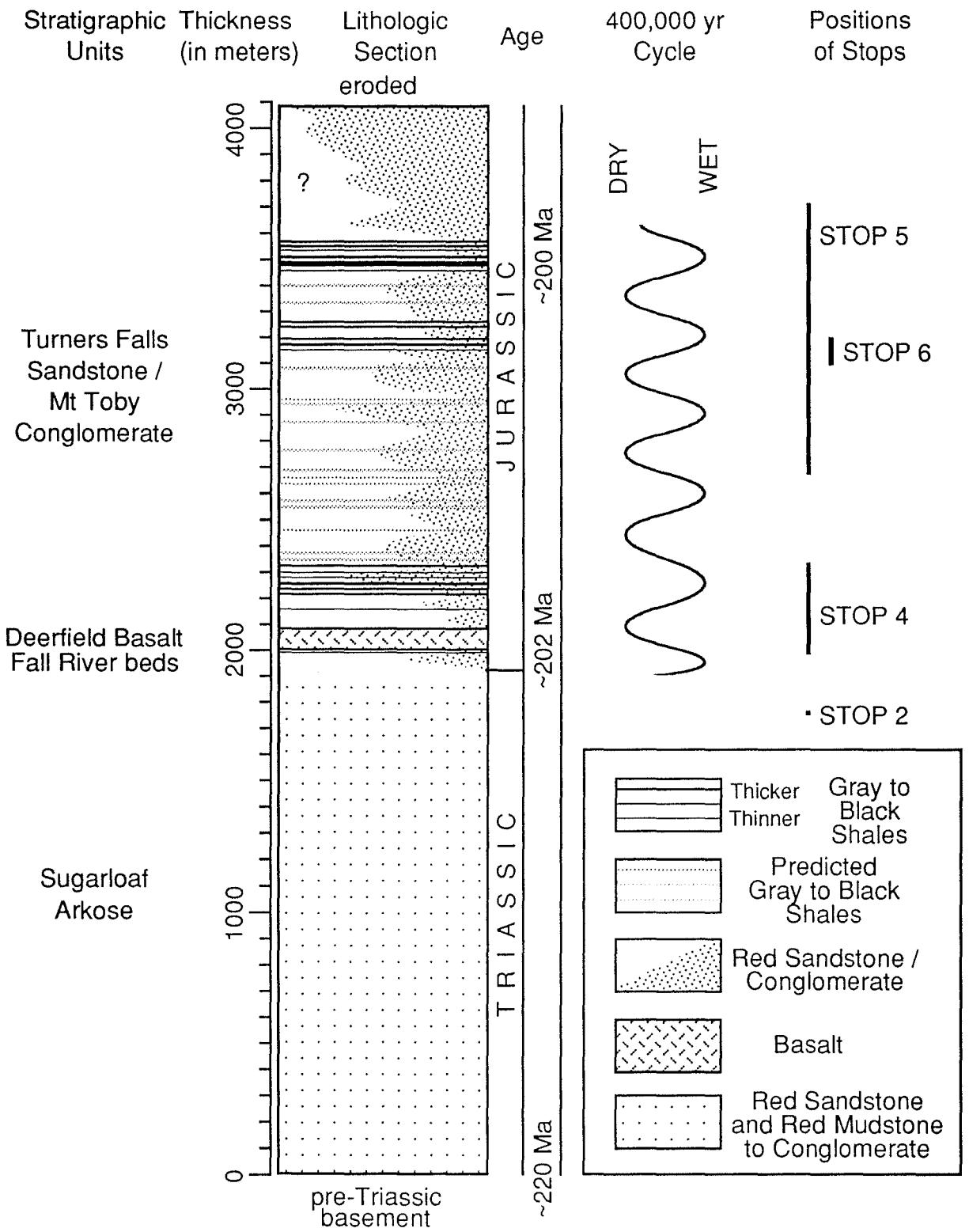



Figure 3. Stratigraphy of the Deerfield basin.

fossil fishes from North America. In the same report, Hitchcock describes and illustrates probable conifer branch and stem fragments and leafy shoots from Sunderland and Turners Falls, and also mentions the occurrence of "phytolites" [probable *Scoyenia* burrows - see Marche, 1992] in red sandstones of the Sugarloaf Arkose near Deerfield. Early figures and descriptions of Sunderland fishes were also included in Louis Agassiz's monumental treatise Poissons Fossiles (1833-1843), although the fishes were incorrectly classified because of poor preservation.

The highly productive fossil fish beds below the dam at Turners Falls (Stop 4) apparently were not discovered until the late 1850's. Hitchcock's works of 1823, 1833, and 1841 make no mention of Turners Falls fishes, although the occurrence of dark, bituminous shales and fossil plant debris at the locality had long been recognized. Emmons (1857) first records the existence of fishes at Turners Falls, but detailed descriptions of the occurrence and systematics of the fossils have only recently been provided by McDonald (1982) and Olsen, *et al.*, (1982; 1989).

In 1802, as the popular legend goes, a boy named Pliny Moody unearthed the first specimens of dinosaur tracks while plowing on his family's farm in South Hadley, Massachusetts in the Hartford basin. The impressions, still preserved in the Pratt Museum of Amherst College (Stop 1), looked as if they had been made by enormous birds, and they were commonly referred to as the tracks of "Noah's Raven" (Figure 4). Although it was not known at the time, these tracks were the first evidence of dinosaurs found in North America (Colbert, 1961). Subsequently, additional examples of these "stony bird tracks" were observed in the Connecticut Valley, but they were not described in the literature until Edward Hitchcock's classic Ornithichnology of 1836. Fascinated by the geological and biological implications of the trackways as well as by their being manifestations of the Divine Hand, Hitchcock devoted the last three decades of his life to their study. He published more than 30 reports on footmarks, and amassed a collection which at one time exceeded 20,000 impressions - still the largest collection of trace fossils in the world (Belt, 1989).

At first many in the scientific community were skeptical of the authenticity of Hitchcock's "footmarks", but his interpretations of their origin were finally sustained when a committee of eminent geologists (Rogers, *et al.*, 1841) visited his localities and examined his evidence. Several of Hitchcock's early specimens were obtained from the Deerfield basin (Turners Falls area) by Dr. James Deane, a successful Greenfield surgeon. Deane also supplied specimens to Dr. J.C. Warren of Harvard University, who in 1854 illustrated a track slab from Greenfield by means of a photograph - the first use of a photograph in an American scientific publication. As word of the spectacular footmarks spread throughout scientific circles in this country and Europe, Deane decided to publish his own findings. In his second report, however, he claimed credit as the discoverer and the first to recognize the scientific importance of the footmarks (Deane, 1844). This ignited a bitter controversy between Hitchcock and Deane, punctuated by a series of vituperative attacks and counter attacks in the literature (Steinbock, 1989). Although Deane's ultimate contributions to paleontology were significant, history has justifiably come to recognize Hitchcock's preeminence in developing the field of ichnology, the study of fossil tracks.

To his final days, in spite of mounting evidence to the contrary, Hitchcock remained convinced that many of his three-toed tracks were those of large, flightless birds, like the extinct moa of New Zealand or the modern rhea. The Connecticut Valley tracks are now known to be primarily those of reptile, including small to medium-sized dinosaurs and crocodiles. But considering the current view that birds are the direct descendents of theropod dinosaurs, Hitchcock's conclusions of more than a century ago cannot be faulted.

Apart from the additions of C.H. Hitchcock to his father's work and the revisions of the footprints by Lull (1904, 1915, 1953) (see below), there have been very few substantial published contributions to Deerfield basin paleontology. Systematic work on the fishes was conducted by J.H. Redfield (1837, 1845), Newberry (1887, 1888), Woodward (1895), and Eastman (1911). The insects (*Mormolucoides* - see below) were described by Scudder (1886). Between 1953 and 1980 the only works that specifically addressed aspects of Deerfield basin paleontology were those of Bock (1949), Cornet (1977a), Cornet and Traverse (1975), McDonald (1975), and McDonald (1992). Especially unfortunate has been a lack of attention to the stratigraphical, paleoecological, and evolutionary context of the fossil remains. Within more recent years, however, there has been a general upsurge in interest, with the works of Cornet, Hubert and his students, McCune, McDonald, and Olsen (see below).

**Paleontology**

Sedimentary strata of the Deerfield basin are very rich in fossils. We have summarized the diversity and of fossils in this basin as currently understood in Table 1. This list represents a very filtered version of information in the literature, and until more thorough studies can be completed it must be regarded as tentative and subjective. All of the major groups of organisms in this list (plants, insects, fishes, and reptiles, and reptile tracks) require some additional description.

**Plants**

Remains of fossil plants are abundant in the rocks of the Deerfield basin. Root traces represented by clay films, linear trains of rhizoconcretions, and rarely organic material are common in much of the section; however, there have been no systematic, paleoecologically-oriented studies of these root traces in the Deerfield basin, or anywhere else in the Newark Supergroup for that matter. This is unfortunate, because roots and the fossil soils they invade are an important key to understanding the metabolism of terrestrial ecosystems, of which we know precious little in the Mesozoic.

Pollen and spores occur abundantly in fine gray siltstones and some claystones in the Fall River beds and Turners Falls Sandstone. Cornet and Traverse (1957) and Cornet (1977a) have described assemblages of fossil pollen and spores (palynoflorules) from several localities within the basin, but as yet no detailed studies have been carried out. The pollen and spores provide some of the most important evidence of a Jurassic-age for the post-Sugarloaf Arkose strata, largely by the complete absence of characteristic Triassic forms (Cornet and Traverse, 1975; Cornet and Olsen, 1985). As is true for Jurassic age strata of eastern North America in general, the relative pollen abundance of the extinct gymnosperm group, the Cheirolepidaceae (Figure 5), in virtually all Jurassic palynoflorules from the basin indicates that this group was dominant. Pollen of cheirolepidaceous confers is distinctive, and is usually ascribed to the genus *Corollina* (or its junior synonym, *Classopollis*). Also present are lower percentages of arauricarian conifer pollen, cycadophyte and ginkophyte pollen, fern and horsetail spores, pinacian conifer pollen, possible pteridosperm pollen, and a variety of enigmatic forms.

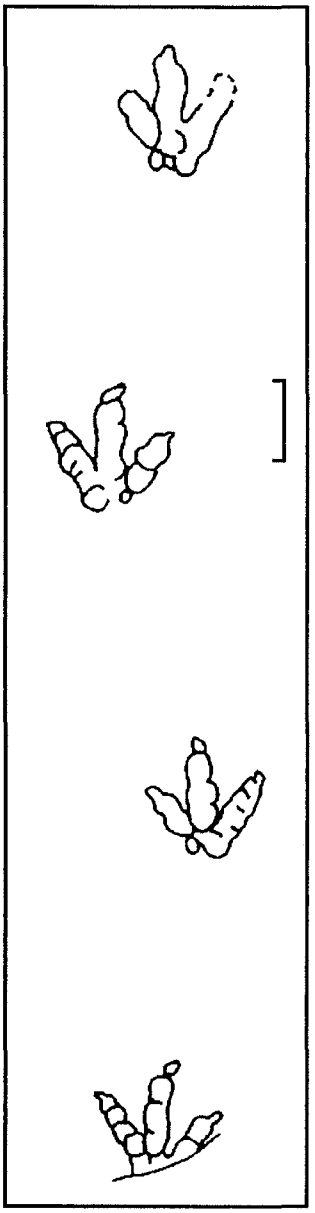
As would be expected from the pollen record, leafy shoots, isolated seed cone scales, male cones, and associated stems and small logs of cheirolepidaceous conifers are the most common plant macrofossils. Apparently, a considerable diversity of leafy shoot taxa and cone scales (Figure 5) are present (Cornet *in* Olsen, *et al.*, 1989). The abundance of well preserved material suggests that it should be possible to find associated organ taxa, and to piece together which cones belong with which pollen and leafy shoots, etc., as has been done elsewhere in the Newark Supergroup (Cornet, 1977b). Leafy shoots of arauricarian and possibly pinacian forms have also been found (Figure 5), along with fragments of cycadophyte and fern fronds (Figure 6).

The fern *Clathropteris* (Figure 6) occurs sporadically in the Deerfield basin. While only fragments of leaves have been found in the Turners Falls Sandstone, complete *Clathropteris* specimens associated with the horsetail *Equisetites* have been found buried *in situ* at one locality in the Fall River beds (Stop 4, locality 1). For reasons not yet understood, complete *in situ* remains of *Clathropteris* occur commonly in other formations correlative with the Fall River beds, but nowhere else. Similar occurences of *in situ Clathropteris* have been found in the lower Shuttle Meadow Formation of the Hartford basin (Hitchcock, 1858), the lower Feltville Formation of the Newark basin (Cornet, 1977a: p. 395; Olsen, 1980b), and the lower Midland Formation of the Culpeper basin. Perhaps these occurrences reflect higher accumulation rates on higher relief flood plains and deltas than was common during the rest of the Newark Jurassic (cf. Olsen and Schlische, 1990).

**Invertebrates**

Invertebrate trace fossils are very common in the Deerfield basin and occur both as burrows and walking traces. The dominant burrow type in the Sugarloaf Arkose is *Scoyenia* cf. *gracilis*. *Scoyenia* was so abundant that many Sugarloaf mudstones are completely bioturbated; it was clearly very important to the continental ecosystem. *Scoyenia* is a back-filled burrow with characteristic rice-grain-like prod marks on the outside and a laminated, "meniscate" fill. The maker of *Scoyenia* is at present uncertain, but was probably a decapod crustacean, most likely a burrowing crayfish (Olsen, 1977; 1988). In the lower Turners Falls Sandstone (Stop 4), small (3-8 mm) *Scoyenia*-like burrows occur abundantly along with small *Planolites* (a curving cylindrical burrow fill); however, these are not well studied.

Arthropod walking and swimming traces are abundant in the Turners Falls Sandstone, especially in fine-grained reptile footprint-bearing beds. Hitchcock (1858, 1865) named many forms of traces, which were summarized by Lull (1915, 1917, 1953). As is the case with the reptile footprints, the taxonomy of the invertebrate traces is in need of revision, and we therefore do not list the taxa in Table 1. For example, Hitchcock applied the name *Aeigmichnus multiformis* to a form characterized by one or more central grooves flanked by closely-spaced, subparallel groves of uniform length. Hitchcock (1865) believed this trace to have been made by annelid worms, while Lull (1915, 1953), Moodie (1930), and Pettijohn and Potter (1964) attribute this morphology to rill marks. In contrast, Gilmore (1928) and Bowlds (1989) identify near identical traces from the Permian of Arizona and New Mexico as the trackways of trilobites, while Hunt, *et al.* (1990), making direct reference to Hitchcock's (1858) published description, refer specimens of this form from the New Mexico Permian to an undetermined arthropod. The extremely uncertain affinities of this trace (?)fossil illustrates the larger problem that inhibits the taxonomic revision of Hitchcock's type and referred specimens. According to the last census (Lull, 1953), Hitchcock named at least 27 genera and 52 species of invertebrate ichnofossils from the Hartford and Deerfield basins alone. We believe a critical reevaluation of these taxa will significantly reduce the number of valid form genera and "species". We note that any taxonomic



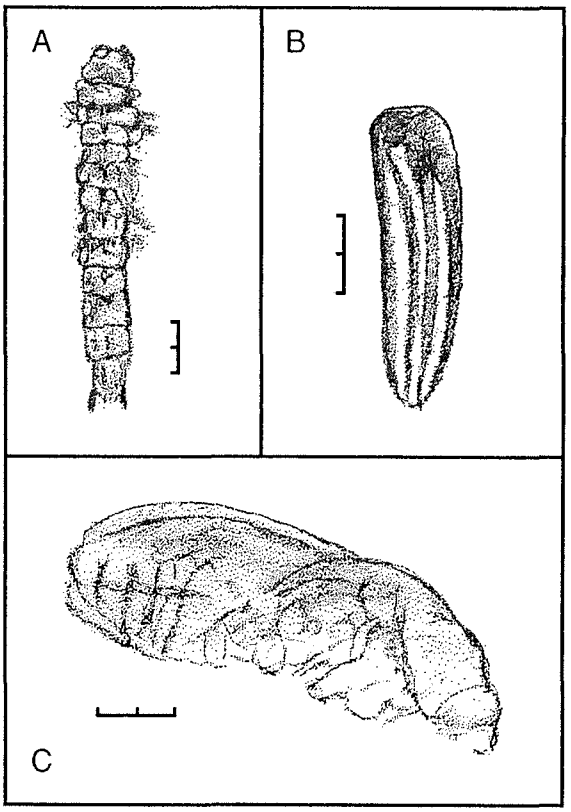



Figure 7. Insects from the Horse Race (Stop 5, locality 4) (above). Drawings were made on a camera lucida, scanned and converted to bit mapped images. **A**, An example of the insect (beetle?) larva *Mormolucoides* (Stop 5); **B**, Beetle elytron of the *Holcoptera-type*; **C**, Whole insect (?*Holcoptera-type* beetle). Scale is 2 mm.

Figure 4. Original slab of footprints dug up by Pliny Moody in 1800-1802. Slab is a natural cast in red sandstone of the ornithischian ichnite *Anomoepus* (AC 16/2) from Moody's Corner, South Hadley, MA (Portland Formation, Hartford basin). These are the tracks that were called "Noah's Raven". Scale is 10 cm.

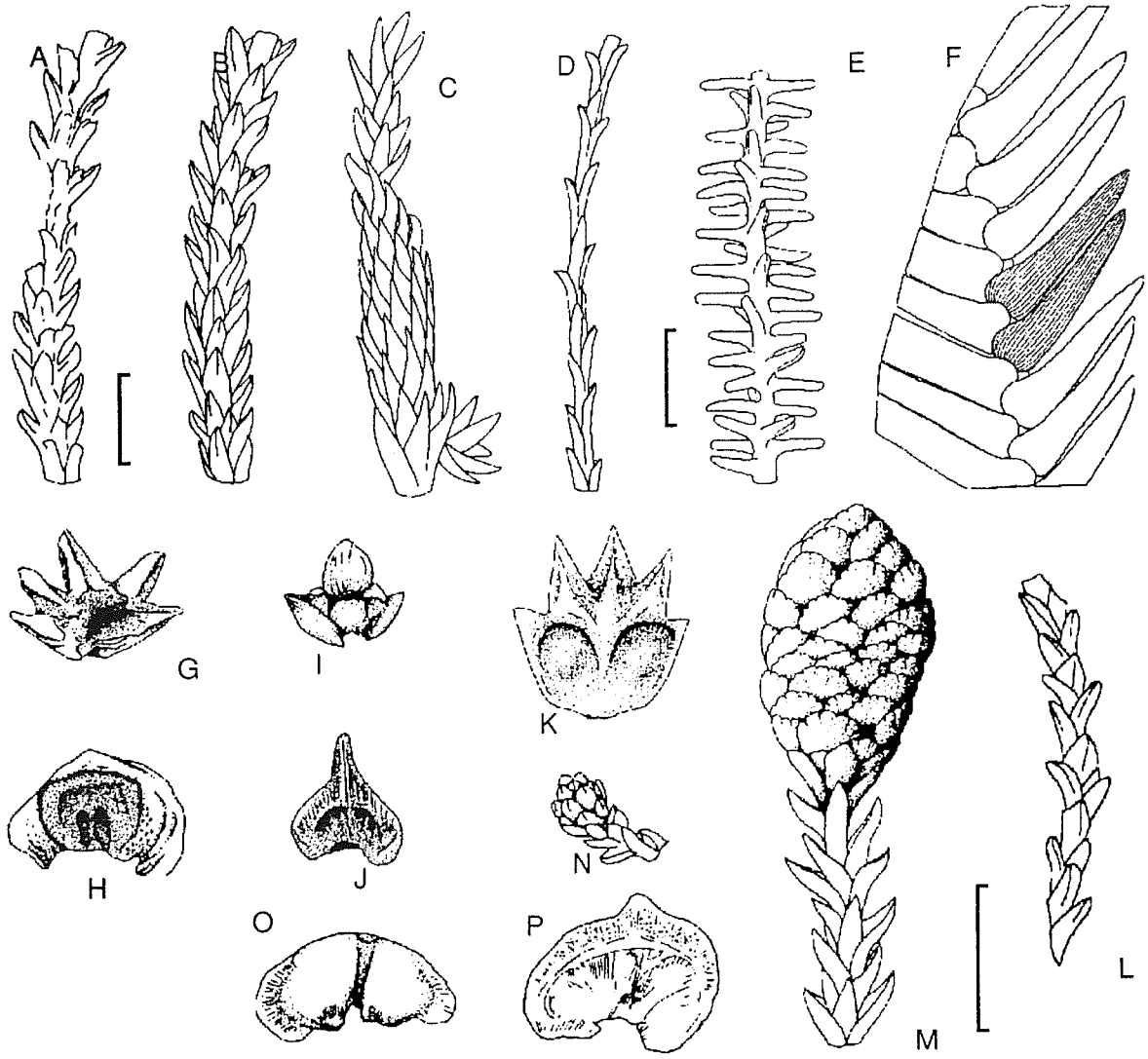



Figure 5. Examples of the kinds of plants found in the Deerfield basin. **A**, *Pagiophyllum simile* (YPM 025, RC, lower Portland Fm., Hartford basin, South Hadley, MA). **B**, Reconstruction of *Pagiophyllum simile* in A. **C**, Reconstruction of *Pagiophyllum* sp. (YPM 020 and YPM 021 - counterparts, RC, lower Portland Fm., South Hadley Falls, MA). Resembles extant *Araucaria imbricata*. **D**, Reconstruction of *Brachyphyllum* sp., Cupressaceae? (YPM 002, RC, R.S. Lull collector, Deerfield basin, Turners Falls, MA). **E**, Reconstruction of *Elatides* sp. Taxodiaceae?, (YPM 024, RC, lower Portland Fm., Hartford basin, Chicopee, MA). **F**, *Otozamites latior*, Bennettitales (WU 824, LC, Shuttle Meadow Fm., Hartford basin, Durham, CT). **G**, Reconstruction of *Hirmerella* sp., Cheirolepidaceae, ovuliferous cone scale (HOL-7lowest Portland Fm., Holyoke, MA). **H**, Reconstruction of *Hirmerella* sp., Cheirolepidaceae, sterile bract from female cone probably belongs with G (HOL-7, lowest Portland Fm., Holyoke, MA). **I**, Reconstructed ovuliferous cone scale of *Hirmerella münsteri* , Cheirolepidacae (YPM 012, RC, Horse Race [Stop 5], Montague, MA). This is virtually identical to the "types" from Rhaeto-Liassic of Middle Franconia, Germany. **J**, Reconstruction of sterile bract of female cone *Hirmerella*? sp. , Cheirolepidaceae? (upper Towaco Fm., Newark basin, Roseland quarry, NJ). **K**, Reconstruction of ovuliferous cone scale of *Hirmerella* sp. , Cheirolepidacae (MC, Turners Falls Ss, Deerfield basin, Horse Race, MA). **L**, Reconstruction of *Pagiophyllum brevifolium* (MC, upper Turners Falls Ss, Deerfield basin, Horse Race, MA - on the same specimen and bedding surface as K). **M**, Ovuliferous cone terminally attached to *Pagiophyllum simile* leafy shoot. Either this specimen is an immature seed cone of *Hirmerella*, or more likely it is closely allied to extant *Cryptomeria japonica* and Early Jurassic *Swedenborgia cryptomeroides*, both of taxodiaceous affinity. **N**, Reconstruction of *Masculostrobus* sp. (pollen cone), Cheirolepidaceae (ER 008 and ER 009 [counterparts], ERT, lower Feltville Fm., Newark basin, New Germantown, NJ). **O**, Reconstruction of pair of winged seeds, Pinaceae? (ER 085, ERT, lower Feltville Fm., Newark basin, New Germantown, NJ). **P**, Reconstruction of *Hirmerella* cf. *H. münsteri* , Cheirolepidaceae, sterile bract from female cone (ER 014, ERT, lower Feltville Fm., Newark basin, New Germantown, NJ). Collection and locality abbreviations are: YPM, Yale Peabody Museum; WU, Wesleyan University; LC, Loper Collection; RC, Redfield Collection; MC, N.G. McDonald Collection; ER, East Round Top Collection of Bruce Cornet; ERT, East Round Top. Scale is 1 cm: for A-C, bar is next to A; for D-F it is next to E; for G-Q it next to P.

revision will have to take into account the exceedingly poor record of Early Mesozoic terrestrial invertebrate body fossils, especially insects, and we therefore doubt that the majority of these invertebrate ichnofossils from the Deerfield basin will ever be confidently related to the organism that made them.

Although insect body fossils are uncommon in early Mesozoic continental strata, the Deerfield basin is one place where there is hope of filling in the record. Hitchcock (1858) described *Mormolucoides articulatu*, the first insect body fossil from the Newark Supergroup, based on several specimens collected by Roswell Field from the Turners Falls Sandstone. Dana (1858) studied Hitchcock's specimens and identified them as insect larvae, possibly Neuroptera. Scudder (1886) examined these and other specimens and noted that they resembled beetle larvae, but concluded again they were neuropteran. We feel that their affinities are still very uncertain. Recently, two of us (NGM and PH) rediscovered one of Hitchcock's localities along the Horse Race in Montague and recovered many additional specimens (Figure 7). Several types of larvae present, as well as associated beetle elytra assignable to *Holcoptera*. The latter are small (7 mm), narrowly ovoid in form and are ornamented by 3-4 longitudinal stripes. *Holcoptera* was originally described from slightly younger, marine deposits of the English Lias (obtusum zone; Whalley, 1985), and we follow Whalley's (1985) concept of this taxon which relegates it to a form genus for beetle elytra displaying 3-4 longitudinal, parallel stripes that cannot at present be assigned to any extant family. We have also found one complete but indeterminate adult beetle that also could be *Holcoptera*, although the elytra are mostly obscured by the abdomen. It seems plausible to us, although far from demonstrated, that *Mormolucoides articulatus* could be the aquatic larva of a beetle with *Holcoptera*-type elytra. At least four other localities produce *Mormolucoides* in the Deerfield basin (see Stop 5 and road log at 44.8 and 45.5 mi.), and it appears to be a facies-indicator fossil. In the Hartford basin, *Mormolucoides* has been found in the East Berlin Formation and possibly the Shuttle Meadow Formation (McDonald, 1992), and *Holcoptera* has been found in the Portland Formation at the now filled in K & F quarry in Suffield (McDonald, 1992). We believe careful searching in similar facies will reveal a much greater diversity of insects.

