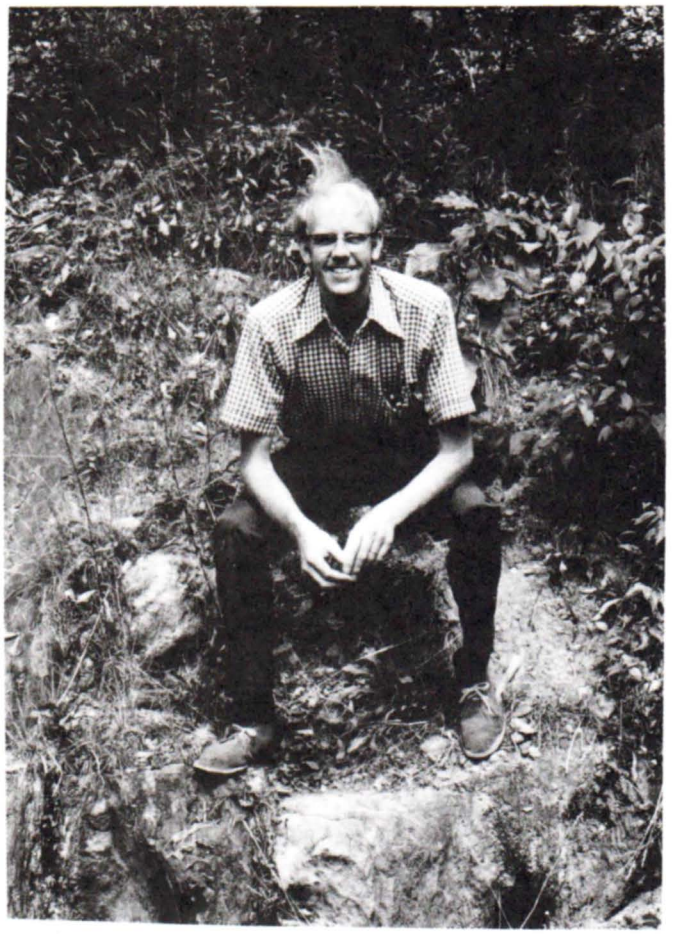
GUIDEBOOK FOR FIELD TRIPS IN THE CONNECTICUT VALLEY REGION OF MASSACHUSETTS AND ADJACENT STATES

Volume 1

84th ANNUAL MEETING
NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE
THE FIVE COLLEGES
AMHERST, MASSACHUSETTS
OCTOBER 9-10-11, 1992

DEDICATION



This past year, NEIGC saw the passing of one of its most stalwart and enthusiastic long-term members, Norman L. Hatch. Norm was a life-long New Englander, with the patience and good ear to hear the message of the rocks, and the ability to persuade and teach all of us how to hear the message more clearly. He was also a common visitor, field trip leader, and friend to the Five College geological community, particularly during his years of mapping in western Massachusetts and when his daughter, Cricket, was a student at Mount Holyoke College. The photograph is a rare one of himself, found in his research collection, and was probably taken by Rolfe Stanley. It shows Norm in 1974 in western Massachusetts, on Route 23 just west of Route 20 in the Woronoco quadrangle. He sits astride a contact zone that occupied a great deal of his attention over many years, the "Richardson Memorial Contact", a sort of personal Holy Grail. Here it is seen in one of the few places in western Massachusetts where it is well exposed and also decorated by a very thin critical quartzite unit. To the left is the Ordovician Cobble Mountain Formation; in the middle, the Silurian Russell Mountain Formation; and on the right the Silurian-Devonian Goshen Formation. Those of us who knew him well will long remember Norm's determination to weigh field relationships fairly and from all points of view. His key contributions were in northern New Hampshire, in western Massachusetts, in the Mount Washington area, and finally in northeastern Vermont, and they dealt particularly with the two great Silurian-Devonian sedimentary belts of New England, the Merrimack synclinorium and the Connecticut Valley synclinorium. Perhaps less well known, was his skill as negotiator between hot-headed compilers and authors during work on the Massachusetts bedrock map. He was dogged in the pursuit of old and new fossil localities, resulting in the identification of key conodonts in the Fitch Formation at Littleton, New Hampshire; the Lower Devonian plant fossils in southern Quebec, and, indirectly, the Devonian conodonts at Bernardston, Massachusetts. Norm's good humor and insightful comments will be sorely missed by many at this meeting. In his memory we dedicate this guidebook.

NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE
84TH ANNUAL MEETING

HONORING
NORMAN L. HATCH
(1932-1991)

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

VOLUME 1

PETER ROBINSON AND JOHN B. BRADY, EDITORS

OCTOBER 9, 10, 11, 1992
AMHERST, MASSACHUSETTS

AMHERST COLLEGE, SMITH COLLEGE,
MOUNT HOLYOKE COLLEGE, HAMPSHIRE COLLEGE,
AND THE UNIVERSITY OF MASSACHUSETTS, AMHERST

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CONTRIBUTION NO. 66,
DEPARTMENT OF GEOLOGY AND GEOGRAPHY, UNIVERSITY OF MASSACHUSETTS'
AMHERST, MASSACHUSETTS 01003

EDITORS' ACKNOWLEDGEMENTS

Without the field trip leaders, with their willingness to lead and their careful and timely preparation of manuscripts, this meeting would not have been possible.

Jack Cheney of Amherst College and George McGill of the University of Massachusetts shared equally with us in organizing this meeting and planned all social events and registration, respectively. Under McGill's guidance, Jennifer Thomson worked hard to make registration a reality. Jack Cheney was assisted by Jackie Newberry. Ed Belt, Director, and Linda Thomas, Curator, made the Pratt Museum available for Friday night Registration and Reception. Walter Coombs kindly agreed to provide tours of the Pratt Museum dinosaur track collection during Registration. Peter Gooding and Amherst College made "cage camping" available at modest cost. The managers and staff of the Newman Center of the University of Massachusetts offered a banquet that students could afford.

Doug Rankin, of the U.S. Geological Survey, located the photograph of Norm Hatch in Norm's personal collection. Carol Vogel at the University of Massachusetts took responsibility for correct positioning of page numbers. Suzanne McEnroe took responsibility for inserting and positioning the line drawings copied by Nancy Vondell from Emerson's *Geology of Old Hampshire County, Massachusetts.* The text was printed at the University of Massachusetts Duplicating Service under the leadership of Leo St. Denis. Cover and binding were produced by Hamilton I. Newell Company in Amherst.

To each of these persons and institutions we offer sincere thanks and acknowledgement.

TABLE OF CONTENTS

GEOLOGY OF THE CONNECTICUT VALLEY REGION: THE LAST 25 YEARS

by
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The New England Intercollegiate Geological Conference was founded in the Connecticut Valley of Massachusetts in 1901 by William Morris Davis, with his trip to visit the Westfield River terraces. Including that time, persons or institutions in this region have hosted the Conference a total of eight times, including 1902 at Mt. Tom, 1904 at Worcester hosted by B. K. Emerson, 1922 at Amherst, 1930 at Amherst, 1941 at Northampton, 1957 at Amherst, and, the last time, 25 years ago in 1967 at Amherst. While it is true that the gap between 1967 and 1992 is the longest between hosted meetings, geologists of the host institutions have been very active in running field trips hosted elsewhere, particularly in 1982, 1985, and 1988.