**Fishes**

Fossil fish from the Deerfield basin (Turners Falls Sandstone) were among the first North American vertebrates to receive formal systematic attention. Agassiz (1833-1843) described two species from the Sunderland fish bed, *Paleoniscum fultus* and *Eurynotus tenuiceps*, which were later recognized as belonging to the holostean genus *Semionotus* (Figure 8). Additional species of *Semionotus* were described from the Turners Falls Sandstone by Redfield (1841), Newberry (1888) and Eastman (1911). Isolated scales and bones occur in several of the gray and black shale beds in the basin, but articulated specimens have been found only in "lake bed 3" in the vicinity of Turners Falls (Stop 4), in the Sunderland fish bed (which may be a coarser equivalent of lake bed 3), and in the gray shales at Sunderland Cave (Bain, 1932). What makes the first two localities interesting is the array of shapes and sizes of the abundant *Semionotus* that prompted the multitude of species names in the older literature. Some variation can be attributed to postdepositional distortion of the fish, which can be quite subtle. Nonetheless, most of the variation is unequivocally real and is reflected not only in body shape, but also in scale and fin ray counts, and scale and skull bone morphology. This is particularly clear in the structure of the prominent ridge of scales along the dorsal midline in front of the dorsal fin (Figure 8) - the dorsal ridge scales. Similar variation is seen in several other Newark Supergroup *Semionotus* assemblages. However, mass mortality assemblages of *Semionotus* from pond and shallow lake deposits from elsewhere in the world show much less variation, and are in this respect similar to modern lacustrine fish species (McCune, 1982).

Recent work by Olsen, *et al.* (1982), McCune (1982, 1986, 1987a, 1987b, 1990) and McCune, *et al.* (1984) attributes the remarkable morphological diversity of Newark Supergroup *Semionotus* to variation between many closely related species (i.e. species flocks: Figure 9) that evolved within giant lake systems. Such species flocks are well known in cichlid fishes from the African great lakes, especially lakes Victoria, Malawi, and Tanganyika (Echelle and Kornfield, 1984). However, recognition of individual species in species flocks of semionotids is difficult, because, as with the African cichlids, there are few non-overlapping consistent characters. Furthermore, we cannot do breeding experiments or observe the natural behavior of extinct fish. Hence, quantification of the actual diversity of semionotids is impeded. Nevertheless, dorsal ridge scale characters and skull bone shape do allow the grouping of specimens into easily recognized and clearly definable supra-specific categories (Olsen, *et al.*, 1982; Olsen and McCune, 1991).

A few years ago, two of us (PEO and NGM) thought that only semionotids were present in the Deerfield basin. This conclusion was substantiated by our own collecting efforts and an exhaustive search of museum collections. The kinds of semionotids present, the apparent absence of other groups of fishes, and the distribution of these forms in the rest of the Newark prompted an attempted correlation of the Newark Supergroup based on fishes (Olsen, *et al.*, 1982). Almost immediately after this paper was published, a specimen of the palaeonisciform genus *Redfieldius* was found (by PEO) at Turners Falls. Subsequently, two more specimens of *Redfieldius* were discovered (by NGM and PH), and two specimens of the coelacanth *Diplurus* were found (by PH and Alasdair Gilfillan). The discovery of *Redfieldius* and *Diplurus* in the Turners Falls Sandstone, and similar discoveries from other parts of the Newark,

demonstrate that correlation by these fishes (at least at the generic level) does not work (cf. Olsen, 1983; Huber, *et al.*, in press).

Of special interest is the *Diplurus* specimen collected and prepared by Alasdair Gilfillan, who has kindly allowed us to examine it. It comes from the nodule horizon in the middle of "lake bed 3" (see Stop 4). The specimen has a standard body length of nearly 40 cm, not including the skull and caudal peduncle, and is complete except for the anterior part of the head, which had weathered prior to collection. There is a phosphatic, conchoidally fracturing probable cololite (partially digested fecal matter) in the region of the body approximating the position of the posterior part of the digestive tract. The cololite is elongate and ovoid and compares well in shape, fracture pattern, and phosphatic composition with coprolites attributed to coelacanths from elsewhere in the Newark (McDonald, 1992). Additional study is needed, however, because some coelacanths have an ossified lung which could be mistaken for a cololite, and a source from other upper-level, aquatic predators such as sharks or protosuchian crocodiles cannot be completely ruled out.

**Reptiles**

The osseous remains of probable tetrapods in the Deerfield basin are represented by two bone fragments: one from the Sugarloaf Arkose; the other from the Turners Falls Sandstone. The Sugarloaf specimen was discovered by Solon Wiley in Greenfield in ?1875 and presented to Professor O.C. Marsh of Yale, where it was catalogued as YPM 6281. Lull (1953) and Galton (1976) regarded this bone fragment as presumably dinosaurian without additional comment or description. We regard its identification as dinosaurian as very suspect. The second specimen was discovered (by PH and NGM) in a large transported block of pebbly sandstone in Turners Falls. The specimen is a blue-weathered, hollow bone fragment about 2.5 cm in diameter, exposed in oblique section. It extends into the matrix an unknown distance. The bone is quite thin ~0.5 cm. Thin, hollow bones are a shared derived character of theropod dinosaurs. We conclude that this fragment may be a portion of the distal end of a long bone of a medium sized theropod, and as such it is the first record of a theropod bone from the Deerfield basin. The block of pebbly sandstone contains clasts up to 10 cm in diameter and excellent armored mud balls (see Little, 1982). The block came from the abutments of the "Red Suspension Bridge", which formerly spanned the Connecticut River upstrean of the Turners Falls-Gill dam (observed by Richard Little, pers. comm., 1992). According to Ms. Therrisa Rice (of Turners Falls) the original stratigraphic origin of the block is from an abandoned quarry (presently occupied by buildings) on the west side of Main St. in the Village of Turners Falls.

This occurrence of tetrapod bones is similar to the reptile bone occurrence described by Huber and McDonald (In Press) in the Pomperaug basin of Connecticut. The Pomperaug bones are in coarse-grained, poorly-sorted fluvial arkose with extrabasinal clasts up to 23 cm in diameter. Thus, the two occurrences are taphonomic analogs, preserved in high-velocity fluvial channel-lags.

**Footprints**

Despite the paucity of osseous remains of tetrapods, the Deerfield basin, specifically the Turners Falls Sandstone, has become famous for its extraordinarily abundant reptile footprints. However, despite superb material, the state of footprint taxonomy has been abysmal. By 1864, the year of his death, Hitchcock had introduced no less than 105 species in 48 genera - not including those names he erected, and subsequently discarded. Unfortunately, most of Hitchcocks taxa are not valid using modern criteria.

Because footprints are the result of an action of an animal and not the actual remains of an organism, they cannot belong to zoological species, which owe their distinctness to genetic isolation, and can be grouped into higher taxa. Historically, however, footprints have been given zoological names corresponding to species and higher categories. Following many years of intense debate, the International Committee on Zoological Nomenclature has formally recognized the ichnospecies, ichnogenus, and ichnofamily as formal names. Unlike the situation for zoological taxa, there is no consensus on how ichnospecies are to be defined, and students of fossil footprints have used radically different approaches. Some recognize every different shaped track as a distinct taxon, regardless of the origin of those differences. Others apply names only to those tracks that show the anatomical structure of the trackmaker's feet (e.g. Padian and Olsen, 1984). These fundamental philosophical differences, plus, quite a bit of sloppiness, have made footprint study very difficult.

Hitchcock began applying zoological names to footprints almost as soon as they were first discovered, which had a very unfortunate and far-reaching effect on the study of vertebrate ichnology. The earlier specimens were of relatively poor quality, but these were the ones described and figured by Hitchcock in his many publications. Most of the thousands of spectacular and beautifully-preserved tracks in the Pratt Museum of Amherst College were collected after the publication of the *Ichnology* (Hitchcock, 1858; Deane, 1861). As a consequence, there is little recognition of the very high quality of Connecticut Valley material in the older literature, and most of the the type specimens of Hitchcock's taxa are indeterminate by modern standards. Some indication of the quality of the later material is to be found in the posthumously published *Supplement to the Ichnology of New England* (1865), in which

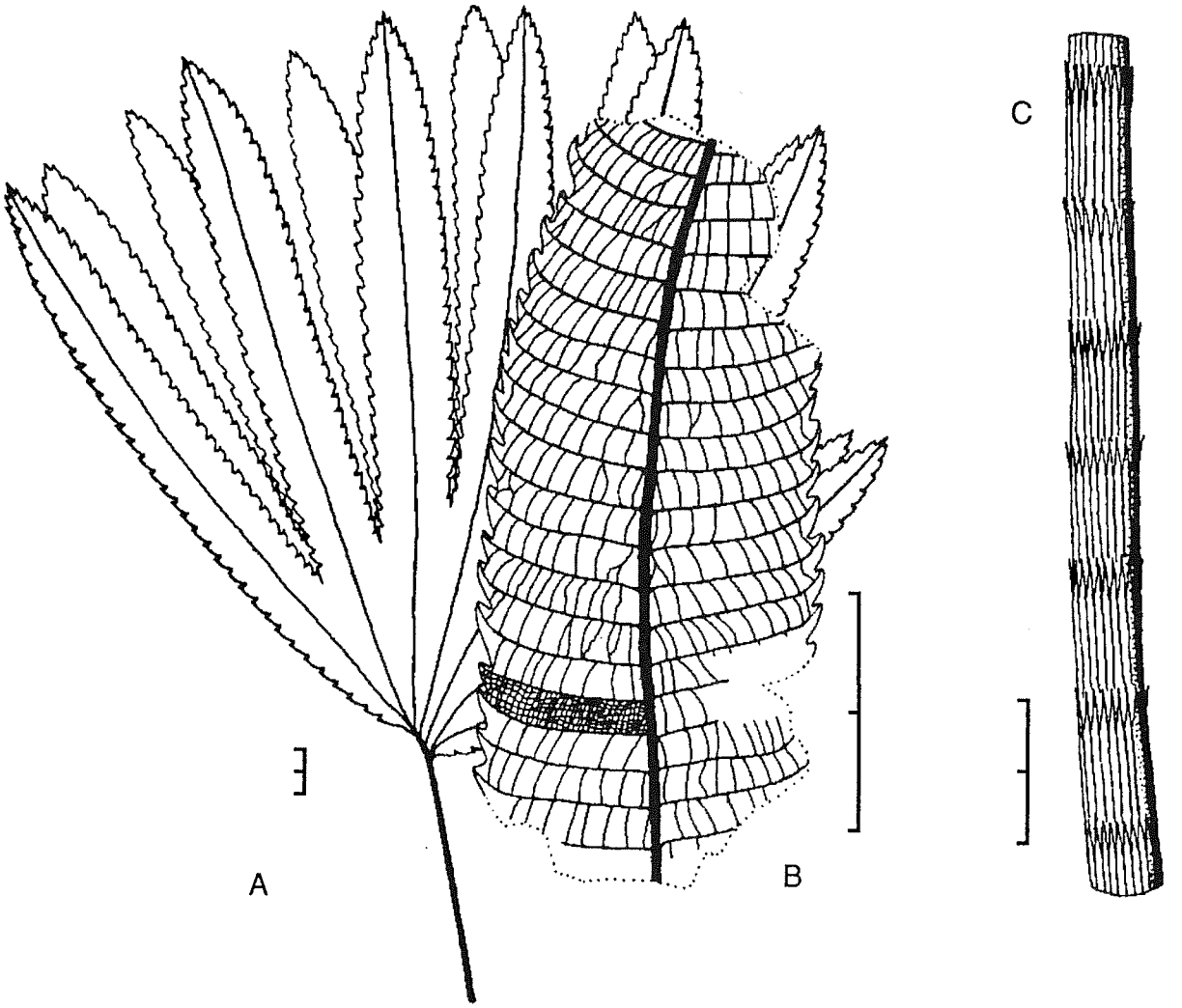



Figure 6. *Clathropteris meniscoides* (Brongniart) Brongniart. **A**, Reconstruction of specimen CL004 from Southampton Road locality, Hampden County, MA. Shuttle Meadow Formation, Early Jurassic. This locality is approximately equivalent in age to the *Clathropteris* locality directly beneath the Deerfield basalt on Rt. 2a (Stop 4, location 1), uppermost Sugarloaf Arkose (i.e. uppermost Falls River beds). **B**, Pinna fragment of *C. meniscoides* (specimen CL001, Southampton Road locality, MA). **C**, *Equisetites* sp., reconstruction of specimen from Southampton Road locality, MA. Equisetites spp. has been found in growth position along side *Clathropteris* both at Stop 4, location 1 and at the Southampton Road locality. Scale bar is 2 cm. Photograph of *Clathropteris* specimen from Stop 4, locality 1 is figured in Olsen, et al. (1989).

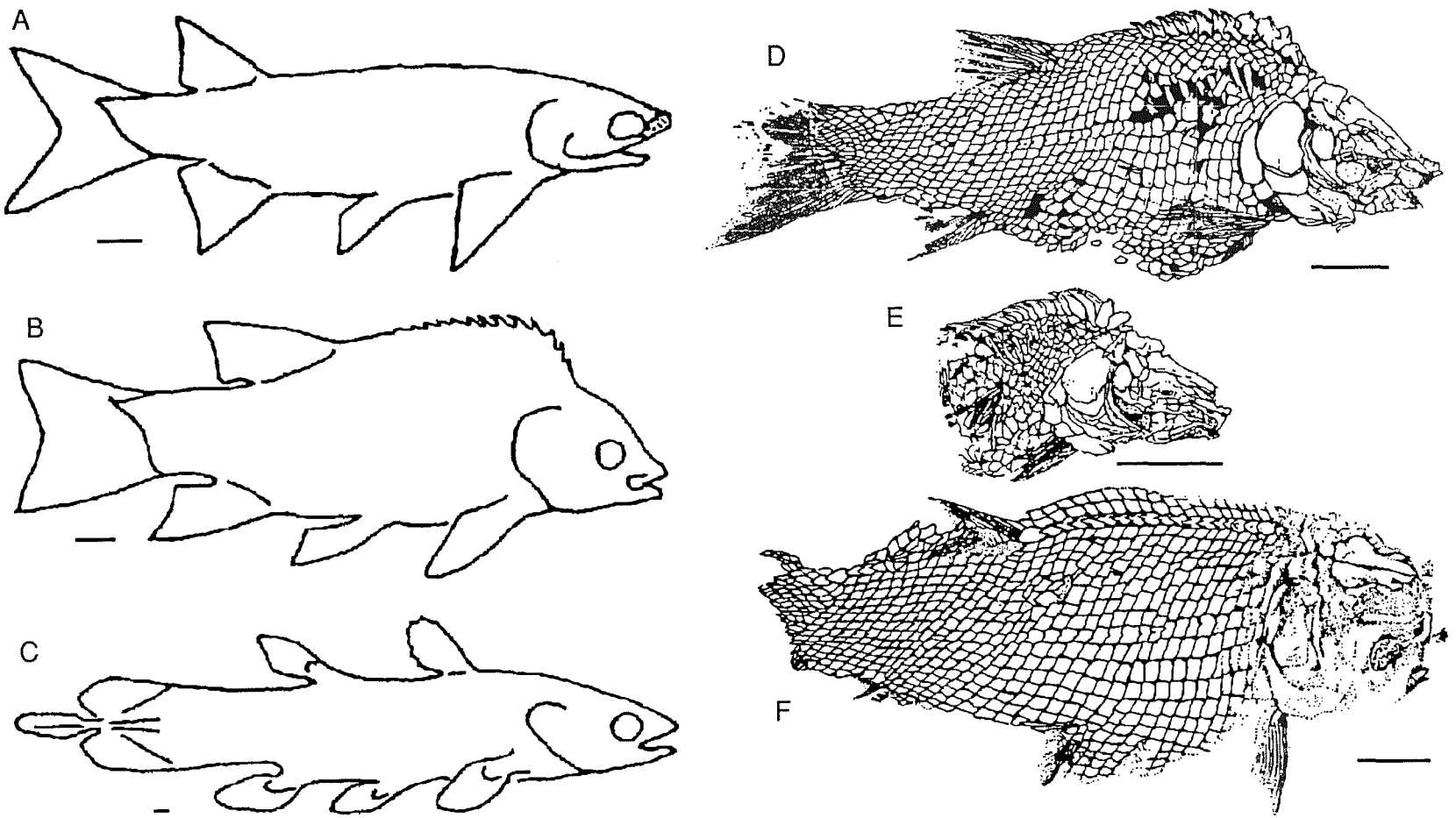



Figure 8. Fish from the Turners Falls Sandstone, Deerfield basin. **A-C**, Outline drawings of fish genera: **A**, the palaeonisciform *Redfieldius*; **B**, the holostean *Semionotus*; **C**, the coelacanth *Diplurus*. **D-F**, *Semionotus* spp., specimens: **D**, member of the "*S. tenuiceps* species group" (YPM 8162) from the Sunderland fish bed, Sunderland; **E**, small member of the "*S. tenuiceps* species group" (YPM 6960) from lake bed # 3 (Stop 4), Turner's Falls; **F**, member of the "small scale group" (YPM 6898) from the Sunderland fish bed, Sunderland. Scale is 2 cm. All adapted from Olsen *et al.* (1982).

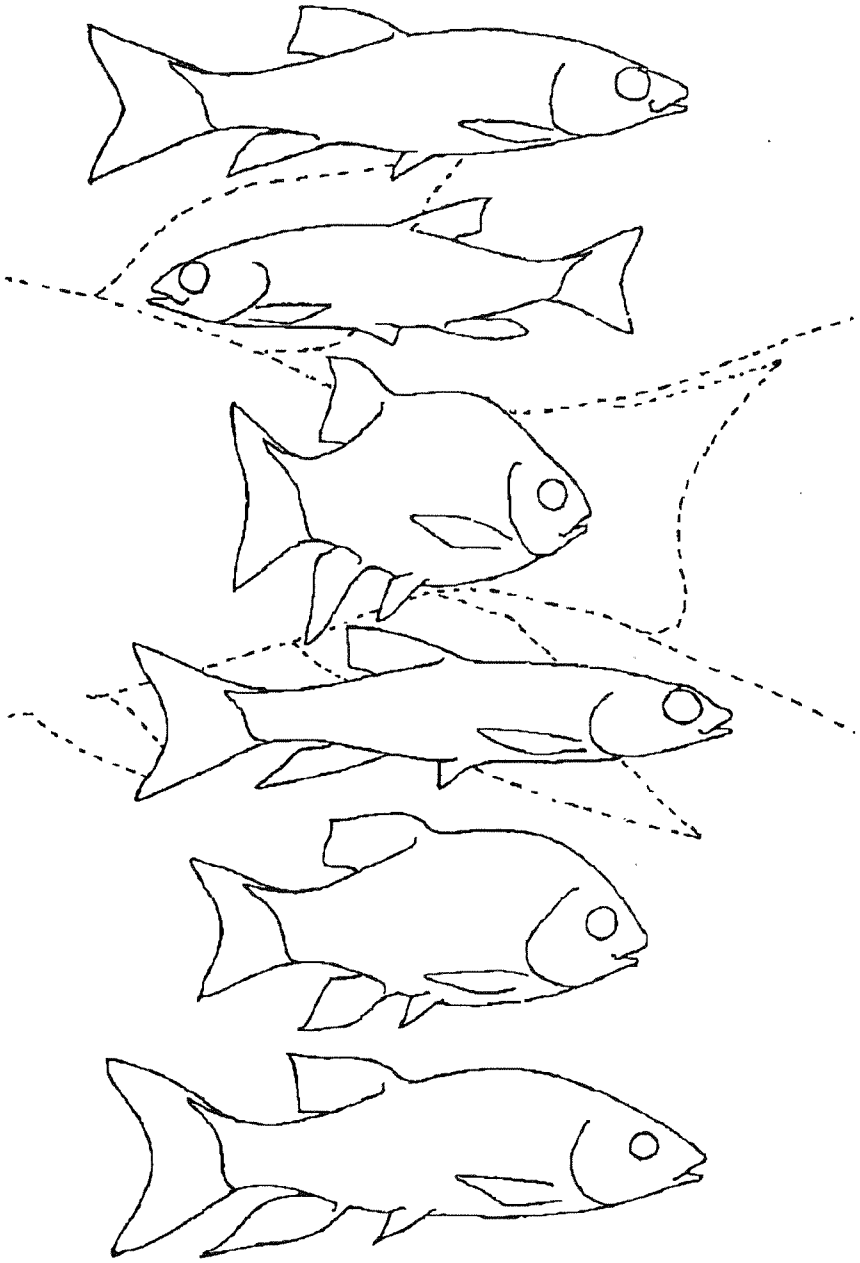


Figure 9. Examples of the range of body shapes seen in *Semionotus* spp. from the Turners Falls sandstone. Based on specimens in the Yale and McDonald collections. Note that the dotted line outlines the posterior portion of a very large *Semionotus*.

all of the footprints in the Hitchcock collection were catalogued by C.H. Hitchcock. In many ways, this is the most useful of the elder Hitchcock's publications because it synthesizes much his previous work of 30 years and attaches specimen numbers to what were only names and/or crude drawings in previous reports. Unfortunately, the taxonomic status of most of Hitchcock's taxa has been completely muddled by repeated renaming of forms and seemingly endless numbers of different names applied to specimens of the same taxon.

Most ichnological studies since Hitchcock have only made matters worse. An excellent example of the confused state of affairs is the history of the ichnotaxon *Sauropus barrattii*, which started out as an isolated, indeterminate manus impression from an unknown locality (Hitchcock, 1837). It ended up as a fancifully "reconstructed" sitting trackway complete with a trackmaker, which was used to infer how Cretaceous hadrosaurs sat (Lull, 1953)! From 1837, "*Sauropus*" went through seven invalid name changes, three invalid changes of type specimens, and four crude redrawings in which dotted lines were slowly replaced with solid ones. The final result was fiction, repeated in every major ichnological compendium to this day (see Olsen and Baird, 1986, for a full history). That is just one taxon out of the 47 listed by Lull (1953). Most of the other genera have had a similarly contorted history.

A footprint is the result of the action of an animal against the substrate. Rarely is the footprint, even at the time it was made, a faithful impression of the foot. As Baird (1980) has pointed out, " a footprint is not the natural mode of a morphological structure but is, instead, the record of that structure in dynamic contact with a plastic substrate." This relationship between the geometric action of the foot (kinematics), the physical properties of the substrate, and the actual morphology of the foot can be represented by a ternary diagram (Figure 10) (Padian and Olsen, 1984a) into which any footprint can be qualitatively placed. All three determinants play important roles in the final footprint. The collected specimen may also have added diagenetic changes, may be structurally deformed (Silvestri and Olsen, 1989), and may have been altered during collection.

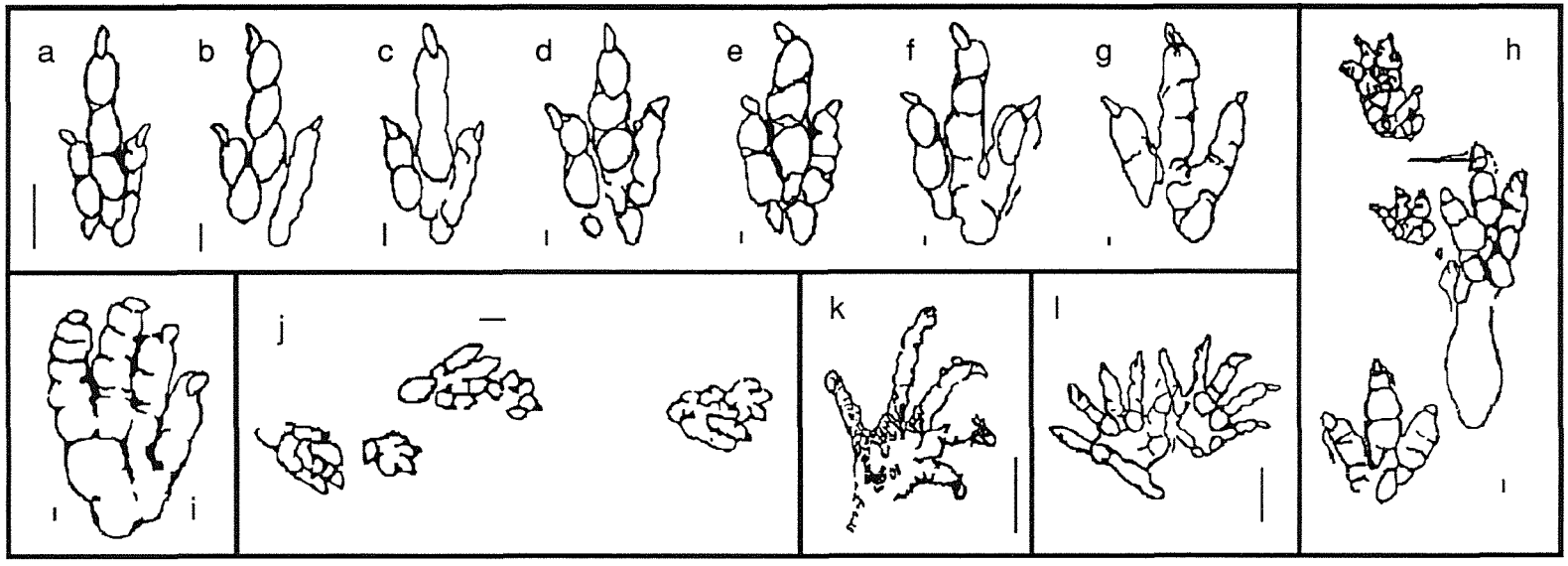
Vital information about the behavior and mechanics of animal motion can be obtained by the study of tracks in which the nature of the substrate and the movement of the animal are most important to track structure. However, we believe that only those tracks in which morphology is faithfully reproduced should be named. These are the forms that are useful for determining the track makers, or as proxies of zoological taxa for biostratigraphic or paleoecological studies. Operationally, this criteria is recognizing when there are well defined impressions of pads or skin impressions, the later being especially definitive. Using these criteria we recognize only four valid ichnogenera from the Deerfield basin. These are *Batrachopus*, *Otozoum*, *Grallator*, and *Anomoepus* (Figure 10).