In 1967 we celebrated the 50th anniversary of B. K. Emerson's Geology of Massachusetts and Rhode Island, which was at that time still the most recent bedrock geologic map of Massachusetts. In 1983 that map was finally superseded by a new map compiled under the leadership of E-an Zen, and already significant changes are being suggested that may lead to a new edition. In 1967 we also celebrated the 30th anniversary of Marland Billings's crucial paper on the Littleton-Moosilauke area of New Hampshire. At that time the consequences of Billings's earlier work were just coming to bear on central Massachusetts geology, and they are continuing to do so.

The 1992 Guidebook is dedicated to our former colleague, Norman L. Hatch, one of the important trip leaders in 1967, and a major contributor to New England geology and particularly to Massachusetts geology in the last 25 years. Norm Hatch played an important though largely hidden role in the 1983 State map compilation, and also undertook to write and/or edit a series of articles explaining what had been put on the map and why. This began with a short general paper in American Journal of Science explaining the rationale for the way the map was organized, and continued with two out of a projected three Professional Papers on the bedrock geology of Massachusetts. It was great good fortune that Norm Hatch made contact with conodont expert Anita Harris, leading to an important paper on the age of the Fitch Formation at Littleton, New Hampshire, and to extensive further explorations for these mysterious but rewarding creatures (see Robinson and Elbert, A-3). One of these efforts produced the following results, which Norm prepared for publication in 1980. This manuscript reflects his sense of humor, often with serious intent, as well as his effort to win R. L. Bates' prize for the highest author/text ratio! I here include in its entirety my best recollection of his draft manuscript , without official U. S. G. S. approval.

ARE THERE CONODONTS CONTAINED WITHIN FINE-GRAINED IMPURE LIMESTONES FROM THE WAITS RIVER AND GILE MOUNTAIN FORMATIONS AT A VARIETY OF LOCALITIES AT LOW METAMORPHIC GRADE IN EASTERN VERMONT AND NORTHERN MASSACHUSETTS?

by
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ABSTRACT

No.

TEXT

No.

REFERENCES CITED

None

In the last 25 years there has been no major change in the grossest features of the geology, the major anticlinoria and synclinoria that dominate region (Figure 1), from west to east: 1) The Adirondack Precambrian massif; 2) The Middlebury synclinorium dominated by Cambrian-Ordovician carbonates and Taconian thrust sheets with clastic rocks of the same ages; 3) the Berkshire - Green Mountain anticlinorium with its Grenville inliers and Lower Paleozoic cover; 4) The Connecticut Valley synclinorium dominated by Silurian - Lower Devonian cover, with its western line of gneiss domes; 5) The Bronson Hill anticlinorium with its plagioclase gneisses, Ordovician, Silurian and Lower Devonian cover, and en echelon gneiss domes; 6) The Merrimack synclinorium with its thick, and now much better known, sequence of Silurian - Lower Devonian strata; and 7) The Boston zone (Milford-

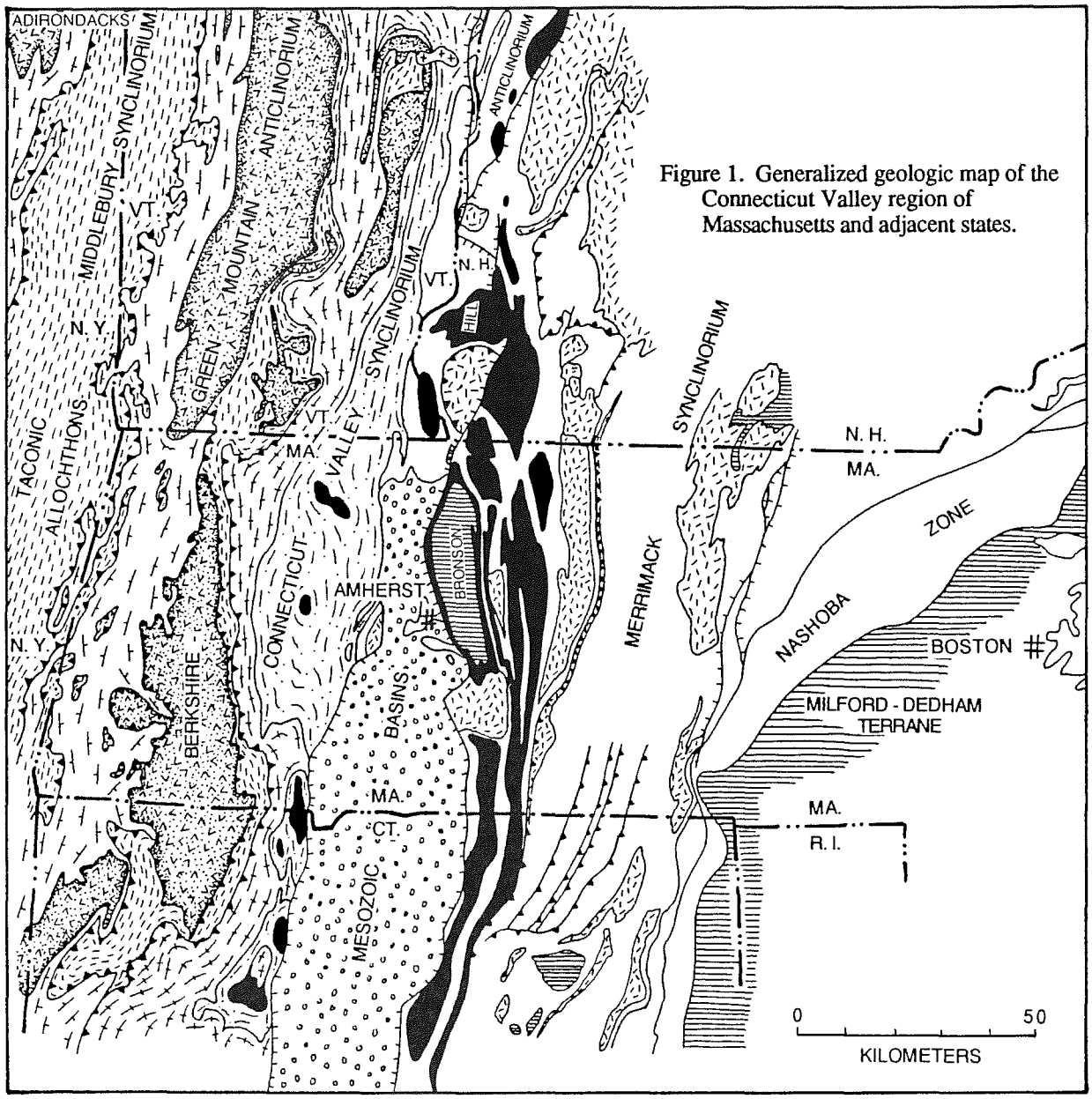

Figure 1. Generalized geologic map of the Connecticut Valley region of Massachusetts and adjacent states.

- Plagioclase gneisses in cores of domes
- Late Proterozoic of Eastern New England
- Cambrian-Ordovician Accretionary Prism
- Cambrian-Ordovician Taconic facies, Hoosac Formation and lateral equivalents
- Cambrian-Ordovician carbonate bank and shelf
- North American Middle Proterozoic basement
- Cretaceous subvolcanic intrusion
- Triassic-Jurassic sedimentary rocks and basalts
- Acadian and younger Paleozoic intrusions
- Silurian-Devonian Vermont sequence
- Ordovician - Silurian - Lower Devonian of Bronson Hill Anticlinorium and Merrimack Synclinorium

Dedham terrane) and the various fault-bounded zones on its west margin. Most paleontological discoveries have only served to reinforce or refine our previous conceptions. There has also been no change in delineation of the major belts of metamorphic rocks in the region, the Vermont metamorphic high to the west, the Connecticut Valley low grade belt, the New Hampshire metamorphic high containing granulite-facies rocks in southern Massachusetts, the Worcester metamorphic low; the eastern Massachusetts metamorphic high; and the weakly metamorphosed area near Boston. However, there has been a profound revolution in how we think about the evolution and timing of these features, brought about by the advent of the theory of plate tectonics, which really had not made a strong effect until just after 1967, and by application of many new analytical techniques. The Connecticut Valley region has now become the sharp focal point for debate about the different plates that shaped the orogenic events of New England, and about when those events took place.

It is now generally accepted that the Taconian orogeny of western New England was produced when a former Cambrian - Early Ordovician passive margin of Laurentia was subducted beneath a magmatic arc, commonly termed the Bronson Hill plate. These events, as summarized by Nick Ratcliffe and Rolfe Stanley, produced a series of major well recognized lithotectonic zones (Figure 1) including the following: 1) The North American Grenvillian basement in an autochthonous position, in various para-autochthonous masses, and in imbricated slices transported for moderate to long distances over their sedimentary cover. A recent development is the discovery by Paul Karabinos and John Aleinikoff of Middle Proterozoic but post-Grenvillian intrusive rocks within some of these basement areas. 2) A Cambrian-Ordovician carbonate bank and shelf with underlying basal clastics seated on the basement, or in para-autochthonous folds and thrusts. 3) A Cambrian-Ordovician continental slope - rise sequence originally deposited outboard of the carbonate bank, but now in large part thrust westward over it to produce the Taconic allochthons, and 4) A Cambrian-Ordovician accretionary wedge complex consisting of Cambrian-Ordovician strata thought to have been deposited on oceanic crust, fragments of that oceanic crust, and strata believed to have been deposited in the trench or forearc region as the arc moved toward its collision with North America. The nature of some of these events will be explored on field trips led by Ratcliffe, Armstrong and Tracy (A-2) in southeastern Vermont; and by Panish (C-2) in western Connecticut.

The exact nature of the land mass that produced the Taconian collision is under constant debate. Philip Osberg named it "craton X". The plagioclase gneisses that physically underlie the Ammonoosuc Volcanics and Partridge Formation, and which had been a candidate for older parts of the arc sequence, have all proved to lie in a narrow age range between 455 and 443. Such an age range is too young for these gneisses to form a basement beneath the Ammonoosuc and Partridge, which themselves have been dated at 453 and 449, respectively. More important, both groups of rocks are too young to have been part of an island arc **before** the Taconian collision. They are contemporary with or post-date the emplacement of the Giddings Brook slice of the Taconic allochthon, generally believed to have occurred about 455 Ma. We thus have a "lost arc" problem of major dimensions. One candidate for the "basement" upon which the magmatic rocks were built is the Late Proterozoic sequence exposed in the core of the Pelham gneiss dome directly east of Amherst (Robinson, Tucker, Gromet, Ashenden, Williams, Reed and Peterson, B-3). In general character, these rocks closely resemble the Late Proterozoic strata exposed in the Avalon zone (Milford-Dedham Terrane) of southeastern New England, leading to speculation that "craton X" may have been an early-arrived fragment from that zone. However, the contact with the overlying magmatic rocks is almost certainly a low angle fault, and one speculation is that the Proterozoic rocks are a piece of Avalon that was underthrust into its present position in the Late Paleozoic.

It is commonly accepted that the Acadian orogeny was produced by a collision between amalgamated craton X and Laurentia with Avalon, but the details are even more murky than the Taconian case. Most of the critical exposures lie in a region strongly affected by Late Paleozoic movements and well east of the region covered in these field trips. The sequence of Acadian tectonic events in much of central New England began with an early phase of west-directed recumbent folding and/or thrust faulting, commonly with a close association with patterns of regional metamorphism, followed by more open folding. In the Connecticut Valley synclinorium this involved westward motion and recumbent folding of the local tectonostratigraphy. In the Bronson Hill anticlinorium it involved both east-over-west recumbent folding of local stratigraphy and overthrusting of stratigraphic units and intrusive igneous rocks from the Merrimack synclinorium to the east.

The tectonic and metamorphic features of the Acadian and younger orogenies are covered in a variety of field trips at this year's meeting. The area of overlap of the Acadian and Taconian orogenies is explored by Ratcliffe, Armstrong and Tracy (A-2), Cheney and Brady (B-4), and Panish (C-2). The Chicken Yard line of the middle Connecticut Valley is a contact between the Littleton Formation of the "New Hampshire sequence" of Silurian-Devonian strata and the Gile Mountain and Waits River Formations of the "Vermont sequence" that follows the

Connecticut Valley chlorite-zone metamorphic low. This contact is examined by Trzcienski, Thompson, Rosenfeld, and Hepburn (A-4) in its type locality and elsewhere along strike, and its nature, whether an unconformity or an early metamorphic thrust is debated. Robinson and Elbert (A-3) examine the Bernardston early fold nappe in its type locality in the Connecticut Valley, where there are Lower Devonian conodonts, and in its root zone within the Bronson Hill anticlinorium. They also examine the relationship of the nappe with the overlying Brennan Hill thrust with its Merrimack-sequence stratigraphy.

In the granulite facies metamorphic region of the Merrimack synclinorium in southern Massachusetts, Berry (B-1) shows evidence for early fold and thrust nappes involving pre-Silurian basement and Lower Silurian Rangeley Formation, now backfolded, overturned, and possibly overthrust to the east. Berry presents a tectonic model for that region. Robinson and Tucker (C-3) make a traverse from the Bronson Hill anticlinorium across the sillimanite-zone region of northern Massachusetts to the Fitchburg plutons. They explore a tectonic sequence involving eastward backfolding of older west-directed fold nappes followed by west-over-east shearing and two later phases of more gentle folding. In the Conant Brook shear zone at the west edge of the Merrimack synclinorium in southern Massachusetts, Peterson (A-1) comes to grips in detail with the overprinting of two high-temperature Acadian shear fabrics in granulite-facies or near granulite-facies rocks. The west-over-east shearing is followed continuously by a phase of strike-parallel right-lateral shear during the Acadian "dome stage", all under dry, high-temperature conditions, demonstrated by quartz *c*-axis fabrics and preserved mineral compositions. In and near the Pelham dome, directly east of Amherst, Robinson, Tucker, Gromet, Ashenden, Williams, Reed, and Peterson (B-3) demonstrate that longitudinal north-over-south ductile shearing is a product of Pennsylvanian deformation, amphibolite-facies regional metamorphism, and minor igneous activity. The pattern of metamorphic ages and of linear structural features, may be explained by overprinting of a Pennsylvanian field of longitudinal ductile strain upon an older field of longitudinal strain produced during the Acadian dome stage. They explore the significance of relict Acadian granulite-facies metamorphism within one part of the Pelham-dome sequence. The nature and timing of folding, faulting, cooling events, and final Late Paleozoic amalgamation on the east side of the Merrimack synclinorium are examined on the basis of the Ar-Ar method by Wintsch (A-5) in eastern Connecticut, and on the basis of overprinted fold and fault fabrics by Goldstein (B-2) in eastern Massachusetts.