*Batrachopus* (Figures 10 and 11) is a small, habitually quadrupedal track characterized by a functionally four-toed pes and an outwardly turned five-toed manus. It is a very common ichnotaxon in the Turners Falls Sandstone, throughout the Jurassic of the Newark Supergroup, and in the very youngest Triassic in the Newark basin. The genus has recently been reviewed and revised by Olsen and Padian (1986). The pes of *Batrachopus* has a pad underlying digit V in combination with a reduced number of pads underlying digit I, indicating that the trackmaker had a reduced number of phalanges in that digit - the latter being a shared derived character of crocodiles (Olsen and Padian, 1986). The striking resemblance between crocodile morphology and *Batrachopus* was first noted by Deane (1861). Lull unknowingly inferred the association of *Batrachopus* with crocodiles as well. In 1904, he suggested that *B. "gracilis"* (i.e. *B. deweyi* - Olsen and Padian, 1986) is probably the trackway of *Stegomosuchus longipes*, a small armored reptile from the Portland Formation of the Hartford basin. It was originally thought by Lull to be a "pseudosuchian" allied to the Triassic aetosaur *Stegomus*, but later proved to be a Jurassic crocodile (Walker, 1968). Small, long-legged crocodilian skeletal remains are quite common in Early Jurassic age deposits elsewhere in the Newark (Fundy basin - Olsen, *et al.*, 1989). We conclude that the ichnogenus *Batrachopus* was made by small, mostly terrestrial crocodilians, although it could probably also have been made by small sphenosuchian crocodilomorphs.

*Otozoum* (Figures 10 and 11) is a generally uncommon and often very large, habitually bipedal ichnite. It is known from relatively small forms (*Otozoum minus* Lull, 1915) from the Turners Falls Sandstone. It is characterized by a functionally four-toed pes that retains a well developed pad for digit V. It more or less looks like a giant *Batrachopus*, but has a full complement of phalanges in digit IV. *Otozoum* also occurs in the the Portland Formation of Hartford basin, and the McCoy Brook Formation of the Fundy basin.

Supposed manus impressions of *Otozoum* are known from only one specimen (AC 5/14 - see Stop 1). AC 5/14 is a slab of natural casts bearing one clear pes of *Otozoum*, many *Grallator* tracks, and what has traditionally taken to be a left and right manus of *Otozoum* with one of the manus and the pes impressions overlapping. The structure of these is far from clear, however. After reexamination of AC 5/14, we conclude that the supposed manus tracks are actually superimposed *Grallator* that fortuitously happen to be adjacent to the pedal track of *Otozoum*. Interpreted as *Grallator* tracks, there is no evidence of the structure of the manus of *Otozoum*.

The structure of the manus of *Otozoum* was the principle objection (Baird, 1980) to assigning the genus to the prosauropod dinosaurs, as postulated by Lull (1953). The structure of the pes of prosauropods agrees well with *Otozoum*, including the pad for digit V, and the fit for the manus is, for the time being, unknown. We would predict

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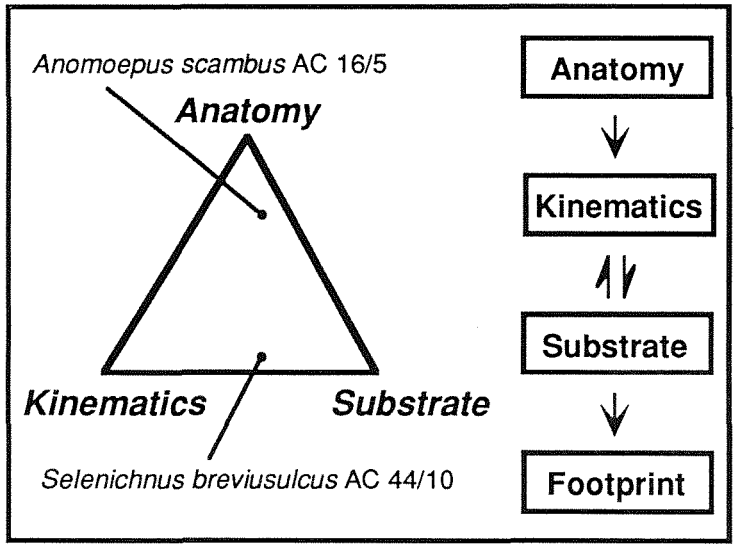



Figure 10. **A**, Valid footprint genera from Early Jurassic age strata of the Newark Supergroup. **a - g**, Theropod dinosaur tracks belonging to *Grallator* spp. from the Towaco Formation of the Newark basin, arranged in order of increasing size: a-c *Grallator* (*Grallator*) spp., d-f, *Grallator* (*Anchisauripus*) spp., g, *Grallator* (*Eubrontes*) sp. **h**, The ornithischian dinosaur ichnite *Anomoepus crassus* (type) from the Towaco Formation of the Newark basin. **i**, Probable prosauropod track *Otozoum* from the Portland Formation of the Hartford basin (based on AC 5/14). **j**, trackway of the crocodilian footprint *Batrachopus deweyi* from the ?East Berlin Formation of the Hartford basin (type, AC 26/5). **k**, Manus impression of the lacertoid track *Rhynchosauroides* sp. from the Towaco Formation of the Newark basin. **l**, Left manus and pes impressions of the possible advanced mammal-like reptile ichnite *Ameghinichnus* sp., from the Towaco Formation of the Newark basin. Note that *Rhynchosauroides* and *Ameghinichnus* have not yet been found in the Deerfield basin. Scale is 1 cm, except for k where it is 5 mm.

**B**, Ternary diagram of the factors influencing footprint morphology and flow chart of the same factors (adapted from Padian and Olsen, 1984). Qualitative positions of *Anomoepus scambus* (AC 16/5) and *Selenichnus breviusulcus* (AC 44/10).

that *Otozoum* should have a manus like that of *Navahopus* (Baird, 1980) from the Early Jurassic Navajo sandstone of Arizona. Therefore we conclude that *Otozoum* is a prosauropod track.

*Grallator*, *Anchisauripus*, and *Eubrontes* (Figures 10 and 11) are names applied to functionally three-toed, bipedal tracks. They are the most common dinosaur footprint forms found in the Deerfield basin and in the Supergroup. The three ichnogenera arrange into a series of increasing size in which larger tracks have relatively shorter middle toes (digit III). Olsen (1980b) showed that the proportional differences that separate these ichnogenera are size-dependant and continuously variable (Figure 10). Practically, specimens of intermediate size cannot be placed to genus. This has led to a massive proliferation of names, most of which we regard as unusable. We recognize *Grallator*, *Anchisauripus*, and *Eubrontes* to be different size classes of the same ichnogenus; the available name with priority is *Grallator* Hitchcock 1858. An older name which would have been more appropriate was *Brontozoum* Hitchcock 1847, and Hitchcock's concept of this ichnogenus was very close to Olsen's idea of the ichnotaxon represented by *Grallator*, *Anchisauripus*, and *Eubrontes*. Unfortunately, *Brontozoum* was suppressed by Baird (1957). Although *Grallator*, *Anchisauripus*, and *Eubrontes*, can not be justified on strictly morphological grounds, it is apparent that the larger forms appear stratigraphically higher in the Newark. Olsen (1980c) suggested that we use *Grallator* (*Grallator*) spp. for the small tracks with a long digit III, *Grallator* (*Anchisauripus*) spp., for the larger, medium size tracks with a medium length digit III, and *Grallator* (*Eubrontes*) spp. for the large tracks with a short digit III. It seems at least the politically correct solution, now that *Eubrontes* is Connecticut's state fossil!

An example of the remarkably confused state of grallatorid nomenclature is provided by *Anchisauripus* Lull 1904. The type species of this genus is *A. sillimani*, the type specimen of which is (according to Lull, 1904) the magnificent slab on display at the Pratt Museum (AC 9/4, Figure 12, Stop 1). from the . This slab, from the Portland Formation of the Hartford basin, was for many years a sidewalk stone in Middletown, Connecticut (Lull, 1915). Unfortunately, this cannot be the correct type specimen. Hitchcock (1858) gave the name *Ornithichnites sillimani* to two specimens he had previously (Hitchcock, 1841; Plate 37, fig. 21, Plate 38, fig. 22) assigned to his *O. tuberosus*. One of these two specimens must be the holotype by original designation; both are from Chicopee Factories, Chicopee, Massachusetts, Portland Formation, Hartford basin). Plate 38, Figure 22 shows a poor impression; while Plate 37, Figure 21 is a much better track (Figure 12). One of us (PEO) has made an exhaustive search for these two specimens, but only that in Plate 37, Figure 21 could be found. It is labeled by deep scratches No. 48, but curiously it has no AC number. Hitchcock (1845) later replaced *O. sillimani* with *Eubrontes dananus* and then put the species in *Brontozoum*. Hitchcock (1858) mysteriously ignores his work of 1841 and substitutes AC 9/4 for the type of the species, which is admittedly a far superior specimen, but it still cannot be the type. Lull (1904) regarded *E. dananus* (not recognizing the older name of *O. sillimani*) as "... undoubtedly the track of *Anchisaurus coelurus*." (p. 487), the latter being the skeleton of a prosauropod. How Lull could have thought that *Anchisaurus coelurus* could have made "*Anchisauripus*" is a mystery, since they could not look more different (Baird, 1957). In any case, the Chicopee specimen No. 48 must be the holotype of *Anchisauripus sillimani* not AC 9/4. It is ironic that Hitchcock (1858; p. 69) states that in regard to AC 9/4, "That upon review of the species, after it is too late to make any alterations because the Plates are struck off, I regret I did not place it under *Grallator*.". Thus, in Hitchcock's own view, what he regarded as the species *A. sillimani* could be classed as *Grallator* to *Eubrontes*. If PEO had his way, *Grallator*, *Anchisauripus*, and *Eubrontes* would all be called *Brontozoum*.

Despite these formidable nomenclatural difficulties, the pedal structure of these grallatorids is perfectly compatible with Triassic-Jurassic theropod dinosaurs (Baird, 1957). The three *Grallator* forms could have been made by several kinds of theropods of different adult sizes, or even one kind of theropod of varying ages. It seems likely to us that there were several kinds of theropods in the valley and that growth and true taxonomic differences are muddled. Needed are detailed studies of single horizons of tracks where different sided individuals are represented and where tectonic deformation can be ruled out or at least corrected for.

*Anomoepus* (Figures 10 and 11) is a usually bipedal ichnogenus with a broadly splayed pes. It has sub-equal toes that frequently have more than one crease between pads, a relatively long digit 1, and a five toed manus print which generally imprints only in sitting tracks. This ichnite is much less common than *Grallator*, but is still fairly abundant. Some of the very best specimens known, including the type, come from the Turners Falls Sandstone. *Anomoepus* also occurs in all other Jurassic formations of the Newark Supergroup.

Lull (1904) recognized the similarity between *Anomoepus* and the best known small ornithischian at that time, *Hypsilophodon*. Now, with the discovery of many small Early Jurassic skeletons of small ornithischians, it is clear that the reconstructed skeleton based on *Anomoepus* is completely compatible with the primitive ornithischians such as *Lesothosaurus* (*Fabrosaurus*), from southern Africa (Olsen and Baird, 1986).

There are some ichnotaxa described by Hitchcock and Lull that do not fit easily into the above four ichnogenera. All of these, however, lack the criteria necessary to show that the tracks reflect the morphology of the track makers, and instead could easily owe their form to behavior or interaction with the substrate.

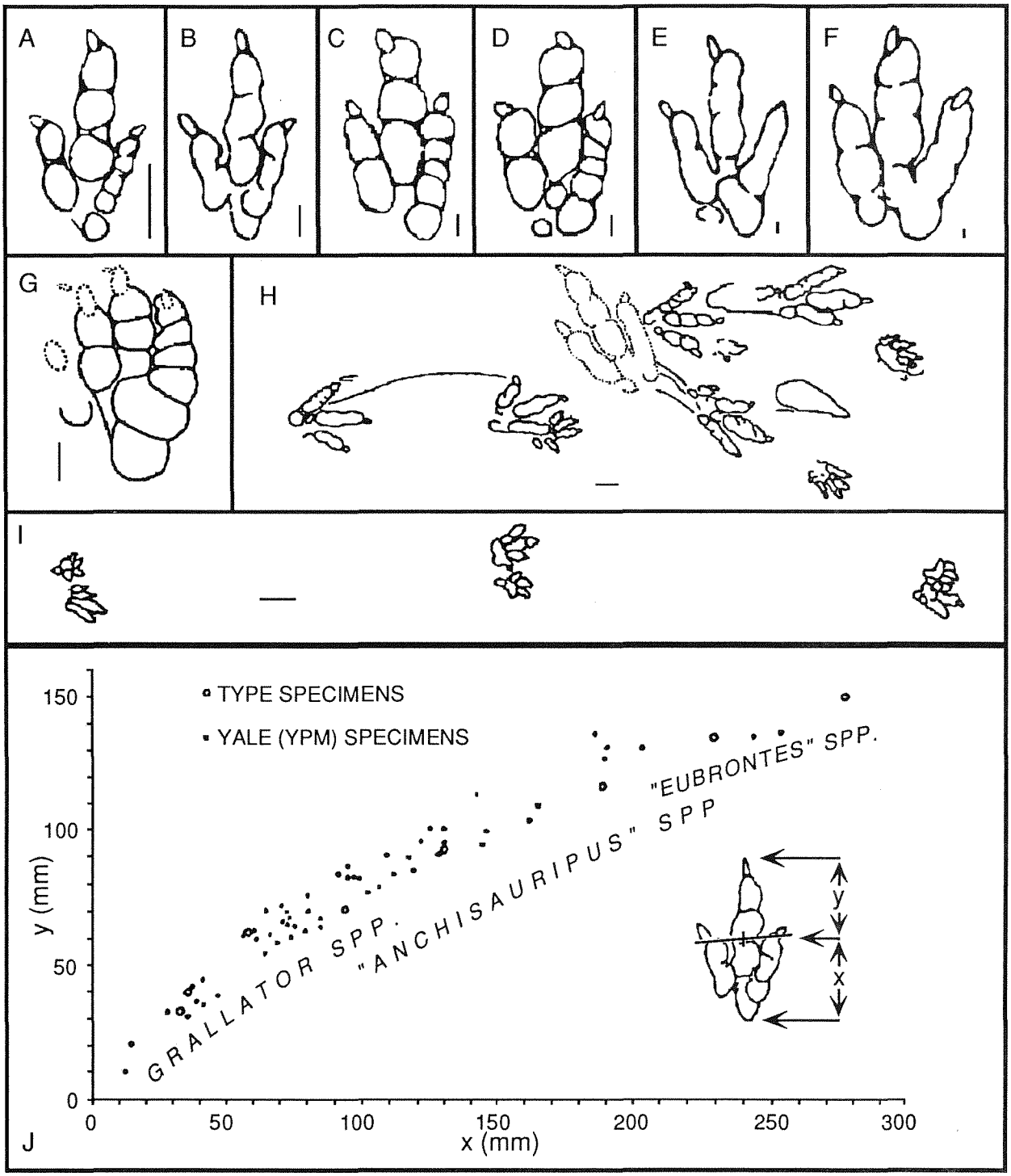



Figure 11. **A-I**, Examples of footprint taxa from the Deerfield basin: **A**, Type of *Grallator tenuis* from Turners Falls (AC 12/3); **B**, Type of *Grallator* (*Anchisauripus*) *hitchcocki* (AC 56/1) from Lily Pond, Gill; **C**, *Grallator* (*Anchisauripus*) sp, (AC 13/4); **D**, *Grallator* (*Anchisauripus*) cf. *tuberosus* (AC 1/1) from Field's Orchard, Gill;, part of the trackways in the confused *Sauropus* saga; **E**, Type of *Grallator* (*Anchisauripus*) *minusculus* (AC 16/1) from Lily Pond, Gill; **F**, *Grallator* (*Eubrontes*) sp. (AC 45/1), locality in Turners Falls Ss. not recorded; **G**, *Otozoum* "*minus*" (drawn flipped, from Lull, 1915) from the Horse Race, Montague (YPM 2046); **H**, Type of *Anomoepus scambus* (AC 16/6), locality in Turners Falls Ss. not recorded; **I**, *Batrachopus gracilis* (AC 42/3); locality in Turners Falls Ss. not recorded; **J**, Relationship between the back of the foot and projection of digit III in *Grallator* tracks (modified from Olsen (1980c).

## PALEOECOLOGY

Ecology is the study of the interactions of organisms with each other and their environment. Paleoecology deals with the fossil record and changes through time in those interactions. The interactions may be direct, such as predation or competition, or indirect interaction, such as control of global biogeochemical cycles. In any case, the emphasis must be on interaction. In paleoecology, the effects of changes in the environment on organisms are as important as the interactions between organisms. Thus, in the strata of the Deerfield basin there is a wealth of ecological information, from the fluvial, alluvial, and cyclical lacustrine sequences and from the record of body and trace fossils.

Very substantial historical change is recorded in the Deerfield basin. The biggest change occurred during the Triassic-Jurassic transition when a global mass extinction wiped out a large proportion of the diversity of higher organisms (Olsen, *et al.*, 1989). In addition, the basin itself shifted environmental and depositional modes. A well drained flood plain, represented by the Sugarloaf Arkose, changed to a largely lacustrine setting characterized by radical and cyclical changes in lake depth through time, represented by strata of the Fall River beds, the Turners Falls Sandstone, and the Mt. Toby Conglomerate. We will therefore discuss the various ecosystems represented in the Deerfield basin in three intervals, representing the basin during Late Triassic, high lake level intervals in the Early Jurassic, and low lake level periods in the Early Jurassic.

### Late Triassic

The paucity of sediments preserving organic matter, body fossils or vertebrate trace fossils in the Sugarloaf Arkose naturally limits what we can say about the Late Triassic ecosystem in the Deerfield basin. However, enough is known about coeval deposits in other Newark basins to at least provide a plausible scenario consistent with the few available fossils.

Sugarloaf Arkose sequences were deposited by braided rivers and streams. Coeval lacustrine deposits in the Newark basin to the south and the Fundy basin to the north show that the climate was strongly seasonal with very distinct wet and dry seasons and with fluctuations in climate from wet to dry at ~20,000 to 2,000,000 year frequencies (Olsen, *et al.*, 1989). This must have had an effect on the ecosystems, but fluvial deposits record those changes poorly.

Primary production in the Deerfield basin during Sugarloaf time was almost certainly dominated by terrestrial plants and additional energy was presumably available for consumers from detritus (also dominated by terrestrial plants) in rivers draining the surrounding highlands. Judging from pollen and spore assemblages from correlative sediments in the Newark basin, the terrestrial vegetation was diverse, with abundant conifers belonging to several major groups, cycad-like plants, seed ferns, true ferns, and horsetails. Forests presumably existed at least sometimes, but even during the longest wetter times, ground cover could not have been as we know it now, because the angiosperms (particularly weeds and grasses) had not yet evolved, even though their ancestors (stem angiophytes) were probably present (Cornet and Traverse, 1975; Cornet, 1989a; 1989b). Perennially wetter areas could have had a cover of ferns and horsetails (fern-savannas *sensu* Crane, 1987: p. 124), but drier areas probably had significant bare ground with widely-spaced cycadophyte shrubs. Without herbaceous angiosperms or grasses, disturbed habitats would have been much slower to recover, and chemical weathering rates would have been lower than they are at the present.

Macroherbivores and omnivores included a diverse suite of lizard- to cow-sized reptiles including procolophonids, sphenodontids, aetosaurs, and small ornithischian dinosaurs, all of which are known from Late Triassic beds elsewhere in the Newark Supergroup, and all but ornithischian dinosaurs are represented by skeletal remains from the correlative New Haven Arkose of the Hartford basin (Olsen, 1980c). Possibly there were some small synapsids ("mammal-like reptiles") and prosauropods (although there is no direct evidence of them in the time-equivalents of the Sugarloaf arkose in surrounding Newark Supergroup basins). Prosauropods, which may not have lived in the Valley during Sugarloaf Arkose time, were the only large herbivores that could have potentially eaten tree foliage.

Direct insect herbivory must have been very different than now. Although there were beetles, diverse hemiptera (true bugs), and various orthoptera resembling grasshoppers, there were no hymenoptera (ants, bees and wasps), or termites. Thus, large scale reduction of woody tissue and leaf material was probably limited. This is supported by the lack of insect damage in fossil wood from Triassic age deposits around the world (Robert Smith, pers. comm., 1991).

Vertebrate carnivores of drier areas almost certainly included rauisuchian "pseudosuchians", small- to medium-sized theropod dinosaurs similar to *Rioarribasaurus* (*Coelophysis*) or *Liliensternus*, and small, terrestrial crocodiles, sphenosuchians, and lizards (all based on footprints from correlative Newark deposits). In and along water courses, there were fish and phytosaurs. A scapula of the latter is known from the New Haven Arkose (Marsh, 1893).

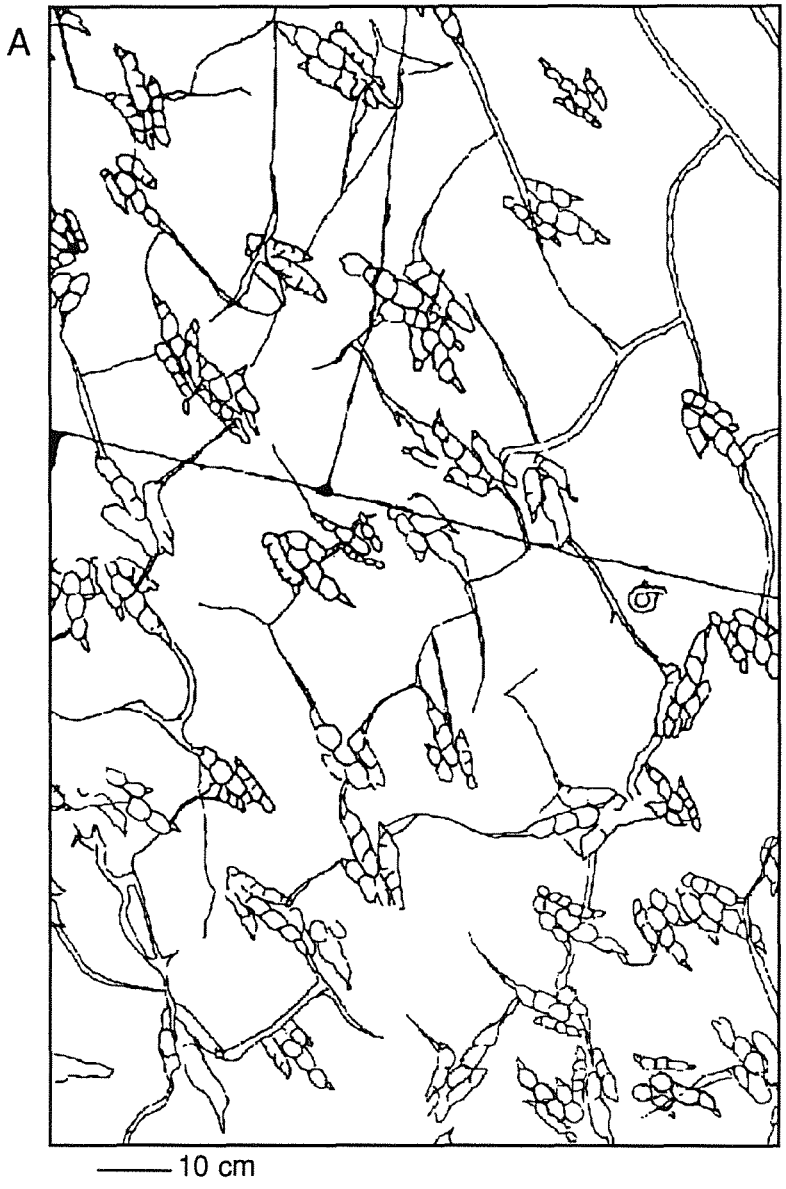




Figure 12. **A**, Slab of *Grallator* tracks (AC 9/14) from the Portland Formation of the Hartford basin, incorrectly assigned as the type specimen of *Anchisauripus sillimani* by Lull (1904).

**B**, The correct type of *Grallator* (*Anchisauripus*) *sillimani* , a left pes impression (AC old collection # 48) from the Portland Formation of Chicopee Factories, Chicopee, MA.