Several workers in New England have been in the forefront in the interpretation of the compositions of zoned minerals, particularly garnets showing evidence of growth zoning or resorption zoning, in terms of major fold and thrust events and subsequent thermal relaxation. Spear (C-5) uses this approach to explore the evolution of the western margin of the New Hampshire Acadian metamorphic high, where fold and thrust emplacement of high grade rocks above low grade rocks is characteristic. Ratcliffe, Armstrong and Tracy (A-2) apply it to understand the P-T evolution of a thickening crustal wedge on the west limb of the Connecticut Valley synclinorium. Panish (C-2) presents detailed evidence concerning the high-grade overlap of Acadian and Taconian metamorphic fields in western Connecticut. Cheney and Brady (B-4) present the details of Acadian overprinting on older Taconian metamorphic features in the Hoosac Schist in western Massachusetts. Thomson, Peterson, Berry and Barreiro (C-5) combine a variety of approaches to understand the western margin of the Acadian granulite-facies high in southern Massachusetts; its unusual P-T path with early low-P heating, followed by increasing P and T, and then high-pressure cooling; and its late Acadian age around 360 Ma.

Not surprisingly, the plate-tectonic theory has had a major effect on thinking about Mesozoic stratigraphy, sedimentation, magmatism and tectonics, with the realization that the Connecticut Valley half graben is a failed arm of a major rift system that ultimately produced the modern central Atlantic Ocean. Nevertheless, the most dramatic single development since 1967 was the finding, by palynologist Bruce Cornet, that about half of the then familiar "Triassic" rocks of the Connecticut Valley, including all of the lavas, are in fact Lower Jurassic. Recent research on the sedimentary rocks and on their fossils has concentrated on interpretation of paleoenvironments including alluvial fans, floodplains, playas, and major lacustrine cycles and how they are related to tectonics, climate change, major global cycles, and possible meteorite impacts (Wise, Hubert and Belt, B-5; Olsen, MacDonald, and Huber, C-6). Thermal maturation studies, spurred by the question of petroleum potential, show large variations in paleotemperature estimates thought to be due to complex fluid circulation patterns (Wise, Hubert, and Belt, B-5) Structural studies, aimed at an understanding of stress history, have demonstrated the importance of listric faulting, with some portions of the eastern border fault reaching dips as low as 20°, and have also shown the importance of a short-lived strike-slip regime (Wise, Hubert, and Belt, B-5).

The familiar term "Triassic diabase", as applied to fine-grained dark intrusive igneous rocks, now proves utterly wrong, not only on the basis of palynology, but on the basis of paleomagnetism and radiometric dating (McEnroe and Brown, B-6). This research shows the dikes and sills range in age from lowest Jurassic to Early Cretaceous

(200-120 Ma), and that they have significant variations in geochemistry that appear to relate to changing plate-tectonic environments. There is debate about which intrusive rocks relate to which of the lavas of the Connecticut Valley, but not about the relationship that can be demonstrated between the Higganum-Holden-Fairhaven dikes of Connecticut and Massachusetts and the lowest Jurassic Talcott Lavas of Connecticut (Philpotts and Asher, A-6). Because this magmatic system is well exposed on both sides of the Connecticut Valley border fault, with perhaps as much as 10 km of vertical displacement, it provides an unusual opportunity to study the behavior of the same magmas at depth and at their site of eruption, including the petrologic effects of pressure release and wallrock assimilation. Discussion continues on whether the magmas for dikes and lavas were generated in the mantle directly beneath their site of intrusion or eruption, or whether some of them migrated laterally for long distances, perhaps as much as 250 km from the young central Atlantic spreading center. Trace element data (McEnroe and Brown, B-6) suggest a within-plate character for the earliest Jurassic magmas, a MORB-like character for the later Lower Jurassic and Middle Jurassic magmas, and a return to within-plate character for the Cretaceous magmas, that have commonly been associated with hot-spot magmatism.

In 1967 the Quaternary field trips were conducted by Joe Hartshorn and the late Richard Jahns. Hartshorn was about to join the faculty at the University of Massachusetts following many years of research in Connecticut Valley Pleistocene geology with the U. S. G. S. Jahns was one of the few geologists of this century to master both Quaternary geology and igneous petrology. Jahns' surficial maps of the Mount Toby and Greenfield quadrangles, as well as his free-hand drawings for the 1967 guidebook, remain as graphic masterpieces. This year's field trips by Stone and Ashley (A-7), Larsen (B-7), and Stone, Lapham, and Larsen (C-7), all of whom were Hartshorn's associates, and three of whom are his former students, show that interest and enthusiasm for Quaternary stratigraphy, and glacial and peri-glacial sedimentary processes continues unabated. The interest has been enhanced by recent popular and governmental concern about past climate and the effects of human activities on climate. The research has moved into a more quantitative phase, with some successes in tying in ^{14}C ages on fossil organic material to detailed analyses of the classic varves of glacial Lake Hitchcock. Highlights of some of the most recent findings include the following: 1) The classic and long-debated lower till of New England is now considered to be a product of Illinoisian rather than Wisconsinan glaciation. 2) The Wisconsinan glacial maximum was probably 22,000-21,000 B.P. rather than 18,000-17,000 as believed previously. 3) The ages of the oldest varves in glacial Lake Hitchcock appear to be older than previously thought. 4) The date of draining of the Lake in Connecticut and Massachusetts is better constrained by new ^{14}C dates and stratigraphic relationships. What is now well known is that the Lake had drained before 12,400 B. P., as compared to Flint's estimate of 10,700 B. P., as shown by a ^{14}C age on plant material trapped within fossil permafrost features formed on the lake-bottom surface after its draining. 5) Dramatic new results (Stone and Ashley, A-7; Stone, Lapham, and Larsen, C-7) indicate that a harsh postglacial climate with permafrost existed in New England for perhaps hundreds of years after the draining of Lake Hitchcock.

In recent decades the largest employment of geologists in the region has been in the field of water resources, including their discovery and protection. Effectiveness in this work may require a sophisticated knowledge of bedrock fracture patterns; Quaternary stratigraphy and sediment characteristics in three dimensions; the characteristics of surface water and groundwater flow in various situations; the nature of chemical interactions between waters, rocks, and sediments including crystal-chemical features of mineral surfaces; and the physical and chemical behavior of living and dead organic matter in surface and subsurface environments. Some of these aspects are covered in two applied geology field trips, one on the contamination of groundwater and sediments by volatile organic compounds and remediation techniques (Morgan and Allen, A-8) in eastern Massachusetts, and one on features and problems of several aquifers in the Connecticut Valley and on water-sediment interactions in the Quabbin Reservoir watershed (Motts, Yuretich, Heeley, Exarhoulakos and Skiba, B-8). These excursions show the importance of fundamental training in basic geology, mineralogy and geochemistry to success in applications.

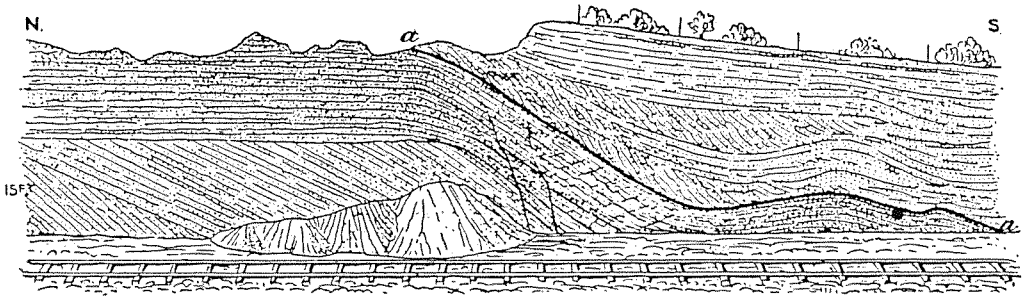

FIG. 42.—Section of north half of a kettle-hole below Dwight's station on the Central Railroad, Belchertown.

Meetings of the New England Intercollegiate Geological Conference

1901 Westfield River Terrace, MA (Davis)
1902 Mount Tom, MA (Emerson)
1903 West Peak, Meriden, CT (Rice)
1904 Worcester, MA (Emerson)
1905 Boston, Nantasket, MA (Johnson, Crosby)
1906 Meriden to East Berlin, CT (Gregory)
1907 Providence, RI (Brown)
1908 Long Island, NY (Barrell)
1909 N. Berkshire Mts, MA (Crosby, Warren)
1910 Hanover, NH (Goldthwaite)
1911 Nahant and Medford, MA (Lane, Johnson)
1912 Higby-Lamentation Blocks (Rice)
1915 Waterbury-Winsted, CT (Barrell)
1916 Blue Hills, MA (Crosby, Warren)
1917 Gay head-Martha's Vineyard, MA (Woodworth, Wigglesworth)
1920 Lamentation and Hanging Hills (Rice, Foye)
1921 Attleboro, MA (Woodworth)
1922 Amherst, MA (Antevs)
1923 Beverly, MA (Lane)
1924 Providence, RI (Brown)
1925 Waterville, ME (Perkins)
1926 New Haven, CT (Longwell)
1927 Worcester, MA (Perry, Little, Gordon)
1928 Cambridge, MA (Billings, Bryan, Mather)
1929 Littleton, NH (Crosby)
1930 Amherst, MA (Loomis, Gordon)
1931 Montreal, PQ (O'Neill, Graham, Clark, Gill, Osborne, McGerrigle)
1932 Providence-Newport, RI (Brown)
1933 Williamstown, MA (Cleland, Perry, Knopf)
1934 Lewiston, ME (Fisher, Perkins)
1935 Boston, MA (Morris, Pearsall, Whitehead)
1936 Littleton, NH (Billings, Hadley, Cleaves, Williams)
1937 NYC-Dutchess Co. (O'Connell, Kay, Fluhr, Hubert, Balk)
1938 Rutland, VT (Bain)
1939 Hartford, CT (Troxell, Flint, Longwell, Peoples, Wheeler)
1940 Hanover, NH (Goldthwaite, Denny, Shaub, Hadley, Bannerman, Stoiber)
1941 Northampton, Ma (Balk, Jahns, Lockman, Shaub, Willard)
1946 Mt. Washington, NH (Billings)
1947 Providence, RI (Quinn)
1948 Burlington, VT (Doll)
1949 Boston, MA (Nicholls, Billings, Shrock, Currier, Stearns)
1950 Bangor, ME (Trefethen, Raisz)
1951 Worcester, MA (Lougee, Little)
1952 Williamstown, MA (Perry, Foote, McFadyen, Ramsdell)
1953 Hartford, CT (Flint, Gates, Peoples, Cushman, Aitken, Rodgers, Troxell)
1954 Hanover, NH (Elston, Washburn, Lyons, McKinstry, Stoiber, McNair, Thompson)
1955 Ticonderoga, NY (Rodgers, Walton, MacClintock, Bartolome)
1956 Portsmouth, NH (Novotny, Billings, Chapman, Bradley, Freedman, Stewart)
1957 Amherst, MA (Bain, Johnson, Rice, Stobbe, Woodland, Brophy, Kierstead, Shaub, Nelson)
1958 Middletown, CT (Rosenfeld, Eaton, Sanders, Porter)
1959 Rutland, VT (Zen, Kay, Welby, Bain, Theokritoff, Osberg, Shumaker, Berry, Thompson)
1960 Rumford, ME (Griscom, Milton, Wolfe, Caldwell, Peacor)
1961 Montpelier, VT (Doll, Cady, White, Chidester, Matthews, Nichols, Baldwin, Stewart, Dennis)
1962 Montreal, PQ (Gill, Clark, Kranck, Stevenson, Stearn, Elson, Eakins, Gold)
1963 Providence, RI (Quinn, Mutch, Shafer, Agron, Chapple, Feininger, Hall)
1964 Chestnut Hill, MA (Skehan)
1965 Brunswick, ME (Hussey)
1966 Katahdin, ME (Caldwell)
1967 Amherst, MA (Robinson, Foose, McGill)
1968 New Haven, CT (Orville)
1969 Albany, NY (Bird)
1970 Rangeley Lakes, ME (Boone)
1971 Concord, NH (Lyons, Stewart)
1972 Burlington, VT (Doolan, Stanley)
1973 Fredericton, NB (Grenier)
1974 Orono, ME (Osberg)
1975 Great Barrington, MA (Ratcliffe)
1976 Boston, MA (Cameron)
1977 Quebec City, PQ (Beland, LaSalle)
1978 Calais, ME (Ludman)
1979 Troy, NY (Friedman)
1980 Presque Isle, ME (Roy, Naylor)
1981 Kingston, RI (Boothroyd, Hermes)
1982 Storrs, CT (Joesten, Quarrier)
1983 Greenville-Millinocket, ME (Caldwell, Hanson)
1984 Danvers, MA (Hanson)
1985 New Haven, CT (Tracy)
1986 Lewiston, ME (Newberg)
1987 Montpelier, VT (Westerman)
1988 Keene, NH (Bothner)
1989 Farmington, ME (Berry)
1990 La Gaspesie, PQ (Trzcienski)
1991 Princeton, ME (Ludman)
1992 Amherst, MA (Robinson, Brady, Cheney, McGill)

THE CONANT BROOK SHEAR ZONE: AN ACADIAN HIGH STRAIN ZONE ALONG THE BRONSON HILL - MERRIMACK BOUNDARY IN SOUTHERN MASSACHUSETTS

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INTRODUCTION

In the New England Appalachians the width of the belt of rocks, primarily deformed and metamorphosed during the Acadian orogeny, is much narrower in southern New England (Connecticut and Massachusetts) than it is further north in Maine. This narrowing of the belt is accompanied by a general increase in metamorphic grade, depth of burial, and intensity of deformation and may be a result of oblique Acadian collision (Rogers, 1981; Chamberlain and Robinson, 1989). Apparent map-scale truncation and attenuation of rock units, particularly in southern New England (Fig. 1) gives the impression that a significant amount of the strain associated with narrowing of the Acadian orogen may have been accommodated along or near the Bronson Hill - Merrimack belt transition. This transition zone, which follows the eastern contact of the Monson Gneiss in southern Massachusetts and northern Connecticut, has been recognized as a zone of high strain: the Bonemill Brook fault in northern Connecticut (Pease, 1982), the Cremation Hill shear zone in southern Connecticut (London, 1988), and the Conant Brook shear zone in southern Massachusetts (Peterson, 1992; Peterson and Robinson, in press) (Fig. 1). Among other things, this boundary has been interpreted to be a terrane boundary (Pease, 1982; Wintsch, 1985), the margin of an accretionary prism (Rogers, 1981), or simply a zone of focused deformation (Robinson and Tucker, 1982). On this trip, we will explore the character of the Conant Brook shear zone in southern Massachusetts and address the significance of this zone in a regional context.