Soil detritivores, other than ants and termites, were fairly well developed (Wing and Sues, 1992) by the Triassic. There is direct evidence for very large-scale detritus processing by decapod crustaceans, namely *Scoyenia* soil burrows. If *Scoyenia* was produced by crayfish or something similar, the water table had to be within a few meters of the surface year-round, and this would have inhibited deep-burrowing soil insects. The bulk of the digestible soil organic matter was probably eaten by crayfish.

As the death of organic fossil preservation testifies, the ecosystem efficiency of the communities inhabiting the Sugarloaf floodplains and rivers was very high, with virtually no organic matter escaping digestion and conversion back to $CO_2$. The ecosystem did, however, bury considerable carbon in the form of $CaCO_3$ in soil caliche. This carbonate results mostly from the direct weathering of Ca-silcates in rock fragments in the arkose itself, greatly aided by the bacterial degradation of plant roots. Thus, the Sugarloaf Arkose was still a net sink for atmospheric $CO_2$.

**Early Jurassic**

The mass extinction event at the Triassic-Jurassic boundary resulted in the extinction of 45% (Olsen, *et al.*, 1987) of all continental tetrapod families as well as a large percentage of marine invertebrates and vertebrates. The best current candidate for what caused this mass extinction is a giant asteroid or impact at an as yet unknown location (Bice, *et al.*, 1992; Olsen, *et al.*, 1990).

Sometime close to the Triassic-Jurassic boundary, the Deerfield basin began to subside at a faster rate, and there was a shift in the depositional mode from fluvial-alluvial in the Sugarloaf Arkose to largely lacustrine-alluvial in the Fall River beds, Turners Falls Sandstone, and Mt. Toby Conglomerate.

Both of these major changes combined to make Deerfield basin ecosystems of the Jurassic substantially different than those of the Triassic. Most dramatic is the spectacular cyclicity recorded by the sedimentary sequence in the Turners Falls Sandstone (Stop 4). This cyclicity was caused by the rise and fall of lake level, as a result of periodic climatic changes. As a direct consequence, ecosystems, as recorded by the sediments, become alternately perennial lake-dominated then lake shore and playa-lake-dominated.

Evidence of Jurassic terrestrial ecosystems in the basin consists of *in situ* trace fossils in the lake shore and playa strata, and allochthonous elements (mostly wood, plant leaves, stems, pollen, and spores). The Jurassic sediments record a massive change in the flora, with at least local elimination of many of the most common plant forms (based on pollen and spore assemblages from the Newark basin, Cornet and Olsen, 1985; Olsen, 1990). An additional very dramatic change is the sudden rise to dominance by the cheirolepidaceous conifers. As was true for the other Newark Supergroup basins, terrestrial ecosystems were no doubt in a constant state of flux because of dramatic changes in precipitation, following Milankovitch cyclicity. Forests were dominated by cheirolepidaceous conifers, but at times araucarian and even pineacian conifers were abundant, perhaps responding to changes in seasonality. In contrast to the Triassic forests, however, those of the Early Jurassic in the Deerfield basin were very low in diversity. This is apparently also true of the marshy areas and understory, a bit of which is preserved *in situ* at Stop 4 in the Fall River beds. Here the dominant forms were the leathery fern, *Clathropteris*, and the reed-like horsetail, *Equisetites*.

The larger herbivores of the Deerfield basin Jurassic were clearly dinosaurian, almost exclusively small "fabrosaurid" ornithischians (e.g. makers of *Anomoepus*) and small to large prosauropods (e.g. makers of *Otozoum*). The former, as evidenced by their footprints, frequently visited the lake shores and drying lake floors, while the latter kept to drier areas. The "fabrosaurids" almost certainly were restricted to browsing on low vegetation, although it is possible they could climb trees. The pad structure seen in the foot and hand prints of *Anomoepus* seem to indicate some grasping adaptations. The prosauropods, however, were large enough to reach foliage in higher trees, and they could use the big claws on their hands to pull down branches. Ornithischian dinosaurs had complex teeth and chewed much of their food, while the prosauropods had peg-like teeth and processed most of their food in their gizzards. Prosauropods were probably specialists in eating cheirolepidaceous conifer trees, as the ranges of both, in time and space, are parallel. Most herbivorous dinosaurs probably had a fermenting gut, and that of prosauropods was very large. The trend to larger sizes seen in prosauropods and sauropods is probably a direct result of the need to increase the residence time of food in the gut, which is itself positively correlated to size. The Jurassic dinosaur-conifer-dominated ecosystems throughout much of the world may have resulted in a global increase in ecosystem efficiency and a decrease in chemical weathering. This would have allowed atmospheric $CO_2$ to build up, increasing the hot-house conditions already partially in effect in the Triassic.

In the Deerfield, perhaps because of more frequent flooding and a higher sedimentation rate, bioturbation is much less prevalent in Jurassic age strata than in the Triassic rocks. In addition, organic matter preservation is higher, even in rooted, lake shore strata. Regardless of the global trend towards higher efficiency, locally the lake shore and lake bed soils had low ecosystem efficiency. Unfortunately, as was the case for the Sugarloaf Arkose, the more evidence there is for *in situ* plant communities, the less is preserved of body fossils and vertebrate trace fossils.

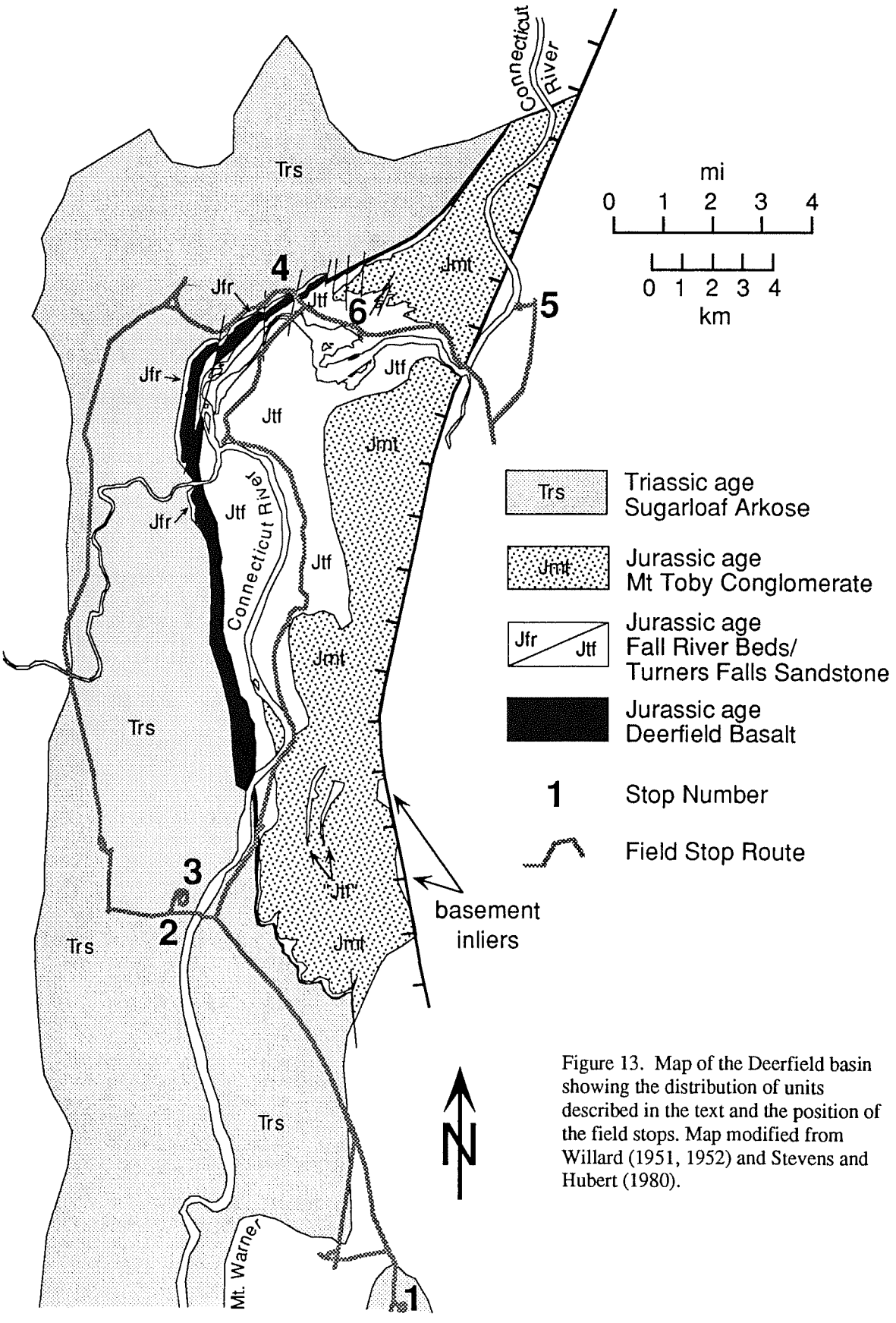



Figure 13. Map of the Deerfield basin showing the distribution of units described in the text and the position of the field stops. Map modified from Willard (1951, 1952) and Stevens and Hubert (1980).

Not surprisingly, therefore, the best footprints come from lake floor or playa floor thin bedded, flaggy mudstones deposited during lake transgression or regression (see Stops 4, 5, and 6). The more bioturbated intervals of the drier phases of the cycles tend to preserve very poor footprints.

Primary productivity in the lakes was probably dominated by blue green algae (i.e. cyanobacteria), because diatoms and dinoflagellates had not yet invaded fresh waters. The amorphous organic material and preserved hydrocarbons that are common in the black shales of Turners Falls Sandstone are probably remains of blue green algae. A significant fraction of the organic matter in the lakes was contributed by terrestrial plants. Wood tracheid cells, leaf cuticles, and pollen and spores are common components of the organic matter in the lacustrine shales. Clam shrimp were probably the dominant zooplankton (they occur in the East Berlin Formation and we expect them to be found in the Deerfield basin), but the now dominant cladocera (water fleas) were absent. Insect larvae of various sorts may also have been important.

During highest water times, the Deerfield basin was evidently filled by very large perennial lakes that may have been connected to even larger lakes in the Hartford basin. During these times, the lakes were perennially stratified with an anoxic hypolimnion. All of these giant lakes had fish, and some of them seem to have had species flocks of semionotid fishes (lake bed 3 - see Stop 4). Whatever benthos existed was limited to the edges of the lake, above the chemocline. Ecosystem efficiency in this perennially stratified lake was low because of the limited residence time of organic matter in the oxygenated zone. As a direct consequence, organic matter preservation was high. Of course, all this changed as the lake level fell in response to increased evaporation during drier times of the climate cycles.

There is substantial evidence that some of the lakes of the upper Turners Falls Sandstone were hypersaline (see Stop 5). Apparently, these lakes were sodium- and chloride-dominated, and at times deposited halite. Such hypersaline waters supported a limited assemblage of aquatic insects (see Stop 5), and were occasionally very shallow. Interestingly, the sedimentation rate was rapid enough or the ecosystem efficiency low enough to preserve considerable organic matter, despite the shallow water conditions.

Fluctuations in lake depth produced fluctuations in ecosystems, the most obvious effect of which was the preservation of much more organic matter because of low ecosystem efficiency in the perennial lake sediments. In addition, when the lakes deepened, the creation of new ecospace provided room for the generically depauperate colonizing fish (especially semionotids) to evolve into many species - most if not all of which were wiped out as the lake dried up. The surrounding highlands and their plant and reptile communities also must have responded to these climatic fluctuations and also to the effects of a moving shoreline.

## ROAD LOG (Figure 13)

**Mileage**

0.0 Road log begins in front of the Pratt Museum, Amherst College.

**STOP 1. PRATT MUSEUM AND THE HITCHCOCK FOOTPRINT COLLECTION.** The Hitchcock footprint collection of the Pratt Museum is the largest fossil footprint collection in the world. The footprint collection of the museum is the fruit of Hitchcock's labors and consistis almost entirely of tracks collected from the Early Jurassic rocks of the Deerfield and Hartford basins. The footprints comprise the basis for the Connecticut Valley-type assemblage (Baird, 1957). C.H. Hitchcock, Edward's son, was reponsible for curating his father's collection as well as clearing up what would have been unfathomable systematic mysteries. Highest on the list of his achievements in ichnology was his editorial help in publishing his father's posthumous *Supplement to the Ichnology of New England* (1865) which is essentially a field guide to the Hitchcock collection. Presumably, C.H. Hitchcock also wrote an unpublished synopsis of tracks in the collection. This museum guide lists examples of track taxa and their type specimens and is invaluable in deciphering the collection. We reproduce the information in that guide in Appendix 1.

When the collection was at its zenith, the footprints were housed in a building named the Appleton Cabinet, erected specifically for that purpose in 1855. The footprints were oriented to the large windows so they could be viewed with the appropriate oblique lighting. Several decades later, however, the space was required another purposes, and the tracks were moved to the basement, the so-called "tombstone room" of the new Pratt Museum. Presently, about two-thirds of the collection is on display, the rest being stored in the closed wooden cabinets. Fortunately, most of the collection remains intact, although it is still largely unstudied.

The bulk of the specimens in the collection are numbered in a way that reflects the original layout of the Appleton Cabinet. Each specimen bears a fraction in which the numerator represents the case, wall, or table that contained it, and the denominator, the specimen number itself. Little remains of the original layout of the collection in the present Pratt Museum. Especially frustrating are specimens in the cabinets; for these, there is no remaining order whatsoever. While a wall or table specimen may be found by a quick scan about the room, a specimen which

is in a cabinet must be found by going through drawers one by one. There are over 2,000 reptile tracks in these cabinets!

Many tracks of unique historical interest are the core of the collection. The original "Noah's Raven" of Pliny Moody (AC 16/2) is on display (Figure 4). According to C.H. Hitchcock (in E. Hitchcock, 1865, p. 52), this slab is "...from South Hadley, near Moody Corner. This specimen was ploughed up by Pliny Moody, in 1800. It was subsequently used for a door-step, then obtained by Dr. Dwight of South Hadley, and finally bought by E.H., one of his heirs. This is the earliest specimen of fossil footmarks anywhere preserved. The tracks were called by Moody, those of Noah's Raven." E. Hitchcock (1858, p. 3), however, records the date of the discovery as 1802. The main tracks on the slab clearly belong to *Anomoepus* sp. (*A. minor* according to C.H. Hitchcock, 1865) and thus, they represent the earliest discovery of dinosaurs in North America (Colbert, 1961).

The main slab of the giant *Otozoum moodii* (AC 4/1) on display is the natural cast of the trackway from Moody Corner pictured in the charming Plate I of Hitchcock's *Ichnology* (1858). Note that the south portion of 4/1 , which appears to be the counterpart of the south portion of AC 3/1, is actually a separate slab, not naturally connected to the north end of AC 4/1 or the under-prints of AC 3/1 (Hitchcock, 1865).

Most of the critical type specimens are present in the collection. All of the important specimens assigned to *Sauropus* (including the mangy type - AC 20/40) are on display, as are the types of the ichnogenera *Anomoepus* (AC 52/10), *Grallator* (AC 4/1 - on the same slab as *Otozoum* ), *Grallator* (*Anchisauripus*) (No. 48 - old collection), *Grallator* (*Eubrontes*) (AC 45/8), *Batrachopus* (AC 26/5), *Otozoum* (AC 5/14), *Hyphepus* (AC 1/3), and *Gigandipus* (AC 9/9). One of us (PEO) has outlined many tracks for illustration purposes in white, water soluble paint (see procedure in Olsen and Baird, 1986). It is extremely instructive to compare the actual specimens with the drawings of the same specimens in publications.

Most of the footprints in the Hitchcock collection come from the Turners Falls Sandstone. Quite a number of tracks, especially those from the upper Turners Falls Sandstone [such as A/C 16/1- *Grallator* (*Anchisauripus*) *minusculus* ) are so finely preserved that skin impressions are visible. This is more common than one would suppose from the literature. A large number of other tracks have very well preserved pad impressions. A few of these require special note. AC 52/10, the type specimen of *Anomoepus curvatus* from Lily Pond, Gill (Stop 6), bears two trackways of the ichnospecies. The two extreme ends of the slab are covered by raindrop impressions which slightly obscure the tracks, and hence were formed after them. The middle third of the slab is smooth and the superposed tracks are clear. A very shallow puddle must have covered the tracks in a low spot at the time of the shower. Another slab from Lily Pond (AC 1/7) is a sitting trackway of *Anomoepus*. A superb but tricky slab (AC 1/1) from Field's Orchard in Gill has about 36 impressions of *Grallator* (*Anchisauripus*) tracks of the *A. tuberosus* variety. This is the slab which caused so much trouble when it was interpreted as *Sauropus* (see Olsen and Baird, 1986, for the messy details). These exquisite tracks clearly lie in the anatomy third of the ternary diagram of Figure 10 and thus are examples from which useful systematic and zoological information may be obtained.

Excellent examples of tracks in the kinematics-substrate portion of Figure 10 are also on display. An example of how behavior and substrate strongly influence morphology of the track is the type trackway of *Selenichnus* (AC 42/6). At the bottom of the slab it appears as an apparently bipedal trackway with a strong tail trace. The pedes appear trydactyl and comma-shaped. However, by the top of the trackway, it is clearly quadrupedal, resembling *Batrachopus* (which it most likely is). A spectacular additional example is the famous "fossil volume" (AC 27/4) on the east side of the room adjacent to the type trackway of the totally indeterminate *Selenichnus breviusulcus* (AC 44/10). The fossil volume consists of two successive tracks deeply impressed in laminated siltstone. Several layers have been split apart revealing the same track on different surfaces, with the layers bound as in a book. The apparent structure of the track changes through the successive layers. On no surface is the actual structure of the foot represented. An example of substrate influence on morphology are the so called 'under-prints' or 'shadow tracks' present on the counterpart slab (AC 3/1) of AC 4/1, the slab bearing *Otozoum*. The actual surface on which the animals walked was a 1/4 inch claystone layer which could not be collected (Hitchcock, 1865). Note the differences between the relatively high fidelity natural casts of the original track surface and the rather different under-prints morphology of the same footprints.

In addition to the unparalleled reptile tracks in the Hitchcock track collection, there are also important examples of Connecticut Valley plants, fish, and invertebrate trails, as well as several excellent specimens of trace fossils from other parts of Eastern North America. The Hitchcock collection is a remarkable, priceless archive, and although it is very old, it has yet to be fully analyzed and described.

Leave Pratt Museum and follow campus circle road to exit.

0.2 Turn right (north) onto S. Pleasant Street (Rt. 116 N).
0.3 Rt. 116 turns left, we go straight on S. Pleasant St.
0.4 Entering Mount Toby Quadrangle mapped by Willard (1951).

0.5 Amherst Center - go straight ahead on N. Pleasant.
0.85 Turn left onto Triangle St. at light.
1.2 Merge with continuation of N. Pleasant.
1.3 University of Massachusetts.
1.4 Leave inlier of arkose according to Willard (1951).
3.0 Intersection with Rt. 63.
3.8 Merge with 116 N.
4.0 Crossing the border fault and entering the main part of Sugarloaf Arkose according to Willard (1951). Chandler (1978) places the border fault about 0.2 mi further to the west, based on gravity data.
4.25 Entering Sunderland and Franklin Co.
5.6 Bull Hill Rd.is on right. Bull Hill is underlain by heavily cemented Sugarloaf Arkose overlain by Deerfield basalt and Mt. Toby Conglomerate.
7.6 View of South Sugarloaf Mountain. Sunderland Center. Continue on 116 N.
7.9 Crossing Connecticut River.
8.3 Turn right into parking area on north side of Rt. 116.

**STOP 2. SUGARLOAF ARKOSE.** Walk ~90 paces west to exposure on north side of Rt. 116. This 10 m section of Sugarloaf Arkose consists of crudely bedded, poorly sorted arkosic sandstone and conglomeratic sandstone. Channel bases are marked by cobble lags. There are vague muddy intervals which might be overbank deposits. Clasts are comprised of granite, schist, and quartzite, and are fairly well rounded. The largest clast we could find is about 25 cm along its long axis. According to Stevens and Hubert (1980), there is an absence of clasts of the kyanite-grade metamorphic rocks that presently crop out east of the border fault. This suggests that the source of the clasts in the arkose was low to medium grade metamorphic rocks structurally high in the Acadian nappes that have since been removed by erosion. There is an obscure hint of cross bedding to the east. Stevens and Hubert indicate a mean direction of transport of 288° for outcrops in this vicinity - close to the 258° mean for all of their Sugarloaf Arkose data. There is vague root mottling the at base of the bench marking the top of this exposure that is probably overlain by a muddier interval. The root mottling suggests that most sedimentary structures are obscured by pervasive bioturbation. This bioturbation, as well as the lack of organic matter preservation, is evidence of the high ecosystem efficiency of the Sugarloaf Arkose communities. According to Stevens and Hubert (1980), most of the Sugarloaf Arkose, as at this site, was deposited by braided streams, as was the case for the contemporaneous New Haven Arkose of the Hartford basin.

Return to car, turn right on 116 N.

8.4 Turn right onto Sugarloaf Street.
8.45 Turn right again immediately into access road for Sugarloaf State Reservation.
Follow access road up the mountain. Note the good outcrops of Sugarloaf arkose along the way.
9.0 Keep right at beginning of access road loop.
9.2 Especially good outcrops are present at left.
9.5 Park in parking lot of Sugarloaf Mountain State Reservation.

**STOP 3. VIEW OF CONNECTION BETWEEN HARTFORD AND DEERFIELD BASINS.** Walk south to overlook area where there is a superb view of the Connecticut Valley, especially the neck between the Hartford and Deerfield basins. This neck is in fact a gentle anticline with several gentle superimposed warps. These kinds of folds are typical of Newark Supergroup basins (Figure 13). The axis of the main fold is roughly NW-SE, at a high angle to the border fault system, and is about parallel with the inferred Late Triassic-Early Jurassic extension direction (Chandler, 1978). As is typical with this type of fold, both the amplitude of the folding and the frequency of minor folds increases towards the border fault. This is seen in this view as the continuous strip of Triassic age arkose bordering the western basement uplands and the alternating patches of arkose and basement on the east, adjacent to the border fault system. In the core of the gentle anticline outcrop basement rocks (Paleozoic sillimanite-grade schists and conglomerate) that comprise Mt. Warner, seen at the bend in the Connecticut River about 5 mi to the SSE.

The homotaxiality of the stratigraphy on the limbs of the anticline is clearly visible here, and the possible connection between the Hartford and Deerfield basins appears less abstract when both basins can be seen from the same vantage. The wavy ridge to the south is supported by Holyoke basalt which is exactly correlative with the Deerfield Basalt. This basalt can be seen to the west underlying Mount Toby Conglomerate and Turners Falls Sandstone. Immediately beneath both basalts are correlative fluvio-lacustrine strata which form a marked bench between the basalt and strong ridge of the underlying arkoses. The induration of the arkose may be a consequences of cements generated from the overlying Early Jurassic age lacustrine strata, either during deposition or early diagenesis. Some of this very hard arkose can be seen poking through the grass at the overlook area.

Return to cars and proceed to exit road, down the mountain.

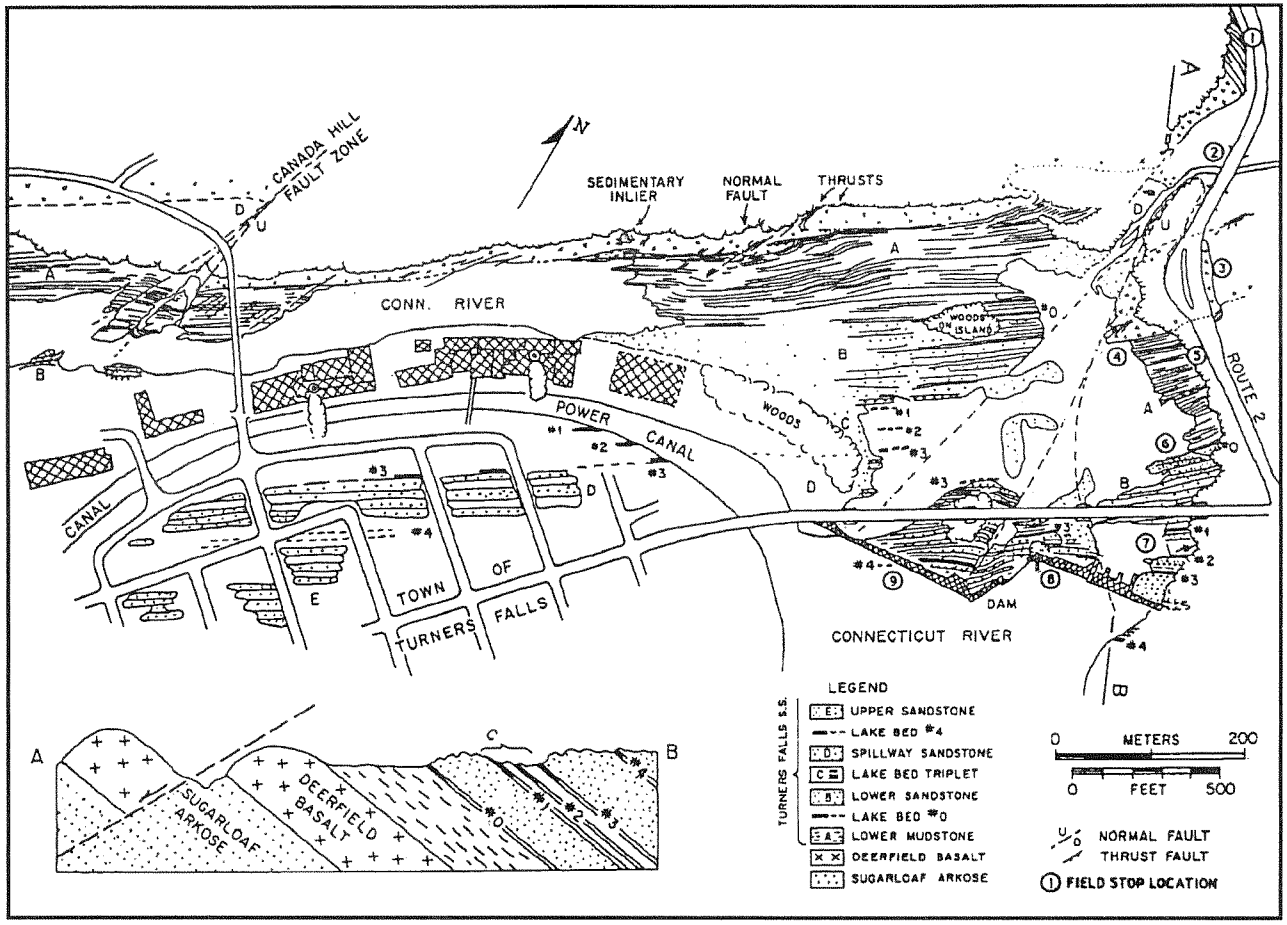



Figure 14. Geologic map of the Turner's Falls area, showing localities for Stop 4. Modified from Wise in Olsen et al. (1989).