GEOLOGIC SETTING

Paleozoic rocks in southern New England have been divided into a number of tectono-stratigraphic provinces (Hall and Robinson, 1982). Two of these, the Bronson Hill anticlinorium and Merrimack synclinorium (Fig. 1) in central Massachusetts contain structures believed to have formed primarily during the Devonian Acadian orogeny (Robinson, 1979). The Bronson Hill anticlinorium contains a series of structural culminations or domes, with cores of Late Proterozoic and Late Ordovician basement (Thompson et al., 1968; Tucker and Robinson, 1990). The Merrimack synclinorium (or Merrimack belt) to the east is a highly deformed sedimentary belt of Late Ordovician volcanics and Silurian continental slope-rise sediments overridden by Acadian (Early Devonian) flysch (Robinson, 1979). Both trend north-northeast, extending from southern Connecticut into northern Maine (Hatch et al., 1983) (Fig. 1).

The Acadian deformation of these rocks in central Massachusetts and southwest New Hampshire has been divided into three main stages: the nappe stage, backfold stage, and dome stage (Robinson, 1979; Robinson and Hall, 1980; Hall and Robinson, 1982; Robinson et al., 1982). An early stage of west-directed fold nappes is well documented in the distinctive rocks of the Bronson Hill anticlinorium (Thompson et al., 1968) and observed in map pattern and locally in minor folds throughout central Massachusetts (Field, 1975; Tucker, 1977; Robinson, 1979; Robinson et al., 1982; Zen et al., 1983).

Thompson's (1985) interpretation of stratigraphy in the Mount Monadnock area (Fig. 2) allowed him to recognize two major thrust nappe surfaces, the Brennan Hill and Chesham Pond thrusts, that truncate earlier fold nappes. The western of these two, the Brennan Hill thrust, carries rocks of the Monadnock sequence of the Merrimack belt over those of the Bronson Hill sequence (Fig. 2). The Brennan Hill thrust has been traced south through the Royalston (Springston, 1990) and Mount Grace areas (Robinson et al., 1988; Robinson et al., 1991) and is thought to continue south along the eastern margin of the Monson Gneiss (see also Robinson et al., 1991) (Fig. 2). The Chesham Pond thrust juxtaposes rocks of the central New Hampshire sequence and Monadnock sequence within the Merrimack Belt (Fig. 3) (Thompson, 1985). It is thought to project southward from New Hampshire, along or near the eastern contact of the Coys Hill Granite (Fig. 2). Imbricate stacked thrusts observed in the Brimfield-Sturbridge area (Fig. 2), interpreted as part of a large scale duplex structure, may also have formed during this thrust nappe stage (Berry, 1989; this volume).

The complex backfold stage caused large-scale eastward overturning of these earlier fold and thrust nappes (Field, 1975; Robinson, 1979; Thompson, 1985; Berry, 1989) and is responsible for the steep west dips observed in south-

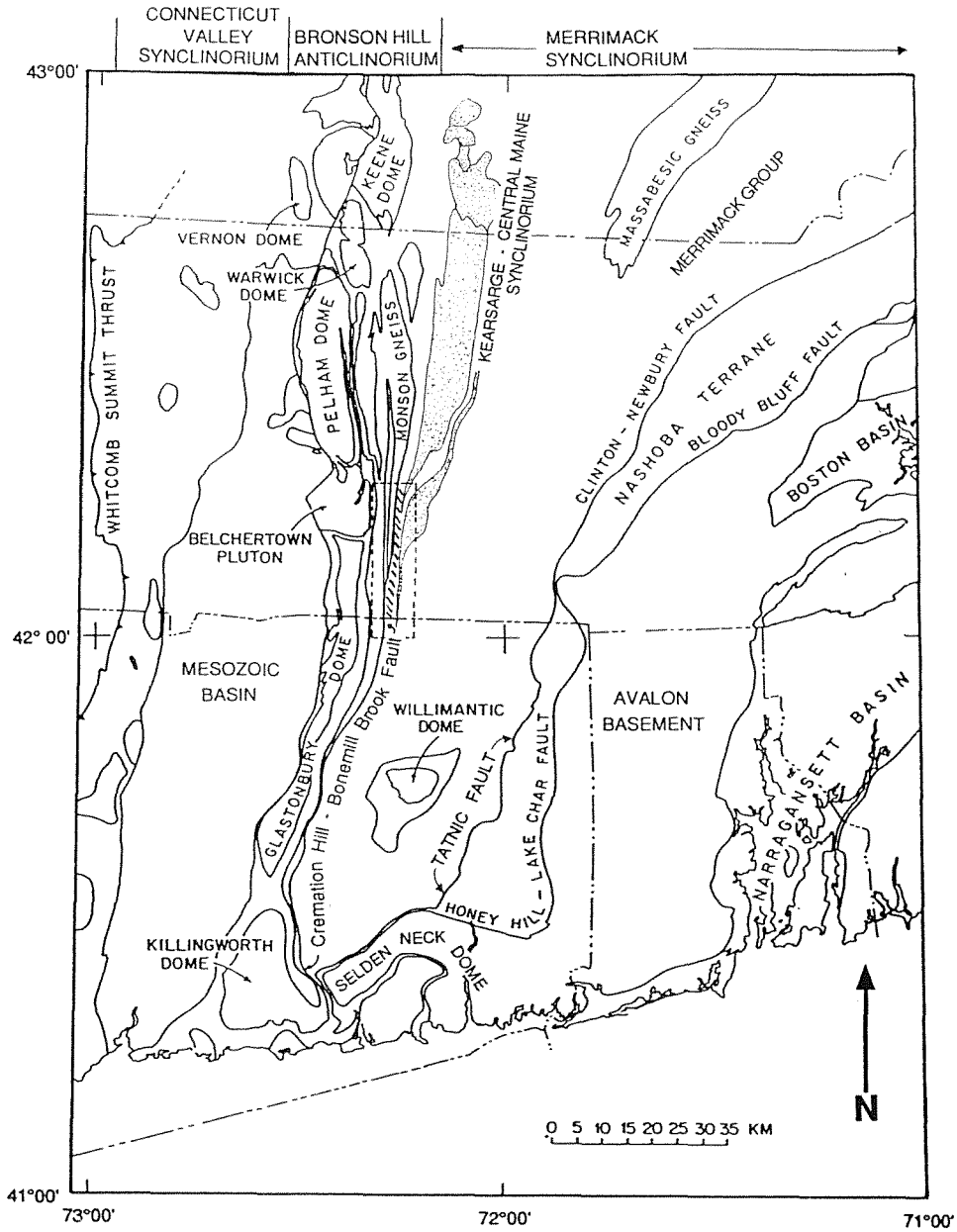

Figure 1. Outline map of southern New England showing major tectonic features (after Robinson and Hall, 1980; Hall and Robinson, 1982). Dashed outline indicates the map area of this study. Shaded areas are approximate outlines of selected Acadian intrusions. Diagonally ruled area is the Conant Brook shear zone.

central Massachusetts. Backfold-stage fabrics include strong east-west-trending mineral lineations (Robinson, 1979; Peterson, 1984) (Fig. 4), asymmetric north-trending, west verging minor folds (Thompson, 1985), minor folds with east-west-trending axes (Robinson, 1967; Peterson, 1984) and local mylonite zones (Robinson et al., 1982; Finkelstein, 1987; Berry, 1989). A west-over-east or reverse-slip sense of motion is consistently indicated by fabrics within steeply west-dipping, backfold-stage mylonites in south-central Massachusetts (Finkelstein, 1987; Berry, 1989; Peterson, 1992). Within central Massachusetts, east-west lineations and other backfold-stage fabrics are best developed to the east in the Merrimack belt (Fig. 4).

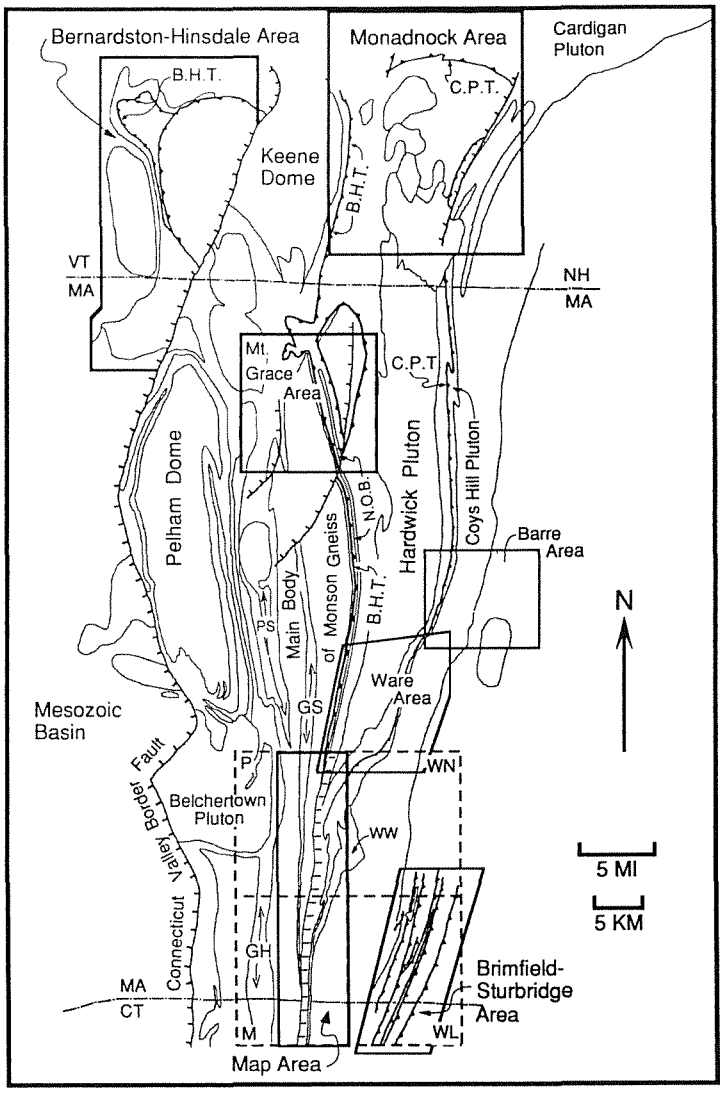

Figure 2. Location map of central Massachusetts and adjacent states, based on Zen et al. (1983), showing major geologic features and areas of previous mapping pertinent to this study including the Monadnock area (Thompson, 1985), the Mount Grace area (Robinson et al., 1988), the Ware area (Field, 1975), the Barre Area (Tucker, 1977), the Brimfield-Sturbridge area (Berry, 1989) and the Bernardston-Hinsdale area (Elbert, 1988). The Royalston area (not shown) (Springston, 1990) links the Monadnock and Mount Grace areas. Quadrangle boundaries shown with dashed outlines are: P = Palmer (Peper, 1976), M = Monson (Peper, 1977), WN = Warren (Pomeroy, 1977), WL = Wales (Seiders, 1976). (See also Peper et al. (1975) for synthesis of these quadrangles.) Labelled fold and thrust nappes are: N.O.B = North Orange band of Monson Gneiss; B.H.T. = Brennan Hill thrust; C.P.T. = Chesham Pond thrust. Labelled synclines are: GH = Great Hill syncline; PS = Prescott syncline; GS = Greenwich syncline. The horizontally ruled area is the Conant Brook shear zone.

Dome-stage deformation produced the most pervasive linear fabrics observed in rocks across west-central Massachusetts (Fig. 4). These consist of mineral lineations with a north-south trend, locally parallel or sub-parallel to intersection lineations. Other common dome-stage fabrics include small- to large-scale, open to tight, north-south-trending, upright folds and locally well developed crenulation cleavage (Robinson, 1963; Ashenden, 1973; Onasch, 1973; Michener, 1983; Tucker, 1977; Peterson, 1984; Thompson, 1985; Berry, 1989). Penetrative shear fabrics with a north-south transport direction and top-side-south shear sense (see Robinson et al., this volume), observed in the Pelham dome area of the Bronson Hill anticlinorium, have also traditionally been attributed to the dome stage (Onasch, 1973; Reed, 1990).

Metamorphic conditions in central Massachusetts range from lower amphibolite to granulite facies. Rocks within the Conant Brook shear zone reached granulite or near-granulite-facies conditions during peak metamorphism. The high grade metamorphism and ductile deformation observed in central Massachusetts is thought to be primarily a product of Acadian orogenesis (eg. Robinson et al., 1986). Acadian metamorphic ages obtained across central Massachusetts include U-Pb monazite ages of 367±3 Ma. for schists from the Pelham dome (Robinson and Tucker, 1991), 350±3 Ma. for schists from the Greenwich syncline (Tucker, personal communication), and ≈ 363 Ma. for schists from the Brimfield-Sturbridge area (Berry, 1989 - preliminary age from Barriero; see also this volume). Recent U-PB monazite and sphene metamorphic ages and zircon ages from pegmatites in the Pelham dome region (Fig. 2) suggest that there is also localized Late Paleozoic metamorphism and deformation in that area (Gromet and

	Bronson Hill Belt	Merrimack Belt	
		Monadnock Sequence	Central New Hampshire Sequence
Lower Devonian	Littleton	Littleton	Littleton
Silurian	Fitch	Warner	Madrid
		Francestown	Smalls Falls
	Clough	Perry Mountain	
		Rangeley	
Late Ordovician	Partridge		
	Ammonoosuc		
	Monson ("Oliverian gneisses")		
Late Proterozoic	Poplar Mountain		
	Dry Hill		

Figure 3. Stratigraphic column for the Bronson Hill belt (Bronson Hill anticlinorium) and the Merrimack belt (Merrimack synclinorium) in central Massachusetts and New Hampshire (from Robinson et al., 1991; Thompson, 1985). The Merrimack belt is divided into a more proximal sequence described in the Monadnock area (Thompson, 1985) and a more distal sequence characteristic of central New Hampshire (Hatch et al., 1983; Duke, 1984).