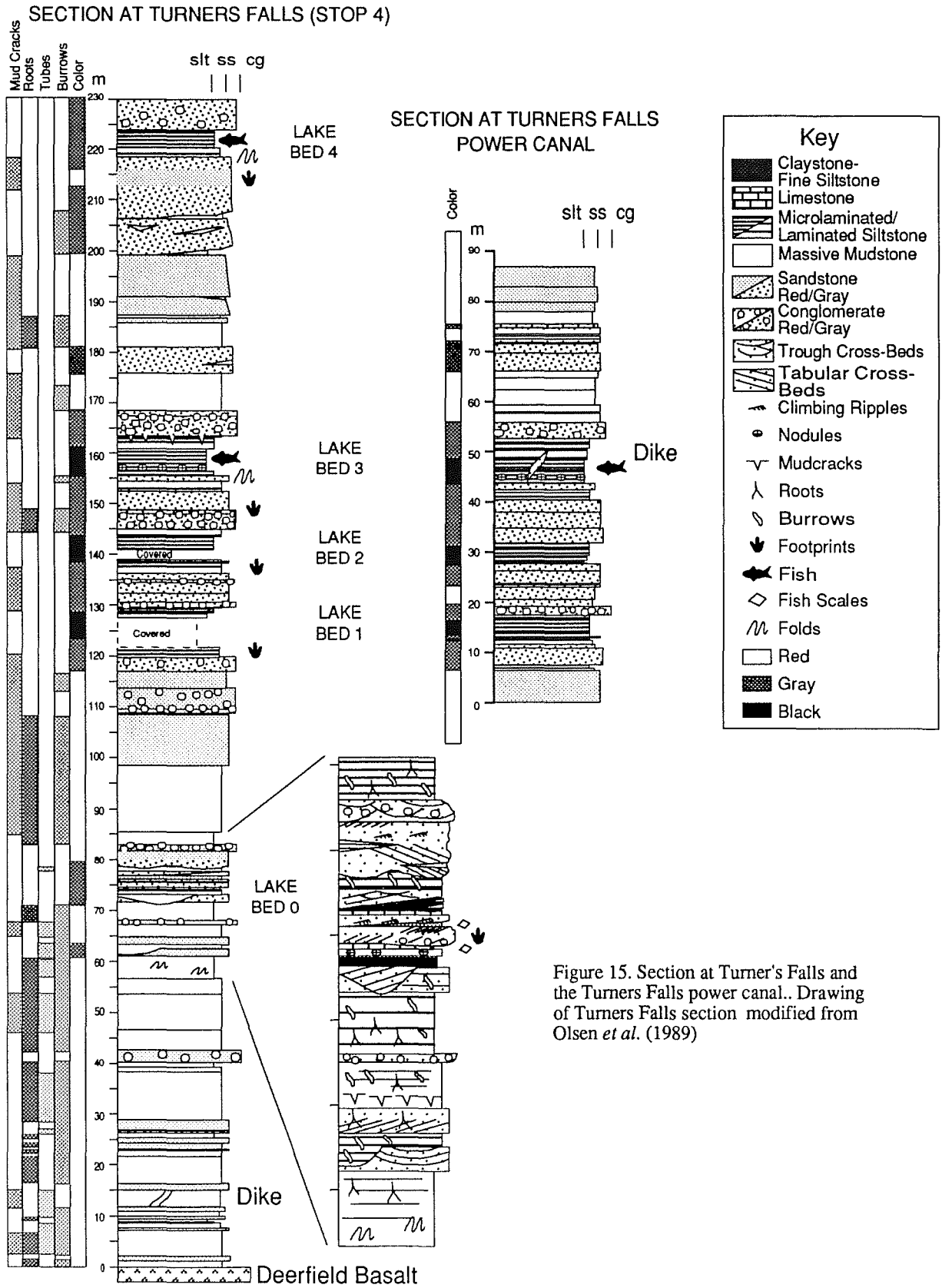



Figure 15. Section at Turner's Falls and the Turners Falls power canal.. Drawing of Turners Falls section modified from Olsen *et al.* (1989)

9.7 Intersection with main road.
10.2 Intersection with Sugarloaf St. Turn left and almost immediately turn right onto 116 N. Highlands at eastern edge of basin visible straight ahead.
11.2 Turn right onto 12-10-116 N (South Deerfield Bypass) following sign for Interstate Rt. 91 N. This road becomes Greenfield Rd.
12.1 Turn left onto 116 N following signs to 91 N.
12.2 Turn right on entrance ramp for 91 N.
14.4 Leaving Mt. Toby Quadrangle of Willard (1951).
15.0 Crossing Deerfield River.
16.3 Exposures of the Sugarloaf Arkose on the east side of the highway have been described by Stevens and Hubert (1980, their locality 6 and Figure 5).
17.3 Exposures on left are of Sugarloaf Arkose.
18.4 Entering Greenfield. About 2.5 km east of here, the Cheapside Quarry exposes the full 80 m thickness of the Deerfield basalt, as well as the upper 10 m of the Fall River beds. Footprints referable to *Batrachopus* sp. are common in this interval. This is the on-strike equivalent of the Fall River beds at Stop 4, location 1. As of June, 1992, quarrying activity had also exposed red beds of the basal 50 m of the Turners Falls Sandstone, which we will examine at Stop 4, location 4.
19.0 Intersection for Rt. 2W Greenfield exit. View of western highlands as they wrap around to east.
20.6 Crossing Green River.
20.9-23.1 Excellent outcrops of Sugarloaf Arkose
23.8 Exit right for Rt. 2 E, ATHOL AND BOSTON; excellent outcrop of Sugarloaf Arkose.
23.1-24.4 Outcrops of uppermost Sugarloaf Arkose on left.
24.5 Outcrops on right of uppermost Fall River beds of Sugarloaf Arkose overlain by Deerfield Basalt.
24.6 Crossing Fall River and Fall River Fault (Emerson, 1898). Pull over just after bridge into parking area on right.

**STOP 4. EXPOSURES AT TURNER'S FALLS** The outcrops at Turner's Falls (Figures 14-17) are some of the most famous in the eastern United States. The are located in the towns of Greenfield and Gill and in the villages of Riverside and Turners Falls. The cataracts called Turner's Falls was originally called Peskeompscut by the local native Americans (Stoughton, 1978). After settlement by Europeans it was called for a short while as Millers Falls or Great Falls, but was finally dubbed Turner's Falls in honor of the massacre of native americans led by Captain William Turner on May 19, 1676, the retaliation for which was additional fighting in which Turner was killed (Stoughton, 1978). According to Stoughton (1978) the name Turner's Falls became fixed in print because of a fictional work about a imaginary native American warrior by a "Julius" (AKA Edward Hitchcock) published in 1828. (By that time the local natives had been exterminated and could be "safely romanticized"). Hitchcock used the name Turner's Falls profusely in his writings afterward. The village of Turners Falls received that name from the adjacent falls when it was first planned by Colonel Alvah Crocker of Fitchburg, his brother William P. Crocker and their associates in 1866-1888 (Stoughton, 1978).

The stop consists of a series of "locations" over a fairly large area. Begin by walking west along the south side of MA 2 (Figures 14 and 16), and cross the bridge over Fall River to the end of the outcrops where the stratigraphic and structural transect will begin. Walk east along exposures (Location 1) to Fall River Bridge (Location 2) then cross road to the section on the north side of MA 2 (Location 3). Descend the hill between the parking area and the river to look at the outcrops upstream (Locations 4-9). The latter exposures are immediately downstream of the Turners Falls dam, and are inaccessible when the dam gates are open.*Water rises rapidly when the dam gates are opened.*

### Stratigraphy and Cycles

The outcrops at Turners Falls and the immediately adjacent area reveal 16 m of the uppermost Sugarloaf Arkose, the full 80 m of the Deerfield Basalt, and 250 m of the Turners Falls Sandstone. The Turners Falls Sandstone (Figure 17) consists of red, gray, and whitish fluvial to lacustrine sandstone and minor conglomerate; red brown fluvial to lacustrine sandstone and siltstone; brick red to maroon lacustrine siltstone and mudstone; and five gray to black lacustrine shale, siltstone, and very minor limestone beds which commonly contain calcareous siltstone concretions. The prominent gray and black shales and limestones present in the Turners Falls Sandstone at this outcrop are numbered from the bottom up: lake beds 0, 1, 2, 3, and 4. These are deepest water portions of lake cycles. They are not the full compliment of cycles present, but they are the most prominent ones and serve as reference and discussion points. This nomenclature follows that of Wise (in Olsen, *et al.*, 1989). These sorts of cycles are called Van Houten cycles after their discover who first recognized them in the Newark basin (Van Houten, 1964; Olsen, 1986).

By analogy with precisely the same pattern of cycles in the Newark basin (Olsen, *et. al.*, 1989), the Van Houten cycles of Turner's Falls were produced by the rise and fall of lakes controlled by climate cycles averaging about 20,000 years. The climate changes were controlled, in turn, by the precession of the equinoxes, modulated by the deformation of the orbit of the Earth. Van Houten cycles vary in the magnitude of the deepest water unit, forming larger cycles of ~100,000, 413,000, and ~2,000,000 years (Olsen, 1986; Olsen and Kent, 1990). The origin of these cycles is as old and persistent as the solar system itself. The precession of the equinoxes is produced by the gravitational pull of the Moon and Sun on the Earth's equatorial bulge, while the longer cycles are caused by deformation of Earth's orbit by the attraction of the other planets to the Earth-Moon system. Ultimately, these celestial mechanical cycles influence the distribution of sunlight on the Earth's surface and thus control climate.

Lake bed 0 is the wettest phase of the first ~100,000 year cycle in the Turners Falls Sandstone (Figure 15). Lake beds 1, 2 and 3 occur in the wettest phase of next 100,000 year cycle, and Lake bed 4 is the first of (probably) three lake beds marking out the next 100,000 year cycle. These upper two ~100,000 year cycles occur in the wettest phase of a 413,000 year cycle, while the lowest (with lake bed 0) is in the driest phase. The Fall river beds are part of the previous 413,000 year cycle. The whole cyclical sequence from the Fall River beds though the exposures at the dam (two almost complete 413,000 year cycles) occur in the wettest phase of a ~2,000,000 year cycle. The same pattern occurs throughout the entire thickness of the Turners Falls Sandstone and Mt. Toby Conglomerate (see Stops 5 and 6).

The cycles seen at Turner's Falls are almost certainly very laterally continous, as has been demonstrated in other Newark basins (Olsen, *et al.*, 1989). The black shales and gray limestones reported at Sunderland and at various other places along the Connecticut River in the Deerfield basin are almost certainly correlative with those at Turner's Falls, although this has yet to be demonstrated. In addition, the climate cycles recorded at Turner's Falls are exactly the same as cycles in at least the Hartford and Newark basins. That allows, along with the geochemical signature of the interbedded basalt sequences (Tollo, in Olsen, *et al.*, 1989; Philpotts and Reichenbach, 1985) a precise correlation of individual cycles in the Newark Supergroup over a distance of at least 500 km (Olsen, *et al.*, 1989).

**Paleontology**

Turner's Falls is one of the premier fossil localities in the Connecticut Valley. Fish are most abundant in the microlaminated shale beds (preserved whole but flattened), and in the center of calcareous siltstone concretions (somewhat dissociated but more three-dimensional). Better preserved, more robust specimens are occasionally found in siltstone beds. All of the articulated fish found so far come from the dark gray to black portions of lake beds 2, 3, and 4 (Figure 15). By far the most common fish are semionotids of the "*Semionotus tenuiceps*" (Figure 8) and "small scale" groups of Olsen, *et al.* (1982). The fish average 7-15 cm in length but can attain sizes up to 40 cm. Much less common are the subholostean *Redfieldius* and the coelacanth *Diplurus*.

In the mid 1800's, Turners Falls was one of the most productive footprint localities in the Connecticut Valley, and was a favorite of Edward Hitchcock and fellow collectors James Deane, Dexter Marsh, Roswell Field, and Timothy Stoughton. Tracks are now uncommon on the mainland, but the islands in the river occasionally yield fine specimens. *In situ* footprints are most common in transgressive portions of Van Houten cycles; less distinct examples are also present in the red beds. The most common ichnotaxa are *Grallator* (*Eubrontes*) spp. and *Grallator* (*Anchisauripus*) spp., but *Anomoepus*, *Batrachopus* and *Otozoum* have been reported (Hitchcock, 1858). Unfortunately, even Hitchcock was sometimes not specific about the precise localities from which the tracks came. He often used the term Turners Falls for the entire stretch of exposures from Fall River to the present French King Bridge (Hitchcock, 1858).

**Location 1: Fall River beds and lower Deerfield Basalt.** Gray, tan, and red micaceous siltstones and sandstones comprise the upper 16 m of the Fall River beds at this locality (Figures 15 and 17). Lithologically, the sequence is unlike any known lower in the Sugarloaf Arkose and much more closely resembles the strata of the upper Shuttle Meadow Formation in the Hartford basin. Based on the cyclostratigraphy of the overlying Turners Falls Sandstone, this outcrop should be the time equivalent of the upper Shuttle Meadow Formation, the upper Feltville Formation of the Newark basin, and the upper Turkey Run Formation of the Culpeper basin (Olsen, *et al.*, 1989).

Stevens and Hubert (1980) described this section (Figure 16) and interpreted the thin, even beds of pyritic gray mudstone as the deepest water sediments of a perennial lake. We believe these beds are the marginal facies of a very shallow lake. The red units are also very shallow water deposits and contain caliche. According to Stevens and Hubert (1980) the lake originated as a playa and later expanded to a closed, alkaline perennial lake with substantial amounts of dissolved sodium, calcium, magnesium, and bicarbonate.

Abundant plants occur in muddy gray siltstone about 8.75 m below the Deerfield Basalt (Cornet, 1977a; Stevens and Hubert, 1980). *Equisetites* sp. is the most common plant, but large, three-dimensional individuals of *Clathropteris meniscoides* in growth position were found (Figure 6) in a fallen block from this zone in 1988.

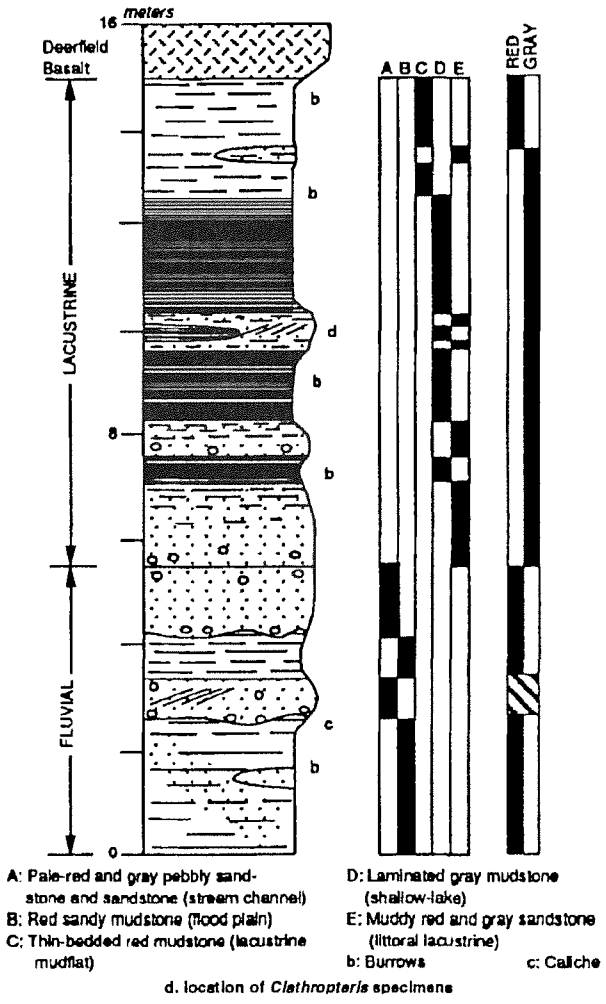



Figure 16: Measured section of the upper Sugarloaf Arkose, location 1 of Stop 4. Modified from Stevens and Hubert (1980).

The basal 17 m of the 23-m-thick lower flow complex of the Deerfield basalt consists of well-developed pillows which then pass upward into a 6-m-thick reddened vesicular zone, poorly exposed at this locality (Figure 16). The upper flow complex of the Deerfield Basalt is exposed in the woods to the southeast. The Deerfield Basalt is a high-titanium, quartz-normative basalt, compositionally similar to the Holyoke Basalt of the Hartford basin, Preakness Basalt of the Newark basin, and Sander Basalt of the Culpeper basin.

**Location 2: Bridge over Fall River and Fall River fault zone.** The offset in the basalt ridge is caused by the Fall River fault zone (Figure 14) that displaces the upper contact of the Deerfield Basalt about 200 m to the south. Displacement is probably due to normal faulting, as shown at location 9, a continuation of the same zone.

**Location 3: Road cut on Route 2 exposing Deerfield Basalt.** The top of the flow is visible at the east end of the outcrop (Figure 14), and the base can be seen on the east bank of the river in the woods to the northwest. The upper reddened zone of the lower flow complex is well displayed here, as is the upper flow complex. Several sill-like gabbroic zones 5-10 m thick are present; these are typical of the middle extrusive complex in the Newark Supergroup (Tollo, in Olsen, *et al.*, 1989). These gabbroids probably contain zircons or baddeleyites and should be amenable to U-Pb dating as has been done on the North Mountain Basalt in the Fundy basin of Nova Scotia (Hodych and Dunning, 1992).

**Location 4: Upper part of Deerfield Basalt and contact with Turners Falls Sandstone.** Proceed down the path on the basalt to the river's edge (Figure 14). Beds visible on the island along strike are 90 m are higher in section, and thus the Falls River fault zone must pass directly off the bank from basalt outcrop. Neptunian dikes and pockets of red mudstone are locally visible in the vesicular basalt along the contact.

**Location 5: Basal Turners Falls Sandstone and large clastic dike.** The red siltstones and fine sandstones of the Turners Falls Sandstone commonly show intense bioturbation by roots and burrows, some of which appear to be varieties of *Scoyenia* and *Planolites*. In the lower 70 m of the Turners Falls Sandstone, intervals of mudcracked massive mudstone alternate with lightly-burrowed to densely-burrowed and rooted strata, outlining a vague cyclic pattern. Some units show considerable relief and some of the sandy and gravelly units display dune-scale trough cross bedding. This section is a mixed fluvial-shallow water lacustrine system. The cyclic pattern, so obvious in overlying beds, is rendered somewhat obscure by the abundance of coarse-grained fluvial and deltaic strata in this dryier portion of a ~100,000 year cycle that is also in the drier part of a 413,000 year cycle.

A large clastic dike, cutting red sandstone and mudcracked massive mudstone, is present about 14 m above the base of the section. The dike (Figure 17) appears to have no connection with the beds above it or below it at this level of exposure. A bedding-parallel septum in the middle of the dike may be a remnant of a sand bed located down-dip, which was mobilized by the injection. The dike is segmented with mudstone septa, constricted between growing dike segments. The dike appears to be the result of lateral injection within mudstone units. A likely scenario is seismically-triggered fluidization of an unconsolidated sand; the sand was prevented from dewatering by enclosure in the less permeable mudstone. We discovered larger dikes of sandstone in lake bed 3 in the Turners Falls power canal when it was drained on July 31, 1992 (Figure 14).

**Location 6: Lake bed #0.** Lake bed #0 contains carbonate beds, concretions, and sparse organic matter in a fine-grained matrix. The limestone near the base of the gray interval (74.9 m above the base of formation) and a nodular calcareous siltstone about 30 cm higher have dissociated fish (*Semionotus* sp.). The weathered color of the fish bones is white and blue (from vivianite). A gray sandy siltstone just below the upper fish-bearing unit has abundant poor *Grallator* (*Anchisauripus*) spp.-type tracks. About 1 m above the base of the lower limestone is an upward-coarsening sequence of sandstone and siltstone with large, north-tilted surfaces. Individual sandstone beds thin down-the-paleoslope and pass into gray siltstone. The geometry suggests a small, shallow water delta. However, some groove marks and current lineations trend E-W, appropriate for a current tangential to the downstream portion of a point bar of a meandering stream. Along-strike color changes toward the river suggest oxidation or reduction reactions from circulating fluids. This same lake bed is exposed on the island in the middle of the Connecticut River to the immediate south.

**Location 7: Triplet of gray lacustrine cycles.** The prominent ledge-forming sandstone of the cycle containing lake bed #1 lies under the bridge. Lack of outcrop immediately above the sandstone marks the trace of the black shale, exposed only during very low water levels. The second and third Van Houten cycles (with lake beds 2 and 3 are well exposed, the third being at the foot of the dam. Reptile footprints occur in many of the transgressive and regressive gray sandstones and siltstones of all three cycles. Thick black shale with dolomitic laminae in lake beds #2, #3, and #4 contain fish. In the upper parts of the high stand portion of cycles #2 and #3 are abundant tan-weathering, septarian, dolomitic concretions probably related to incipient soil forming processes. Cycle #3 has deep, conglomerate-filled desiccation cracks where black shale is in contact with the overlying conglomerate bed.These same cycles are completely exposed in the Turners Falls power canal (Figure 14).

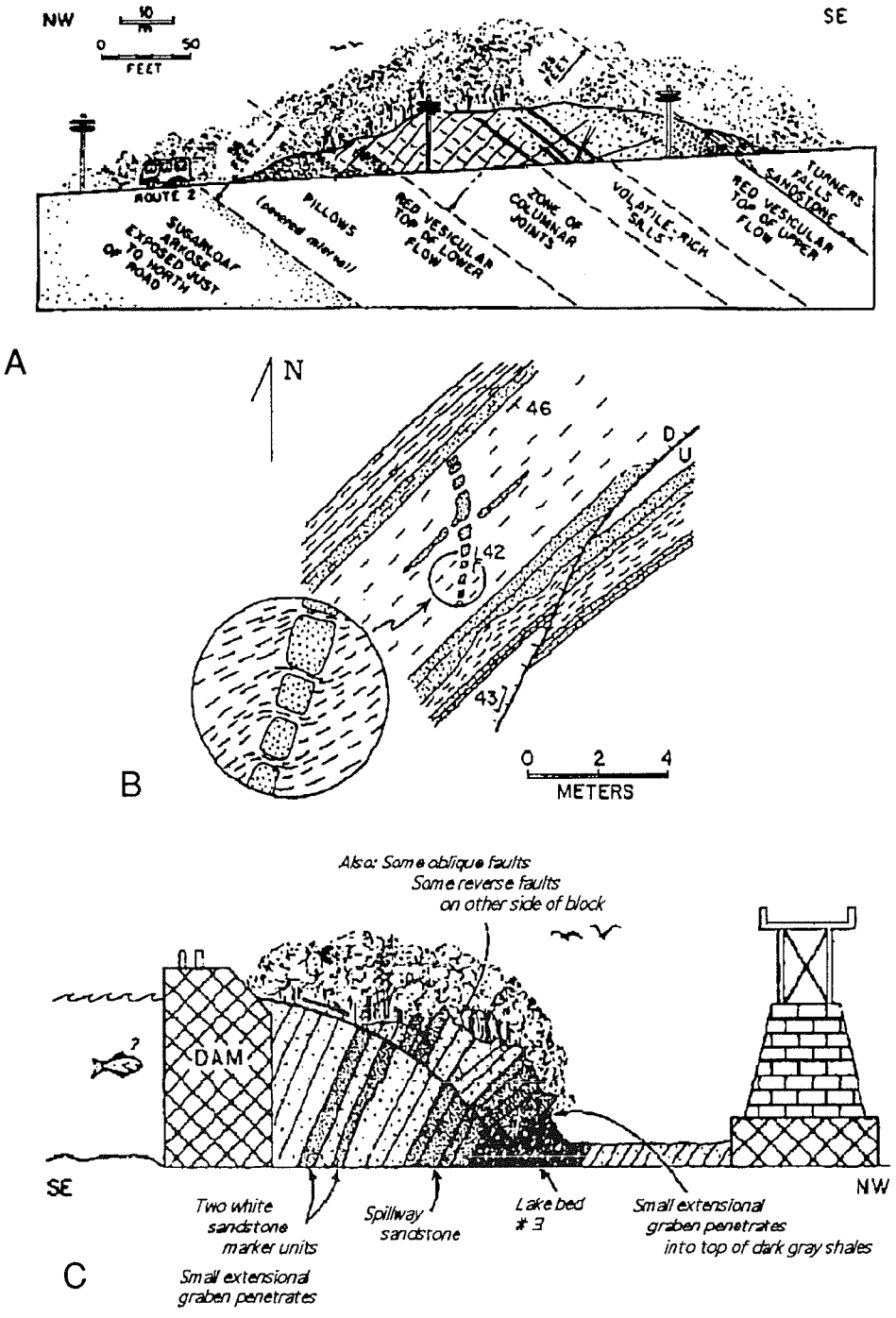



Figure 17. Details of Stop 4 at Turner's Falls. **A**, Diagram of outcrop along Route 2, location 3, showing contact between Deerfield Basalt and Turners Falls Sandstone. **B**, Clastic dike at location 5. **C**, Geologic relations on northern black of Spillway Island, location 8. All figures modified from Wise in Olsen *et al.* (1989).

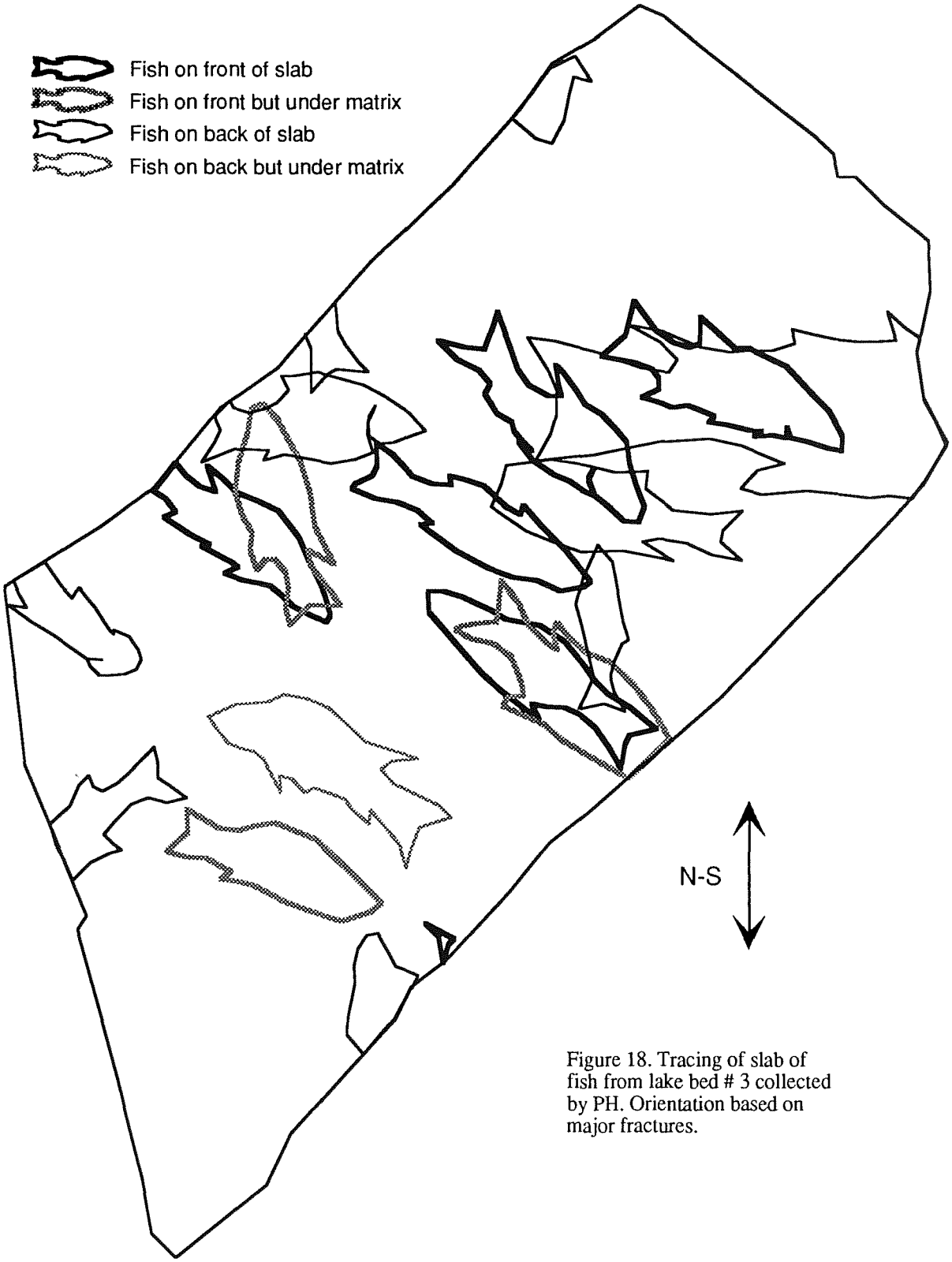



Figure 18. Tracing of slab of fish from lake bed # 3 collected by PH. Orientation based on major fractures.

Fish are most abundant and well preserved in lake bed #3, and occur throughout the lower three quarters of the unit. A silty limestone nodule bed occurs in the middle of the black shale, and every nodule contains either a fossil fish or a coprolite. At very low water, nodules washed out of this layer can be collected in the river bed near the dam. Fish are also extremely common in the microlaminated, crinkly beds, which contain many silt turbidites. Fish from at least some of the microlaminated beds seem to be oriented, with their long axes perpendicular to the main extensional joint set (Figure 18). The joint set trends NE-SW which is slightly oblique to the long axis of the Deerfield and Hartford basins. While there may be some component of tectonic shortening in the alignment of these fishes (Figure 18), most of the alignment was probably caused by delicate bottom currents. These currents could have been underflows running down the axis of the lake, or more likely internal seiches (very large scale, but delicate sloshing movements in stratified lakes).

Early folds with ~SE-NW axes occur in the lower three quarters of lake bed 3 and at the base of the shales there are transposed folds. These folds have an unknown effect on the thickness of the unit, and their origin is obscure. They are much more ductile appearing than the folds at Barton Cove (Stop 6), and hence appear to have formed earlier in the lithification history. Bedding plane faults and veins with fibrous calcite and bitumen are common in all of the black shale units. Displacement seems to be dip-parallel (as at Stop 6).

**Location 8: NE end of spillway island; lake bed #3.** The dam foundations are in the resistant spillway sandstone (Figure 17). Access to this island is difficult. It can be reached at low water by stepping over the rocks under the bridge, by wading across the river, or by seeking permission from Northeast Utilities to walk along the dam. The spillway sandstone also forms the resistant base for the north end of the dam and for the bridge pier to the south. Correlation with lake bed # 3 on the mainland is based on matching of thicknesses, lithology, deep clastic-infilled mudcracks, and septarian concretions comprised of dolomite (Wise, 1988).

The "island" is a fault block splay of the Falls River fault (Figures 14 and 17). The fault must pass very close to the bridge pier and very close to the south end of the spillway gates. At extreme low water, the disturbed beds are visible next to the gates. Much of the surface of the island is a large, curving, normal (?) fault surface. With present dips, the fault would be regarded as normal, but it is essentially perpendicular to bedding, and thus is a paleovertical structure. At the contact with the underlying dark shales, the fault and others parallel to it bend and drag the lake beds (Figure 19).

**Location 9: Lake bed #4.** Cross with care onto concrete the concrete apron of the dam. The largest of the Falls River fault splays passes under concrete here (Figure 15). Note breccia and mineralized fractures.

Lake bed #4 (Figures 13, 14, and 17) differs from those exposed on the mainland in having a more finely-laminated dolomitic unit and fewer interbedded turbidites. The transgressive sandstones below lake bed #4 have abundant *Eubrontes* footprints. Lower in the section are gray sandstones representing the wetter phases of two poorly developed Van Houten cycles (Figure 14) in the dry phase of the 413,000 year cycle. The upper cycle has well developed deltaic sequences, with thin, plant-rich gray siltstone intervals (?prodelta muds). The gray sandstones of the lower cycle have very well developed hummocky cross stratification with ~1 m crests suggesting water of considerable depth (>10 m).

At the west end of the island, lake bed #3 is exposed at the water line (Figure 17). It contains the characteristic fish and bedding style seen at locations 3 and 8.

Return to cars. Turn right onto Rt. 2 E.

| | |
|---|---|
| 24.8 | Intersection with Main Street and Turners Falls Bridge over the Connecticut River - go straight on Rt. 2 east. |
| 24.9 | Village of Riverside. |
| 25.0 | Red and gray conglomerates and sandstones are on left. |
| 25.7 | Entrance to Barton Cove on right (Stop 6). |
| 26.1 | Thin gray shales in red siltstones and sandstones on left. |
| 26.3 | Thin gray shales in red beds on left opposite parking area, this is Cornet's (1977a) locality TFSS-B (see discussion for Stop 6). |
| 26.6 | Thick gray siltstones and sandstones on left. |
| 26.8 | Gray, grading up into reddish sandstones and siltstones. |
| 27.2 | Good outcrops of gray and ?red conglomerates. |
| 28.2 | French King Bridge - Connecticut River. |
| 29.4 | Turn left onto "to Rt. 63 north" toward Northfield. |
| 29.6 | Turn right onto Rt. 63 north. |
| 31.2 | Entering Northfield. |
| 31.6 | Northfield Environmental Center on right. |
| 31.62 | Turn left on Lower Farms Rd. to Riverview Picnic area (opposite cemetery). |

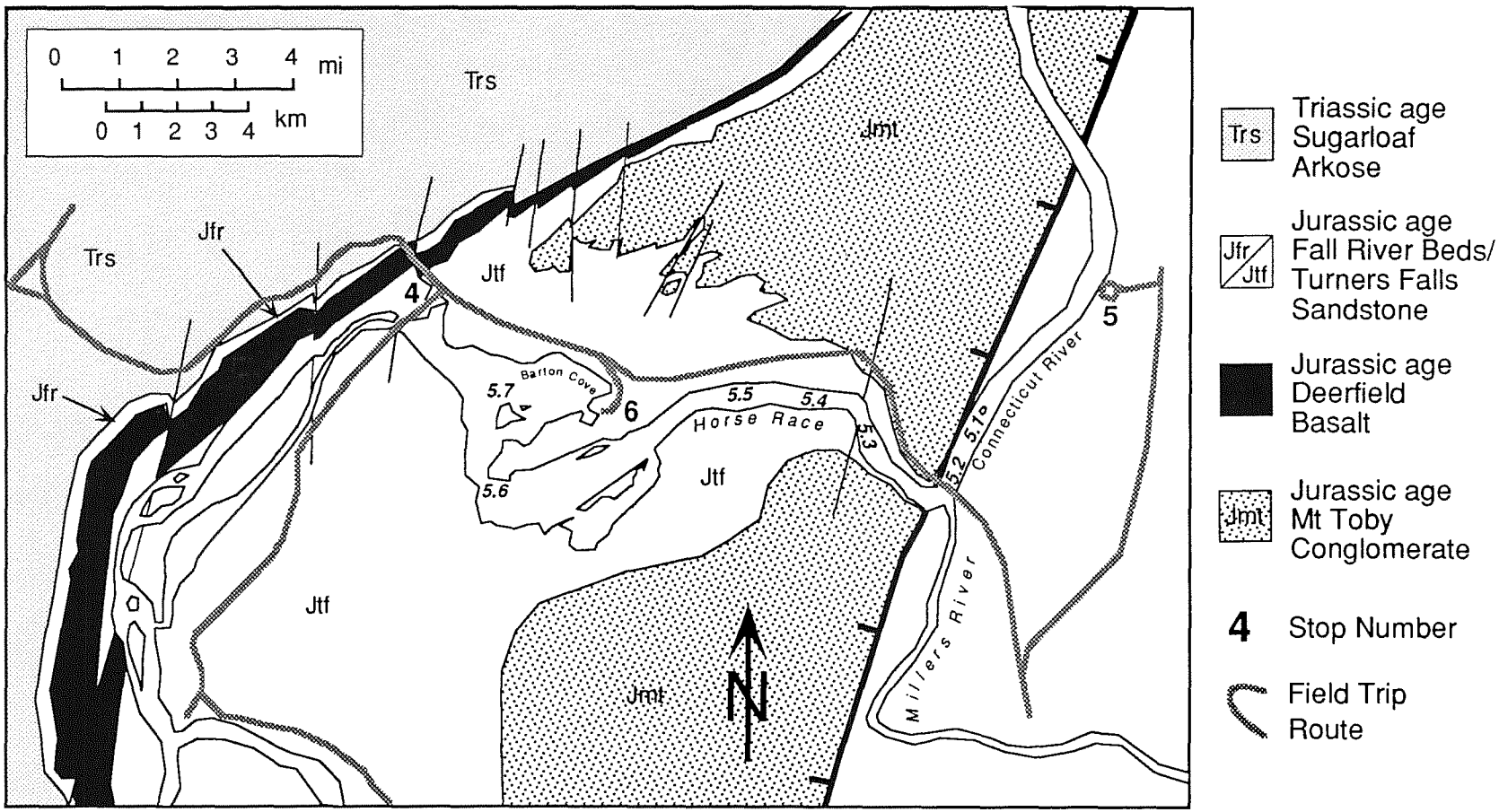



Figure 19. Map of the part of the northern Deerfield basin showing the distribution of units described in the text and the position of the field stops 4 and 5. Map modified from Willard (1951, 1952) and Stevens and Hubert (1980).

31.7 Drive through gates to circle and park.
Follow path to river and the boat, noting the blocks of Turners Falls Sandstone.

**STOP 5. RIVERBOAT TOUR OF CONNECTICUT RIVER ABOVE DAM AT TURNER'S FALLS.** Board the Quinnatucket II. The boat trip will go south and then east into the Deerfield basin to examine excellent (but mostly unstudied) outcrops of Turners Falls Sandstone along the river banks. Locations discussed in the text are shown on the map in Figure 19.

**Location 1: Border Fault Zone, French King Rock.** The River follows the border fault zone of the Deerfield basin in this area. On the right and left along the river are outcrops of sheared metamorphic rocks. On the left the seemingly bedded units are fault slivers of sheared and brecciated Giles Mt. slates and limestones. The fault separating Jurassic rocks of the Deerfield basin from these fault zone slivers is further up the hill (Stopen, 1988). The large, rounded mass of conglomerate in the center of the river is apparently a glacial erratic (Stopen, 1988; Peter Robinson, pers. comm.).

**Location 2: Conglomerate at Border Fault.** Mount Toby Conglomerate outcrops on right with some clasts in excess of 30 cm. Outcrops opposite Millers River consist of interbedded sandstones, shales, and conglomerates. An old quarry for footprints apparently existed on the hill on right. Just after the French King Bridge and Millers River, the Connecticut turns west, following close to the strike of the Turners Falls Sandstone. Note the proximity of fine grained beds to the present border fault. Was this the border fault at the time of deposition, or is the present fault just one of a broader zone that formed a pediment during deposition?

**Location 3: Fault zone in Turners Falls Sandstone.** Outcrops on both sides of the river are part of a fault zone that strongly deforms the adjacent strata. The beds are mostly lacustrine strata of Turners Falls Sandstone.

**Location 4: Lacustrine Cycles and Footprint and Insect Beds at the Horse Race.** Superb outcrops along the south bank of the river comprise a long, more or less continuous section of cyclical Turners Falls Sandstone (Figure 20). In older literature, this section is usually referred to as the Horse Race, Montague. According to Stoughton (1978), prior to the construction of the dam at Turners Falls in 1867, there was a stretch of swift water in this area called the "Horse Race" because the rocks protruding from the rapids appeared to be racing forward like a line of race horses while the water stood still.

An important footprint quarry operated by Timothy Stoughton from 1859 to 1867 (Stoughton, 1978) is present on the bluff to the south (Figure 20). The quarried beds are in the regressive portions of a very well developed Van Houten cycle that also produces insect fragments and conifers in the underlying, deeper water, dark gray shales. Footprints are present in most of the flaggy beds in the Horse Race section, and are absent only from the best-laminated dark gray or black shales.

Lower in the Horse Race section (Figure 20) there is an excellent outcrop of gray shales in a lacustrine cycle that produces fairly abundant *Mormolucoides*, occasional whole insects (Figure 7), and abundant conifer remains. Some thin beds with *Mormolucoides* also bear the clear impressions of cubic crystals (?halite). These crystals are preserved as sole marks and therefore must have formed very early (if not at the sediment water interface). Because some larvae are preserved with partially filled guts, they represent dead insects, not exuvia. It seems plausible that the larvae died as the waters became saturated with salt and the crystals began to grow. Reptile footprints have also been found on *Mormolucoides*-bearing surfaces indicating extremely shallow water. Another cycle, close to the base of the measured section (Figure 20) has produced *Mormolucoides* as well.

The average Van Houten cycle thickness in this area is 14 m, which is about the thickness of the cycles at both Turners Falls and Lily Pond, however, the range is 9 to 18 m (Figure 15, 20 and 21). There is a clear hierarchy of larger cycles with the ~100,000 year cycle being about 70 m and the ~400,000 year cycle being about 300 m. The upper 100 m of the section contains most of the gray sequences and represents the wetter part of the ~400,000 year cycle.

**Location 5: Lacustrine Cycles on North Bank.** Outcrops of two black shales on the north bank in this area mark out the wet phase of the next lower ~400,000 cycle from that at locality 4. Fragments of *Mormolucoides* and conifers occur in these shales.

**Location 6: The Narrows.** Excellent outcrops of two gray shale-bearing Van Houten cycles (Figures 19 and 21) are present on the southern tip of the peninsula jutting into the Connecticut River at this point. The upper cycle outcrops compeletely and has a deeper water portion riddled with dolomite pseudomorphs after aragonite (Rita Ricconi, pers. comm., 1992). The upper portion of the gray interval of this cycle contains numerous conifer fragments, and is folded and brecciated. The same cycles can be seen at Barton Cove (Stop 6) However, the deformation is not nearly as intense as seen at Stop 6.

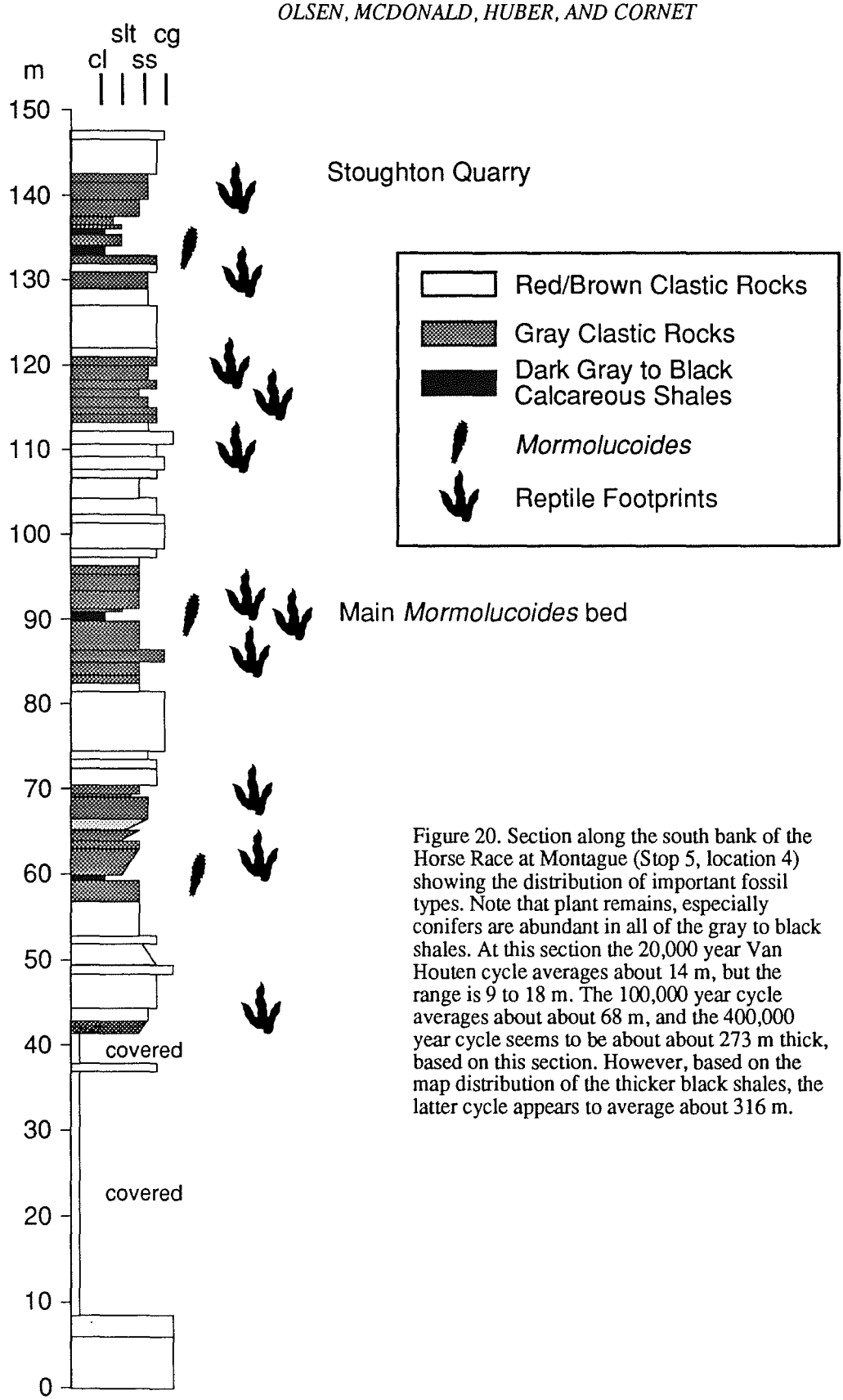



Figure 20. Section along the south bank of the Horse Race at Montague (Stop 5, location 4) showing the distribution of important fossil types. Note that plant remains, especially conifers are abundant in all of the gray to black shales. At this section the 20,000 year Van Houten cycle averages about 14 m, but the range is 9 to 18 m. The 100,000 year cycle averages about about 68 m, and the 400,000 year cycle seems to be about about 273 m thick, based on this section. However, based on the map distribution of the thicker black shales, the latter cycle appears to average about 316 m.

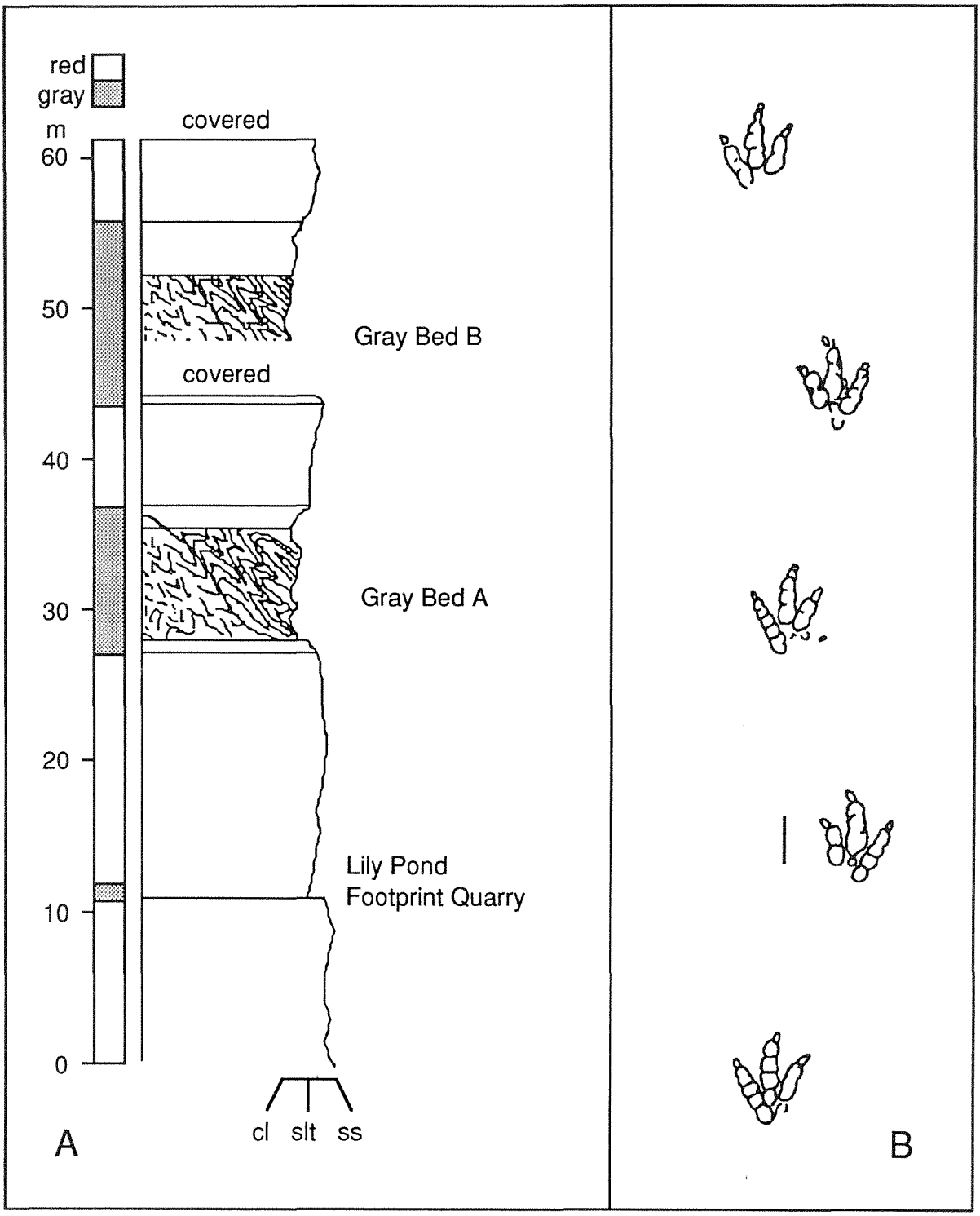



Figure 21. **A**, Section at Barton Cove (Stop 6) showing the distribution of cycles and deformed beds. **B**, Type specimen of *Anomoepus curvatus* (AC 52/10) from the Lily Pond Quarry at Barton Cove (scale is 5 cm).

**Location 7: Barton Island and Barton Cove.** Barton Island was once a hill surrounded by a low plain owned by Roswell Field. According to some maps (Beers, 1871), there was a footprint quarry on or near the island. In addition, we suspect that the location of the quarry at Field's Orchard in Gill was near here. Most outcrops visible from the boat are laminated Pleistocene lake deposits.

Note that this island is a protected bird sanctuary and cannot be accessed without special permission. American bald eagles nest on this island and are often visible, sometimes with young.

Outcrops on the north side of the river to the west included a productive track site near the location of the former "Red Suspension Bridge" (Stoughton, 1978). This was also the site of the old ferry across the river and the locality called "ferry above Turners Falls" is probably the same locality. Lull lists an inflated 17 genera and 27 species of footprints from this place.

Here the boat turns around and heads back to Northfield. After arriving at the dock, return to the cars and leave via Lower Farms Road.

32.15 Turn right back onto Rt. 63 south.
34.2 Bear right on Road to Rt. 2 west.
34.4 Turn Right onto Rt. 2 west.
35.5 French King Bridge.
36.0 Begin series of outcrops along Rt. 2 we passed on the way to Northfield.
37.8 Entrance to Barton Cove recreation area.
38.1 Drive along access road to parking lot and park.

**STOP 6. BARTON COVE.** The extensive exposures of gray and red lacustrine siltstones and sandstones of the Turners Falls Sandstone in the Barton Cove area reveal several Van Houten cycles. Two cycles display complex structural deformation, and another includes the classic Lily Pond footprint quarry (Figure 21).

Barton Cove occupies an area that was, prior to the construction of the new dam, a lowland with a number of small ponds. The ridge that underlies the peninsula forming the eastern boundary of Barton cove was a natural dam during the last deglaciation and the crescent shaped small coves along its western side were large plunge pools (Jefferson, 1898). The ridge itself is comprised of the lower parts of the wet portion of a ~400,000 year cycle and the upper parts of the drier portions of an underlying ~400,000 cycle.

Near Lily Pond, Handy (1976) identified ledge-forming zones of coarse breccia, composed of 2-100 cm blocks of gray and black to reddish dolomitic siltstone and fine sandstone in a muddy matrix of dark sandy siltstone. Handy mapped the distribution of these beds to southern Montague and thought they were produced by large scale sediment slumping. At Barton Cove these breccias occur in the gray portions of two thick (15 m) Van Houten cycles. Both clasts and matrix weather gray, yellow, and white. On a smaller scale, the breccia units are discontinuous and pass laterally into relatively undisturbed beds. Thrust faults and associated (in some places, recumbent) folds are present in beds below and adjacent to the breccia.

Both gray shale sequences show a variety of folding and faulting styles and intensities that suggest a sequence of progressive deformation (Figure 22). Relatively undeformed areas have kink-band-like trains of box and chevron folds which propagate upward through the entire thickness of gray shales within single cycles, dying out in the red units above and below. Thrust faults are present which have spatially associated chevron, box, and isoclinal folds. These faults tend to cut the entire gray shale sequence, but frequently pass into the surrounding red beds where they shallow and pass into bedding, where the folds disappear. The attitudes of the thrust faults and the axial planes of the fold trains tend to be parallel (Figure 22). In more strongly deformed areas, strata with abundant folds pass laterally into zones of transposed folds and chaotic breccia with a pseudostratigraphy (Figure 22). Most southwest trending outcrop surfaces of the breccia beds tend to show "beds" of imbricate clasts which alternate their direction, producing a faint herringbone pattern suggestive of transposed chevron folds. Outcrop surfaces trending northwest appear chaotic. There are rare folded faults within the most deformed units, but all structures are cut by NE-trending normal faults similar to those at Turners Falls (Olsen, *et al.*, 1989).

The orientation of the hinge lines of the folds and their northeast vergence is consistent with the orientation of the thrust faults and axial planes of the fold trains, all of which suggest northeast transport. The absence of evidence of post-deformation erosion or sedimentary draping of younger units on top of the deformed beds, suggests that the origin of these structures is due to between bed deformation, not slumping. If these structures are interpreted as slumps they are consistently heading up the axis of the basin, up the local dip, and away from the depocenter of the basin (Figure 23). This is not the direction of transport which would be expected of a slump sheet under any reasonable model. It is, however, exactly the direction of transport to be expected of thrust faults and folds generated by flexural slip folding of the main syncline which characterizes the Deerfield basin (Figure 23). We conclude that the deformation is tectonic and flexural slip in origin. While the inclination of the NE limb of the Deerfield basin

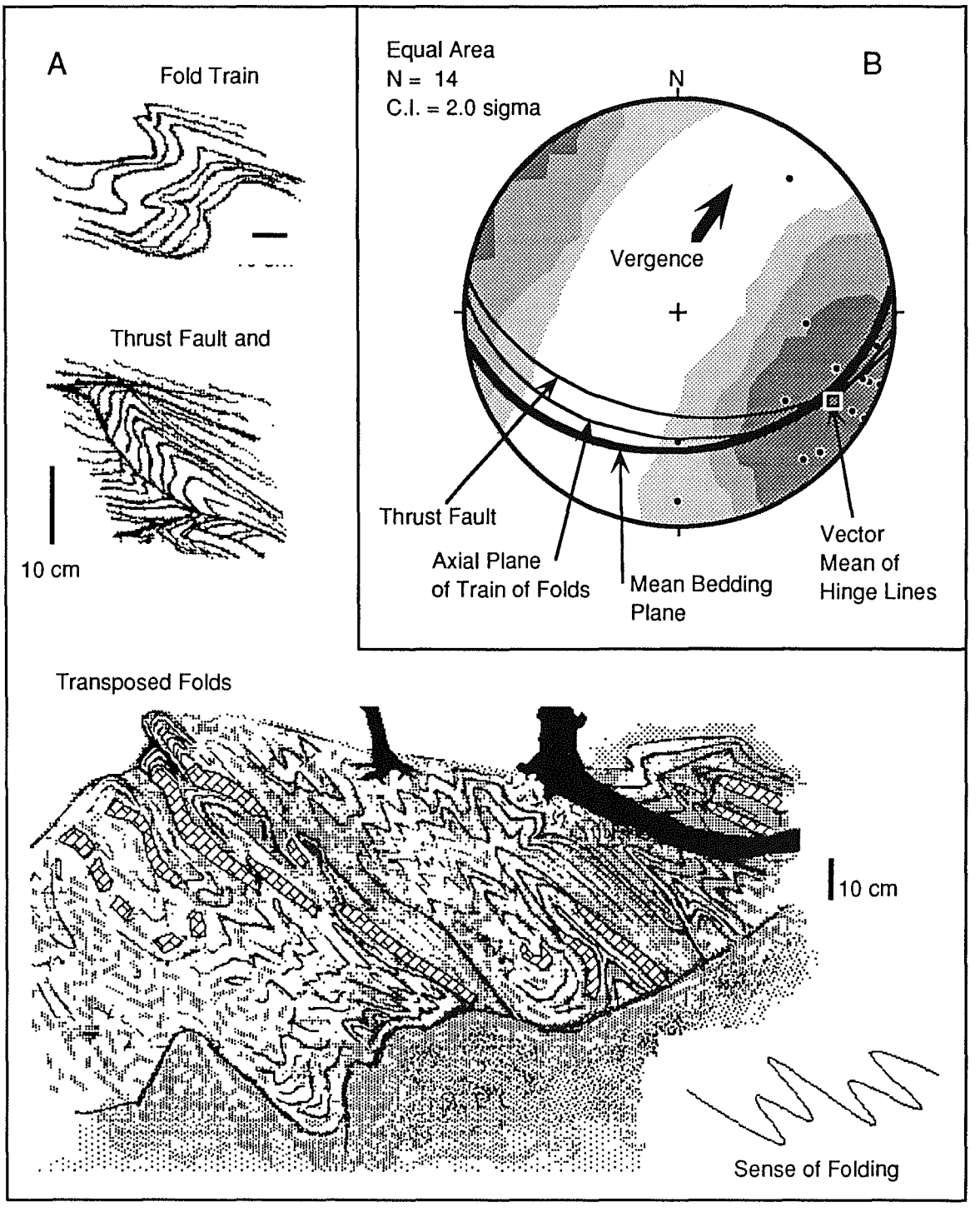



Figure 22. Deformation in the "slump beds" at Barton Cove (Stop 6). **A**, Exposures of deformed beds aligned parallel to the dip direction. **B**, Lower hemisphere projection of structures in the deformed beds.

syncline is relatively slight (~35°), the necessary amount of bedding plane shear can easily be achieved by concentrating most the the slip in a few beds within the basin section. Correspondingly, evidence of flexural slip is not obvious in surrounding strata of the Turners Falls - Barton Cove area.

Timing of the folding and associated flexural slip deformation is somewhat constrained by the NE-trending normal faults which cut the breccia. These faults tend to be mineralized. In contrast, the breccia and associated thrust faults and folds are not mineralized, suggesting early deformation, while the sediment was still somewhat plastic. Because the normal faults are probably synrift, so must be the flexural slip and folding. Paleomagnetic study of these structures could further constrain the timing of deformation, although this may be obscured by the multiple overprints which characterize the Connecticut Valley rocks (Brown, 1988; W. Witte, pers. comm.).

Cornet (1977a) recovered a palynoflorule ("LP") dominated by *Corollina meyeriana* from the breccia; a different assemblage dominated by *Corollina torosa* and *Araucariacites* spp. (TFSS-B) was recovered from what was thought to be an overlying undisturbed gray unit some distance away. This led Cornet to infer that the breccia represented a considerable hiatus, equivalent to the Shuttle Meadow through Hampden Basalt of the Hartford basin. In addition, Cornet (1977a) reasoned that these disturbed horizons might be continuous with thick (~78 m) metamorphic clast-bearing breccia beds within the Mt. Toby Conglomerate in the southern Deerfield. Robinson and Luttrell (1985) confused Cornet's descriptions of the Barton Cove beds in the northern Deerfield basin with the metamorphic clast breccia beds 7 km distant at Sunderland Cave (Emerson, 1898; Bain, 1932) in the southern Deerfield basin, implying that the latter lie above the former in the same outcrops (Robinson and Luttrell, 1985; p. A76), which they do not. Hence, they drew the base of the Mt. Toby Conglomerate and the top of the Turners Falls Sandstone, at the top of the "slump beds", which they regarded as an unconformity. This argument can be criticized on five grounds.

First, the so called "slump beds" are in our estimation early tectonic structures caused primarily by bedding plane slip, not slumps with an upper free surface. Second, it is far from clear that palynoflora TFSS-B comes from beds stratigraphically above the Barton Cove beds (palynoflora LC). The two localities are nearly 1 km apart, and the two intervals are not seen in superposition. Third, there are alternating zones dominated by *C. torosa* and *C. meyeriana* in the broadly correlative lower Portland Formation of the Hartford basin and upper Boonton Formation of the Newark basin in intervals not thought to contain a hiatus (Cornet (1977a). Fourth, the section above the Barton Cove "slump" along the Connecticut River (see Stop 5) is a continuation of the same kind of fine grained cyclical sequence represented by the Barton Cove sequence itself. Hence, there is no lithological break, and certainly no justification for a lithologically-based formational boundary. Fifth, there is no direct evidence that the lacustrine shales interbedded with conglomerate and metamorphic clast breccia at Mount Toby (Sunderland Cave) correlate with the Barton Cove beds. There are no metamorphic clast breccias or "fanglomerates" at Barton Cove. We conclude that there is no evidence of an unconformity or any significant break in sedimentation at or above the level of the Barton Cove "slump zone".

Throughout the Jurassic portion of the Deerfield basin section, finer grained strata, dominated by lacustrine shales, interfinger with and pass laterally into alluvial conglomerates. This can be seen on Mt. Toby at Sunderland Cave (Bain, 1932) where fine grained lacustrine shales with fragmentary to articulated fish (*Semionotus*) are interbedded with coarse conglomerates and metamorphic clast breccias. A similar relationship can be seen along Rt. 2 (between Stops 5 and 6), at Whitmore's Ferry and Chard Pond, along the Horse Race, and even adjacent to the border fault itself (Stop 5). Similar associations of alluvial and lacustrine beds have been described from the roughly contemporary Portland Formation of the Hartford basin (LeTourneau, 1985) and the Boonton Formation of the Newark basin (Olsen, 1980a). We therefore recognize the Turners Falls Sandstone and the Mt. Toby Conglomerate as lateral equivalents of one another, but still mappable formations. The boundary between the two formations can be drawn where conglomerates become dominant over fine grained clastic rocks.

In the mid 19th century the land now occupied by Barton Cove was largely owned by Roswell Field of Gill. He discovered and quarried several footprint localities on his property. One of the most prolific was opened in the thin-bedded, red-gray siltstones and sandstones that underlie the two gray sequences with the deformed beds (Figure 21). This quarry was opened next Lily Pond, and after Roswell Field, it was worked by Timothy Stoughton. It became a major source of many classic specimens for the Hitchcock collection, now housed in the Pratt Museum of Amherst College, Yale, and formerly Dartmouth and Princeton. A nearly complete suite of Connecticut Valley tracks was recovered from this quarry (Figure 21). Valid taxa present include *Batrachopus* spp., *Anomeopus* spp., and *Grallator* spp. (Olsen, *et al.*, 1989), although Lull (1915) lists a monumentally inflated 46 ichnospecies in 25 genera. Apparently most of the arthropod trackways described from the Deerfield basin come form these beds (Lull, 1953). We should note that Roswell Field (1860), who discovered this site, was one of the first to strongly suggest that the three-toed tracks were those of dinosaurs rather than birds, as Hitchcock thought.

Return to cars and leave Barton Cove.

38.4 Turn left onto Rt. 2 west.

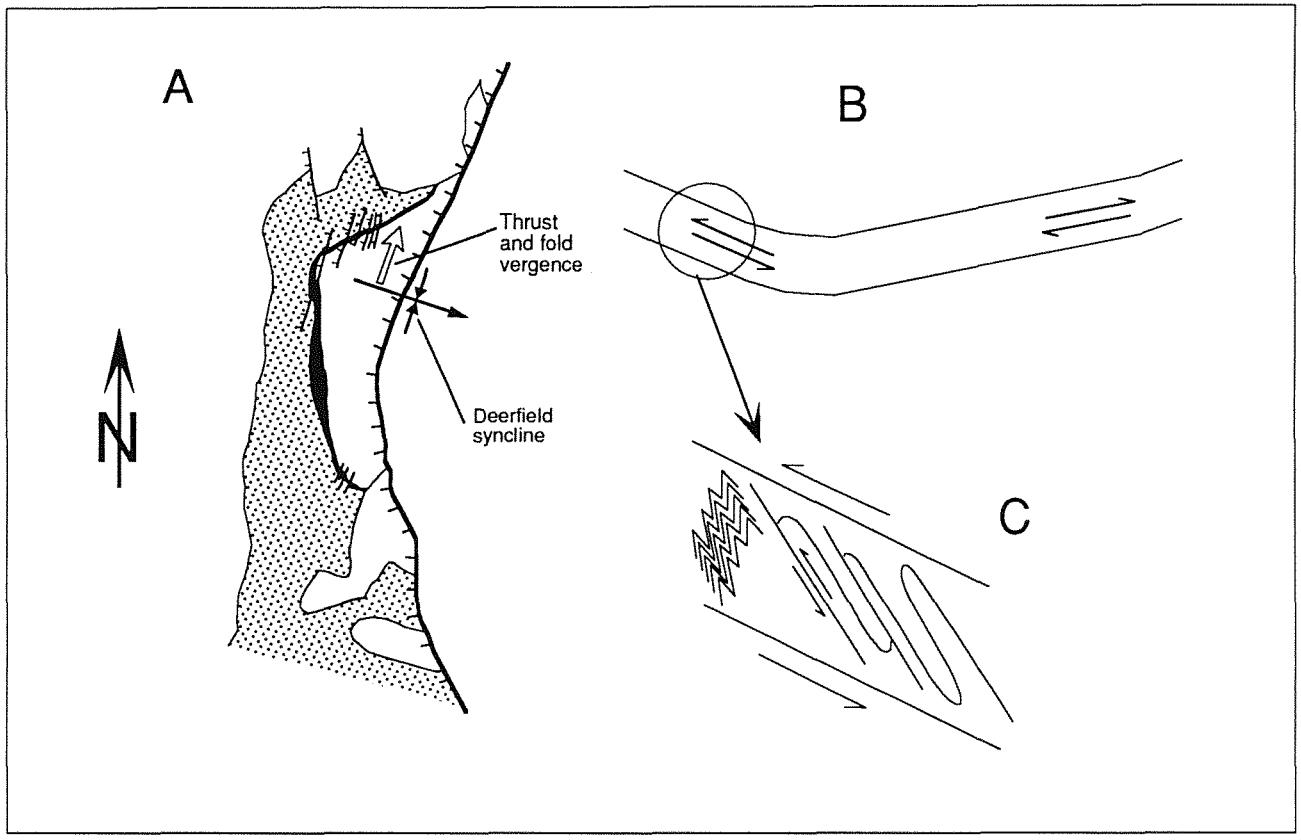



Figure 23. Possible origin of the deformed beds at Barton Cove (Stop 6) by flexural slip. **A**, Map of the Deerfield basin showing orientation of axis of main syncline. **B**, Flexural slip between two beds. **C**, Close up of deformation styles in zone of concentrated slip - note similarity to structures in Figure 22.

TABLE 1

Archeobacteria
- Amorphus algal debris

Plantae (leaf taxa only listed)
- Sphenopsida (horse tail rushes)
  - *Equisetites*
- Ficales (ferns)
  - *Clathropteris*
- Cycadales (cycad-like plants)
  - *Otozamites*
- Coniferales (conifers)
  - *Brachyphyllum*
  - *Pagiophyllum*

Animalia
- Arthropoda
  - ?Malacostraca
    - Crustacea
      - ?Decapoda
        - *Scoyenia*
  - Insecta
    - Coleoptera (beetles)
      - *Mormolucoides*
      - *Holcoptera*
- Chordata
  - Pisces
    - Osteichthyes (bony fishes)
      - Sarcopterygii (lobe-finned fishes)
        - Coelacanthiformes (coelacanths)
          - *Diplurus* cf. *longicaudatus*
      - Actinopterygii (ray-finned fishes)
        - Paleonisciformes (primitive ray-finned fishes
          - *Redfieldius* sp.
        - Semionotiformes (gars and short-snouted relatives)
          - *Semionotus* spp.
  - Tetrapoda (four legged vertebrates)
    - Amniota
      - Reptilia
        - Archosauria
          - Crurotarsi (Reptiles with a crocodile-like ankle)
            - Crocodiliomorpha
              - *Batrachopus* spp.
          - Dinosauria (Dinosaurs)
            - Saurischia (lizard-hipped dinosaurs)
              - Prosauropoda
                - *Otozoum* sp.
              - Theropoda
                - Grallatorids
                - *Grallator*
                - "*Anchisauripus*"
                - "*Eubrontes*"
                - indeterminate bone
            - Ornithischia
              - "Fabrosaurids"
                - *Anomoepus* spp.

40.5 Turn left onto Main Street over Turners Falls Bridge.
40.8 Note Turners Falls power canal on right (Figures 15 and 17).
41.0 Village of Turners Falls.
41.1 Approximate location of quarry that produced armored mud balls and possible theropod bone.
41.5 Outcrops on left of Turners Falls Sandstone.
43.1 Turn left onto Greenfield Road south.
43.4 The William Wilson quarry was located on hill to the left. Dexter Marsh recognized the presence of footprints here in 1835 (Stoughton, 1978). These were the first dinosaur tracks recognized in the Deerfield basin and they were passed along to James Deane. Dexter Marsh became an avid and important collector of footprints within the basin.
44.6 "Slump" beds similar to those outcropping at Barton Cove are exposed on left overlying black shales (Handy, 1976).
44.8 Outcrops of gray shales on right in stream bed contain *Mormolucoides* and conifers in a similar facies to that exposed along the Horse Race (Stop 5), but not in correlative beds.
45.5 Railroad overpass. Outcrops along the river to the north expose "slump beds" and dark gray shales containing *Mormolucoides* and conifers.
45.9 Veer right onto Greenfield Road Extension.
46.6 Turn Left onto South Ferry Road.
46.7 Turn right, staying on S. Ferry Road.
46.8 Keep right on S. Ferry Road.
46.9 Pass Taylor Hill Road on left. Taylor hill is to the south and is comprised of Mt. Toby Conglomerate. Its crest is marked by metamorphic clast breccia similar to that present at Whitmore Falls (see below). This was originally thought to be an inselberg of basement along a step fault (Emerson, 1898). However, drilling done in conjunction with the planning of the Northfield Mountain Pumping facility showed that the breccia is underlain by conglomerate and sandstone, not basement (Peter Robinson, pers. comm.).
48.6 Nice view of Mt. Toby straight ahead - S. Ferry Road becomes Meadow Road.
49.6 Sunderland town line. Meadow Rd. becomes North Sunderland Road. The site of Whitmore's ferry is on the left. Near here, along the river are exposures of the famous Sunderland fish bed. These gray, thin bedded siltstones are probably a lateral continuation of lake bed #3 exposed at Turner's Falls (Stop 4).
49.7 Whitmore falls on left exposes apparent basement. Higher up the hill and to the northeast, there are outcrops of breccia containing metamorphic clasts in Jurassic sedimentary matrix. The origin of these outcrops is highly controversial, with Emerson (1898) and Bain (1932) reasoning that these are outcrops of basement along step faults, while Peter Robinson (pers. comm.) argues they are talus deposits underlain by Turners Falls Sandstone .
50.2 Outcrops of lacustrine siltstones and associated conglomerates and sandstones that probably represent lake beds #1 and #2 at Turner's Falls (Stop 4).
50.5 Chard Pond.
51.2 Conglomerate outcrops on left are in the lower beds of the Turners Falls Sandstone.
51.3 Turn right onto Rt. 47 South (North Main Street).
52.7 This small ridge is Deerfield Basalt.
56.5 Turn left onto Rt. 116 south at Sunderland Village center.
57.4 Intersection with Rt. 63, go straight.
58.8 Veer right onto ramp for exit for University of Massachusetts, Amherst.
59.15 At top of ramp turn left onto Massachusetts Ave.
60.4 Go straight ahead at light onto North Pleasant Street.
60.7 Turn right continuing on North Pleasant.
60.9 Amherst village center.
61.1 Turn left into Amherst College.
61.4 Pratt Museum of Amherst College. End of field trip.

## BIBLIOGRAPHY

Agassiz, L., 1833-1843, Recherches sur les poissons fossiles. Neuchâtel, Switzerland, Imprimerie de Petitpierre, 5 vols., 1420 p.

Bain, G.W., 1932, The northern area of the Connecticut Valley Triassic. American Journal Science, ser. 5, v. 23, p. 57-77.

Baird, D., 1957, Triassic reptile footprint faunules from Milford, New Jersey. Museum Comparative Zoology Bulletin, v. 117, p. 447-520.

Baird, D., 1980, A prosauropod dinosaur trackway from the Navajo Sandstone (Lower Jurassic) of Arizona. In Jacobs, L.L., (ed.), Aspects of Vertebrate History - Essays in honor of Edwin Harris Colbert. Flagstaff, Ariz., Museum of Northern Arizona Press, p. 219-230.

Beers, F.W., 1871, Atlas of Franklin County, Massachusetts.

Belt, E.S., 1989, A brief sketch of Edward Hitchcock (1793-1864); In Jordan, W.M., (ed.), Boston to Buffalo, in the footsteps of Amos Eaton and Edward Hitchcock. 28th International Geological Congress, Washington, D.C., Field Trip Guidebook T169, p. 14-20.

Bice, D.M., Newton, C.R., McCauley, S., Reiners, P.W., McRoberts, C.A., 1992, Shocked quartz at the Triassic-Jurassic boundary in Italy. Science, V. 255, p. 443-446.

Bock, W., 1949, Triassic chimaeroid egg capsules from the Connecticut Valley. Journal of Paleontology. v. 23, p. 515-517.

Bowlds, L. S., 1989, Tracking down the Early Permian. Geotimes, May, 1989, p. 12-14.

Brown, L.L., 1988, Multicomponent paleomagnetic directions from the Sugarloaf arkose, Deerfield basin, Massachusetts. Geological Society of America Abstracts. v. 20, no. 1, p. 10.

Chandler, W.E., Jr., 1978, Graben Mechanics at the Junction of the Hartford and Deerfield basins of the Connecticut Valley, Massachusetts. M.S. Thesis, Contribution no. 331, Deptartment of Geology and Geography, University of Massachusetts, Amherst, MA, 151 p.

Colbert, E.H., 1961, Dinosaurs - Their Discovery and Their World. New York, N.Y., E.P. Dutton and Co., 300 p.

Cornet, B., 1977a, The Palynostratigraphy and Age of the Newark Supergroup. Ph.D. Thesis, Deptartment of Geology, The Pennsylvania State University, State College, PA, 506 p.

Cornet, B., 1977b. Preliminary investigation of two Late Triassic conifers from York County, Pennsylvania. In Romans, R.C. (ed.), Geobotany, Plenum Press, New York. p.165-172.

Cornet, B., 1989a, Late Triassic angiosperm-like pollen from the Richmond rift basin of Virginia. U.S.A. Palaeontographica, v. 313B, p. 37-87.

Cornet, B., 1989b, The reproductive morphology and biology of *Sanmiguelia lewisii*, and its bearing on angiosperm evolution in the Late Triassic. Evolutionary Trends in Plants, v. 3, no. 1, p. 25-51.

Cornet, B. and Olsen, P.E., 1985, A summary of the biostratigraphy of the Newark Supergroup of eastern North America with comments on early Mesozoic provinciality. In Weber, R. (ed.), III Congresso Latinoamericano de Paleontologia. Mexico. Simposio Sobre Floras del Triasico Tardio, su Fitogeografia y Paleoecologia, Memoria. p. 67-81.

Cornet, B. and Traverse, A., 1975, Palynological contributions to the chronology and stratigraphy of the Hartford Basin in Connecticut and Massachusetts. Geoscience and Man, v. 11, p. 1-33.

Crane, P.R., 1987, Vegetational consequences of the angiosperm diversification, In Friis, E.M., Chaloner, W.G., and Crane, P.R. (eds.), The origins of angiosperms and their biological consequences. Cambridge University Press, Cambridge, p. 107-114.

Dana, J.D., 1858, Note on fossil insect remains from Turners Falls, Massachusetts. In Hitchcock, E., Ichnology of New England, etc. Boston, MA., Commonwealth of Mass., p. 7-8.

Deane, J., 1844, On the discovery of fossil footmarks. American Journal of Science, ser. 1, v. 47, p. 381-390.

Deane, J., 1861, Ichnographs from the sandstone of Connecticut River. Boston, MA., Little Brown and Co., 61 p.

DeBoer, J.Z. and Clifton, A.E.,1988, Mesozoic tectogenesis: development and deformation of 'Newark' rift zones in the Appalachians (with special emphasis on the Hartford basin, Connecticut). In Manspeizer, W. (ed.), Triassic-Jurassic rifting: continental breakup and the origin of the Atlantic Ocean and Passive Margins. Amsterdam, Netherlands, Elsevier Sci. Publ., part A, chapter 11, p. 275-306.

Dunning, G.R. and Hodych, J.P., 1990, U-Pb zircon and baddeleyite age for the Palisade and Gettysburg sills of northeast United States; implications for the age of the Triassic-Jurassic boundary. Geology, v. 18, p. 795-798.

Eastman, C.R., 1911, Triassic fishes of Connecticut. Connecticut Geological and Natural History Survey Bulletin no. 18, 77 p.

Echelle, A.A. and Kornfield, I. (eds.), 1984, Evolution of fish species flocks. Orono, Maine, Univ. of Maine at Orono Press, 257 p.

Emerson, B.K., 1898, Geology of Old Hampshire County, Massachusetts, comprising Franklin, Hampshire and Hampden Counties. U.S. Geological Survey Monograph 29, 790 p.

Emmons, E., 1857, American Geology, etc., part VI. Permian and Triassic Systems of the Atlantic Slope. Albany, N.Y., Sprague and Co., 152 p.

Galton, P.M., 1976, Prosauropod dinosaurs (Reptilia: Saurischia) of North America. Postilla, no. 169, 98 p.

Gauthier, J.A., Kluge, A.G., and Rowe, T., 1988, The early evolution of the Amniota. In Benton, M.J. (ed.), The Phylogeny and Classification of the Tetrapods, Systematics Association Special Volume No. 35A, Clarendon Press, Oxford, pp. 103-135.

Gilmore, G.E., 1928, Fossil footprints from the Grand Canyon; Third Contribution . Smithsonion Miscellaneous Collections, v. 80, p. 1-16.

Handy, W.A., 1976, Depositional History and Diagenesis of Lacustrine and Fluvial Sedimentary Rocks of the Turners Falls and Mount Toby Transition, North-Central Massachusetts. M.Sci. Thesis, Department of Geology, University of Massachusetts, Amherst, MA, 115 p.

Hitchcock, E., 1818, Remarks on the geology and mineralogy of a section of Massachusetts on Connecticut River, with a part of New Hampshire and Vermont. American Journal of Science, ser. 1, v. 1, p. 105-116.

Hitchcock, E., 1823, A sketch of the geology, mineralogy and scenery of the region contiguous to the River Connecticut, etc.American Journal of Science, ser. 1, v. 6, p. 1-86.

Hitchcock, E., 1833, Report on the geology, mineralogy, botany and zoology of Massachusetts. Amherst, Mass., Commonwealth of Mass., 692 p.

Hitchcock, E., 1836, Ornithichnology - description of the footmarks of birds (ornithichnites) on New Red Sandstone in Massachusetts. American Journal of Science, ser. 1, v. 29, p. 307-340.

Hitchcock, E., 1837, Fossil footsteps in sandstone and graywacke. American Journal of Science, ser. 1, v. 32, p. 174-176.

Hitchcock, E., 1841, Final Report on the Geology of Massachusetts. Amherst, MA, Commonwealth of Massachusetts, 831 p.

Hitchcock, E., 1845, An attempt to name, classify and describe the animals that made the fossil footmarks of New England. Association of American Geologists and Naturalists, Proceedings., 6th Mtg., p. 23-25.

Hitchcock, E., 1847, Description of two new species of fossil footmarks found in Massachusetts and Connecticut, or of the animals that made them. American Journal of Science, ser. 2, v. 4, p. 46-57.

Hitchcock, E., 1858, Ichnology of New England. A report on the sandstone of the Connecticut Valley, especially its fossil footmarks. Boston, MA., Commonwealth of Massachusetts, 220 p.

Hitchcock, E., 1865, Supplement to the ichnology of New England [with editorial notes by C.H. Hitchcock]. Boston, Mass., Commonwealth of Massachusetts., 96 p.

Hodych, J.P. and Dunning, G.R., 1992, Did the Manicouagan impact trigger end-of-Triassic mass extinction? Geology, v. 20, p. 51-54.

Huber, P. and McDonald, N.G., In Press, Revised stratigraphy and paleontology of the early Mesozoic Pomperaug basin (Newark Supergroup), western Connecticut. Geological Society of America, Abstracts with Programs.

Huber, P., Lucas, S. G. and Hunt, A. P. In Press, Late Triassic fish assemblages of the North American Western Interior and their biochronologic significance; In Morales, M., (ed.)., Museum of Northern Arizona Bulletin.

Hunt, A.P., Lucas, S. G. and Huber, P., 1990, Early Permian footprint fauna from the Sangre de Cristo Formation of northeastern New Mexico. New Mexico Geological Society, Guidebook 41, pp. 291-303.

Jefferson, M.S.W., 1898, The postglacial Connecticut at Turners Falls, Massachusetts. Journal of Geology, v. 6, p. 463-472.

Kutzback, J.E. and Gallimore, R.G.,1988, Pangean climates: Megamonsoons of the megacontinent. Journal of Geophysical Research, v. 94, p. 3341-3357.

LeTourneau, P.M., 1985, The Sedimentology and Stratigraphy of the Lower Jurassic Portland Formation, central Connecticut. M.A. Thesis, Department of Earth and Environmental Sciences, Wesleyan Univ., Middletown, Conn., 247 p.].

Little, R.D., 1982, Lithified armored mud balls of the Lower Jurassic Turners Falls Sandstone, north-central Massachusetts. Journal of Geology, v. 90, p. 203-207.

Lull, R.S., 1904, Fossil footprints of the Jura-Trias of North America. Boston Society of Natural History Memoirs, v. 5, p. 461-557.

Lull, R.S., 1915, Triassic life of the Connecticut Valley. Connecticut Geological and Natural History Survey Bulletin no. 24, 285 p.

Lull, R.S., 1953, Triassic life of the Connecticut Valley (revised). Connecticut Geological and Natural History Survey Bulletin no. 81, 336 p.

MacFadyen, J.A., Thomas, J.E., Jr. and Washer, E.M., 1981, A new look at the Triassic border fault of northcentral Massachusetts. In O'Leary, D.W. and Earle, J.L. (eds.), Proceedings of the Third International Conference on Basement Tectonics. Denver, Colo., Basement Tectonics Committee, Publ. no. 3, p. 363-373.

Manspiezer, W., 1988, Triassic-Jurassic rifting and opening of the Atlantic: an overview. In Manspeizer, W. (ed.), Triassic-Jurassic rifting: continental breakup and the origin of the Atlantic Ocean and Passive Margins. Amsterdam, Netherlands, Elsevier Sci. Publ., part A, chapter 3, p. 41-79.

Marsh, O.C., 1893, Restoration of *Anchisaurus*. American Journal of Science, ser. 3, v. 45, p. 169-170.

McCune, A.R., 1982, Early Jurassic Semionotidae (Pisces) from the Newark Supergroup: Systematics and evolution of a fossil species flock. Ph.D. Thesis, Department of Biology, Yale University, New Haven, CT, 371 p.].

McCune, A.R., 1986, A revision of *Semionotus* (Pisces: Semionotidae) from the Triassic and Jurassic of Europe. Palaeontology, v. 29, part 2, p. 213-233.

McCune, A.R., 1987a, Lakes as laboratories of evolution: endemic fishes and environmental cyclicity. Palaios, v. 2, p. 446-454.

McCune, A.R., 1987b, Milankovitch cycles and endemism: evolution of *Semionotus* (Pisces) in Mesozoic lakes [abstr.]. Journal of Vertebrate Paleontology, v. 7, no. 3, suppl., p. 20A-21A.

McCune, A.R., 1990, Evolutionary novelty and atavism in the *Semionotus* complex: relaxed selection during colonization of an expanding lake. Evolution, v. 44, p. 71-85.

McCune, A.R., Thomson, K.S. and Olsen, P.E., 1984, Semionotid fishes from the Mesozoic great lakes of North America. *In* Echelle, A.A. and Kornfield, I. (eds.), Evolution of Fish Species Flocks. Orono, Maine, University of Maine at Orono Press, p. 27-44.

McDonald, N.G., 1975, Fossil Fishes from the Newark Group of the Connecticut Valley. M.A. Thesis, Department of Geology, Wesleyan University, Middletown, CT, 230 p.

McDonald, N.G., 1982, Paleontology of the Mesozoic rocks of the Connecticut Valley. In Joesten, Raymond and Quarrier, S.S. (eds.), Guidebook for field trips in Connecticut and south-central Massachusetts; New England Intercollegiate Geological Conference, 74th Annual. Meeting., Storrs, Connecticut. Connecticut Geological and Natural History Survey Guidebook, no. 5, trip M-2, p. 143-172.

McDonald, N.G., 1992, Paleontology of the early Mesozoic (Newark Supergroup) rocks of the Connecticut Valley. Northeastern Geology, v. 14, p. 185-199.

McEnroe, S.A., 1989, Paleomagnetism and Geochemistry of Mesozoic Diabase Dikes and Sills in West-Central Massachusetts. M.S. Thesis, Department of Geology and Geography, University of Massachusetts, Amherst, MA, 169 p..

Moodie, R. L., 1930, Vertebrate footprints from the red beds of Texas II. Journal of Geology, v. 38, p. 548-565.

Newberry, J.S., 1887, The fauna and flora of the Trias of New Jersey and the Connecticut Valley. New York Academy of Science, Transactions, v. 6, p. 124-128.

Newberry, J.S., 1888, Fossil fishes and fossil plants of the Triassic rocks of New Jersey and the Connecticut Valley. U.S. Geological Survey Monograph 14, 152 p.

Olsen, P.E., 1977, Stop 1 - Triangle Brick Quarry. In G.L. Bain and B.W. Harvey (eds.), Field Guide to the Geology of the Durham Triassic Basin, Raleigh. Carolina Geological Society p. 59-60.

Olsen, P.E., 1980a, The Latest Triassic and Early Jurassic formations of the Newark basin (Eastern North America, Newark Supergroup): stratigraphy, structure, and correlation. New Jersey Academy of Sciences, Bulletin, v. 25, p. 25-51.

Olsen, P.E., 1980b, Fossil great lakes of the Newark Supergroup in New Jersey. In W. Manspeizer (ed.) Field Studies in New Jersey, Geology and Guide to Field Trips, 52nd Annual Meeting, New York State Geological Association, Newark College of Arts and Sciences, Newark, Rutgers University, p. 352-398.

Olsen, P.E., 1980c, Comparison of the vertebrate assemblages from the Newark and Hartford basins (Early Mesozoic, Newark Supergroup) of eastern North America. In Jacobs, L.L. (ed.), Aspects of Vertebrate History, Flagstaff Museum of Northern Arizona Press, p. 35-53.

Olsen, P.E., 1983, On the non-correlation of the Newark Supergroup by fossil fishes: biogeographic, structural and sedimentological implications. Geological Society of America, Abstracts with Programs, v. 15, no. 3, p. 121.

Olsen, P.E., 1986, A 40-million-year lake record of early Mesozoic orbital climatic forcing. Science, v. 234, p. 842-848.

Olsen, P.E., 1988, Paleontology and paleoecology of the Newark Supergroup (early Mesozoic, eastern North America). In Manspeizer, W. (ed.), Triassic-Jurassic Rifting: Continental Breakup and the Origin of the Atlantic Ocean and Passive Margins, Amsterdam, Netherlands, Elsevier Scientific Publications, part A, chapter 8, p. 185-230.

Olsen, P.E. and Baird, D., 1986, The ichnogenus *Atreipus* and its significance for Triassic biostratigraphy. In Padian, K. (ed.), The Beginning of the Age of Dinosaurs: Faunal Change Across the Triassic-Jurassic Boundary. Cambridge, England, Cambridge University Press, p. 61-87.

Olsen, P.E., Fowell, S.H., and Cornet, B., 1990, The Triassic/Jurassic boundary in continental rocks of eastern North America; a progress report. Geological Society of America, Special Paper 247, p. 585-593

Olsen, P.E. and Kent, D., 1990, Continental Coring of the Newark Rift. EOS, Transactions of the American Geophysical Union, 71, 385, 394.

Olsen, P.E. and McCune, A.R., 1991, Morphology of the *Semionotus elegans* species group from the Early Jurassic part of the Newark Supergroup of eastern North America with comments on the family Semionotidae (Neopterygii). Journal of Vertebrate Paleontology, v. 11, no. 3, p. 269-292.

Olsen, P.E., McCune, A.R., and Thomson, K.S., 1982, Correlation of the early Mesozoic Newark Supergroup by vertebrates, principally fishes. American Journal of Science, v. 282, p. 1-44.

Olsen, P.E. and Padian, K., 1986, Earliest records of Batrachopus from the Southwest U.S., and a revision of some Early Mesozoic crocodilomorph ichnogenera. In K. Padian (ed.), The Beginning of the Age of Dinosaurs, Faunal Change Across the Triassic-Jurassic Boundary, Cambridge University Press, New York, p. 259-273

Olsen, P.E., Schlische, R.W. and Gore, P.J.W. (eds.), 1989, Tectonic, Depositional, and Paleoecological History of Early Mesozoic Rift Basins, Eastern North America. 28th International Geological Congress, Washington, D.C., Field Trip Guidebook T351, 174 p.

Olsen, P.E., Shubin, N.H. and Anders, P.E., 1987, New Early Jurassic tetrapod assemblages constrain Triassic-Jurassic tetrapod extinction event. Science, v. 237, p. 1025-1029.

Olsen, P.E., Withjack, M.O., and Schlische, R.W., in press, Inversion as an Integral Part of Rifting: An Outcrop Perspective from the Fundy Basin, Eastern North America, EOS, Transactions of the American Geophysical Union.

Padian, K. and Olsen, P.E., 1984, The fossil trackway *Pteraichnus*: not pterosaurian, but crocodilian. Journal of Paleontology., v. 58, p. 178-184.

Pettijohn, F. J. and Potter, P. E., 1964, Atlas and glossary of primary sedimentary structures. Berlin, Springer-Verlag, 370 pp.

Philpotts, A.R. and Reichenbach, I., 1985, Differentiation of Mesozoic basalts of the Hartford basin, Connecticut. Geological Society of America, Bulletin, v. 96, p. 1131-1139.

Pratt, L.M., Shaw, C.A. and Burruss, R.C., 1988, Thermal histories of the Hartford and Newark basins inferred from maturation indices of organic matter. In Froelich, A.J. and Robinson, G.R., Jr. (eds.), Studies of the early Mesozoic basins of the eastern United States. U.S. Geological Survey Bulletin, no. 1776, p. 58-63.

Redfield, J.H., 1837, Fossil fishes of Connecticut and Massachusetts, with a notice of an undescribed genus. N.Y. Lyceum Natural History, Transactions, v. 4, p. 35-40.

Redfield, J.H., 1845, Catalogue of the fossil fish of the United States as far as known, with descriptions of those found in the New Red Sandstone. unpublished report, Peabody Museum, Yale University, New Haven, Conn., 19 p.

Redfield, W.C., 1841, Short notices of American fossil fishes. American Journal of Science, ser. 1, v. 41, p. 24-28.

Robinson, P. and Luttrell, G.W., 1985, Revision of some stratigraphic names in central Massachusetts. U.S. Geological Survey Bulletin, no. 1605-A, p. A71-A78.

Roden, M.K. and Miller, D.S., 1991, Tectono-thermal history of Hartford, Deerfield, Newark and Taylorsville basins, eastern United States, using fission-track-analysis. Schweiz. Mineral. Petrogr. Mitt., v. 71, p. 187-203.

Rogers, H.D., Vanuxem, L., Taylor, R.C., Emmons, E. and Conrad, T.A., 1841, Report on the ornithichnites or foot marks of extinct birds in the New Red Sandstone of Massachusetts and Connecticut, observed and described by Prof. Hitchcock, of Amherst. American Journal of Science, ser. 1, v. 41, p. 165-168.

Schlische, R. W., 1990, Aspects of the Structural and Stratigraphic Evolution of Early Mesozoic Rift Basins of Eastern North America. Ph.D. Thesis, Department of Geology, Columbia University, 479 p.

Schlische, R.W. and Olsen, P.E., 1990, Quantitative filling model for continental extensional basins with applications to early Mesozoic rifts of eastern North America. Journal of Geology, v. 98, p. 135-155.

Scudder, S.H., 1886, The oldest known insect larva, *Mormolucoides articulatus*, from the Connecticut River rocks. Boston Society of Natural History Memoir, v. 3, p. 431-438.

Silvestri, S.-M. and Olsen, P.E., 1989, Ichnostratigraphy of the Jacksonwald Syncline: the last 7 million years of the Triassic. Geological Society of America, Abstracts with Programs, v. 21, no. 2, p. 66.

Steinbock, R.T., 1989, Ichnology of the Connecticut Valley: A vignette of American science in the early nineteenth century. In Gillette, D.D. and Lockley, M. (eds.) Dinosaur Tracks and Traces. Cambridge, England, Cambridge Univ. Press, p. 27-32.

Stevens, R.L. and Hubert, J.F., 1980, Alluvial fans, braided rivers and lakes in a fault-bounded semiarid rift valley: Sugarloaf Arkose (Late Triassic-Early Jurassic), Newark Supergroup, Deerfield basin, Massachusetts. Northeastern Geology, vol. 2, p. 100-117.

Stopen, L.E., 1988, Geometry and Deformation History of Mylonitic Rocks and Silicified Zones along the Mesozoic Connecticut Valley Border Fault, Western Massachusetts. M.S. Thesis, Department of Geology and Geography, University of Massachusetts, Amherst, MA, 176 p.

Stoughton, R.M., 1978, History of the Town of Gill, Franklin County, Massachusetts, 1793-1943. Gill, Mass., Published by the Town, 337 p.

Sutter, J.F., 1988, Innovative approaches to the dating of igneous events in the early Mesozoic basins of the eastern United States. In Froelich, A.J. and Robinson, G.R., Jr. (eds.), Studies of the early Mesozoic basins of the eastern United States. U.S. Geological Survey Bulletin, no. 1776, p. 194-200.

Swanson, M.T., 1986, Preexisting fault control for Mesozoic basin formation in eastern North America. Geology, v. 14, p. 419-422.

Van Houten, F.B., 1964, Cyclic lacustrine sedimentation, Upper Triassic Lockatong Formation, central New Jersey and adjacent Pennsylvania. Kansas Geological Survey Bulletin, v. 169, p. 497-531.

Walker, A.D., 1968, *Protosuchus*, *Proterochampsa* and the origin of phytosaurs and crocodiles. Geological Magazine, v. 105, p. 1-14.

Whalley, P.E.S., 1985, The systematics and palaeogeography of the Lower Jurassic insects of Dorset, England. Bulletin of the British Museum Natural History (Geology), v. 39, no. 3, p. 107-189.

Warren, J.C., 1854, Remarks on some fossil impressions in the sandstone rocks of Connecticut River. Boston, Mass., Ticknor and Fields, 54 p.

Willard, M.E., 1951, Bedrock geology of the Mount Toby quadrangle, Massachusetts. U.S. Geological Survey Geological Quadrangle Map GQ-8, 1; 31, 680.

Wing S.L. and Sues, H.-D., 1992, Mesozoic and early Cenozoic terrestrial ecosystems. In Behrensmeyer, A.K., Damuth, J.D., DiMichele, W.A., Potts, R., Sues, H.-D., and Wing, S.L. (eds.) Terrestrial Ecosystems Through Time. ETE, University of Chicago Press, p. 327-416.

Wise, D.U., 1988, Mesozoic stress history of the upper Connecticut Valley at Turners Falls, Massachusetts. In Bothner, W.A. (ed.), Guidebook for Field Trips in Southwestern New Hampshire, Southeastern Vermont, and North-Central Massachusetts. New England Intercollegiate Geological Conference, 74th Annual. Meeting, Keene, NH, Field Trip C-7, p. 351-372.

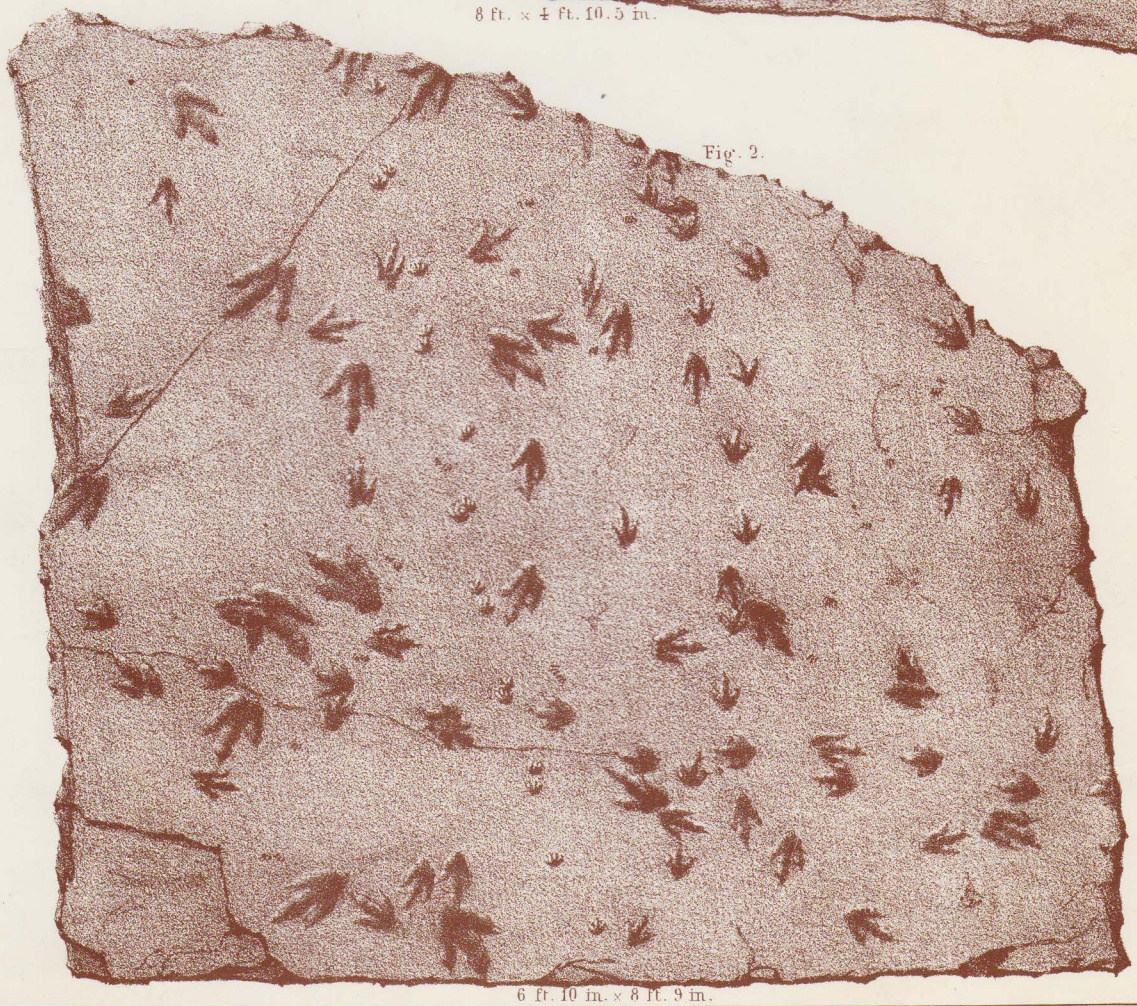
Witte, K.W., Kent, D.V. and Olsen, P.E., 1991, Magnetostratigraphy and paleomagnetic poles from Late Triassic-earliest Jurassic strata of the Newark basin. Geological Society of America, Bulletin, v. 103, p. 1648-1662.

Woodward, A.S., 1895, Catalogue of the fossil fishes in the British Museum (Natural History), part 3. London, England, British Museum (Natural History), 544 p.

Fig. 1.

8 ft. × 4 ft. 10. 5 in.

Fig. 2.



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T. H. Bradford & Co's Lith.