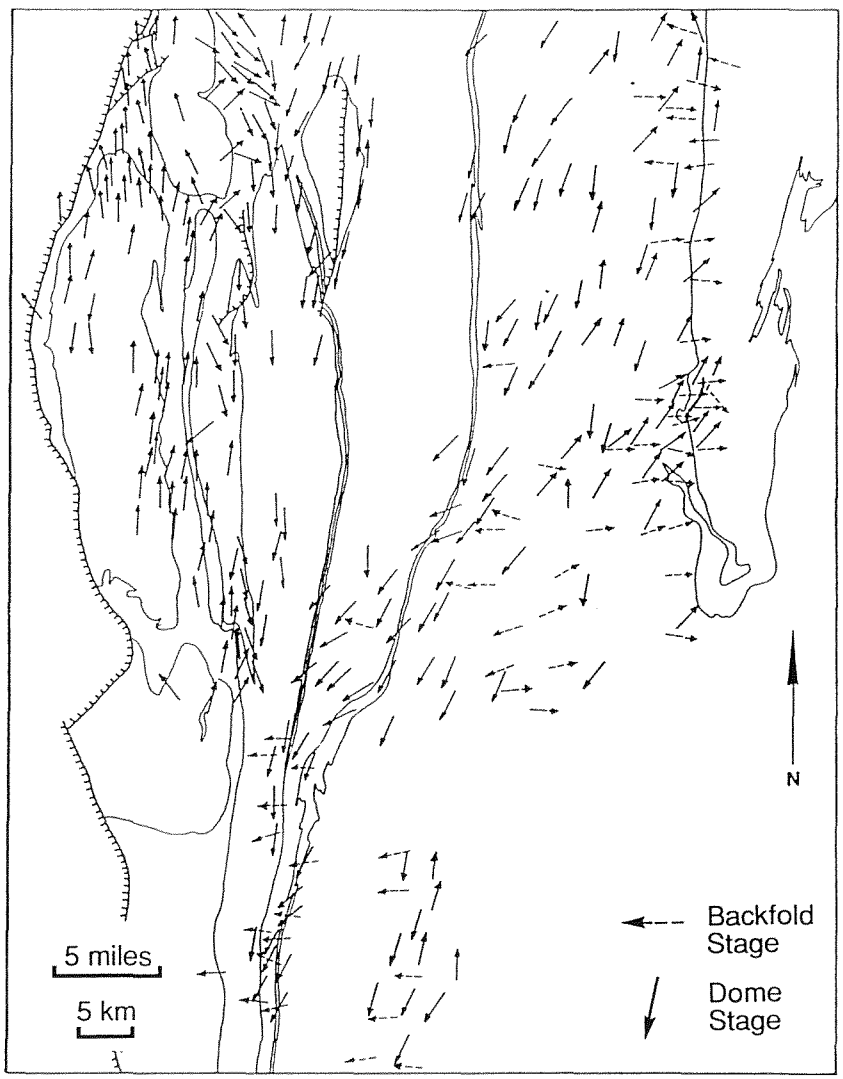

Figure 4. Regional map showing distribution and representative orientations of backfold- and dome-stage lineations across central Massachusetts (from Peterson, 1992).

Robinson, 1990; Tucker and Robinson, 1991; Robinson and Tucker, 1991; Robinson et al., this volume). The dome-stage shearing observed in the Pelham dome may actually be synchronous with this metamorphism. The current distribution of metamorphic ages suggests that the approximate eastern extent of high grade Late Paleozoic metamorphism in the Bronson Hill anticlinorium is near the western boundary of the Monson Gneiss (Fig. 2). This suggests that peak metamorphism in the vicinity of the Conant Brook shear zone is strictly Acadian.

On this trip we will examine the Conant Brook shear zone, a kilometer-wide zone of high strain that follows the eastern flank of the Bronson Hill anticlinorium in south-central Massachusetts (Fig. 2). It is bounded to the west by the main body of Monson Gneiss, with most of the focused deformation observed in pelitic schists of the Rangeley Formation within the Merrimack belt (Fig. 3). It extends ≈25 kilometers along strike from the southern Ware area to the Massachusetts/Connecticut state line (Fig. 2). The shear zone is characterized by well developed shear fabrics that include a dominant, west-dipping mylonitic foliation, two mineral lineations with distinctly different orientations, and numerous asymmetric fabrics that indicate a strong component of simple shear and give a sense of relative motion during deformation. The two lineations in this zone have a similar orientation to the two regionally pervasive lineations shown in Fig. 4, making this one of the few areas in central Massachusetts where both are equally well developed. The eastern margin of the shear zone is marked by a decrease in overall shear strain and an increase in the scale of partitioning between high and low strain domains. It is coincident with the southward and westward attenuation or truncation of Acadian intrusives (Fig. 2) and the western regional granulite-facies boundary (see Thomson et al., this volume). To the south, the shear zone is apparently equivalent to the Bonemill Brook fault and Cremation Hill shear zone in northern and southern Connecticut, respectively (Fig. 1), whereas to the north, the intensity of shear strain appears to decrease.

The Conant Brook shear zone appears to have been an important zone of focused deformation throughout the Acadian orogeny. We will examine evidence for two phases of shearing deformation within the zone that appear to correlate with the regional backfold- and dome-stages of deformation. In addition we will see some possible evidence for earlier (nappe stage?) deformation in this zone. Detailed fabric observations have provided constraints on deformation kinematics and on the relationship between the different phases of deformation as well as the relationship between shearing deformation and metamorphism. On the trip we will examine some of these relationships and look at the different styles of deformation evident in the vicinity of the shear zone. In addition we will look at the rock types present within and adjacent to the Conant Brook shear zone and some of the criteria for stratigraphic assignments.

ROCK UNITS

Rocks west of the Conant Brook shear zone are typical of those described in the Bronson Hill belt (Figs. 3 and 5) (Billings, 1937; 1956; Robinson, 1963). Rocks within and east of the Conant Brook shear zone are correlated with the Monadnock sequence in the Merrimack belt (Figs. 3 and 5). The Silurian rocks in the Merrimack belt represent a thick clastic sequence deposited in a trough to the east of the Bronson Hill shelf sequence. There are no fossils preserved or dateable cover units within the Merrimack belt in Massachusetts. Thus, the stratigraphic section defined for this region and used here is based on correlations with lower grade rocks along strike in northwestern and central Maine, where there is some fossil control (Moench and Boudette, 1970; Robinson et al., 1982; Hatch et al., 1983; Thompson, 1985). An alternative stratigraphic sequence for cover rocks in the Merrimack belt has been suggested for northern Connecticut and southern Massachusetts (eg. Peper et al., 1975; Pease, 1975; Seiders, 1976; Peper, 1976; 1977; Pomeroy, 1977). In this interpretation, much of the cover in this region is part of a thick, westward-younging homoclinal stratified sequence within the Ordovician(?) to Devonian (?) Brimfield Group (Peper et al., 1975), which includes the Bigelow Brook, Hamilton Reservoir, and Mount Pisgah Formations. A number of Acadian intrusive bodies are present just to the east of the Conant Brook shear zone. These include the Hardwick tonalite, the Gneiss of Ragged Hill, the Coys Hill Granite, and the diorite of West Warren. Attenuation and truncation of these bodies marks the approximate eastern margin of the zone (Figs. 2 and 5). More detailed descriptions of these units can be found in Peterson (1992).

Bronson Hill Belt

The western part of the map area (Fig. 5) is underlain primarily by the Monson Gneiss and rocks of the Bronson Hill sequence. A relatively complete Bronson Hill sequence is present along the western margin of the main body of Monson Gneiss, but only limited exposures of Late Ordovician Ammonoosuc Volcanics and Partridge Formation are found near the eastern margin of the gneiss. Of these units, we will only see the Monson Gneiss, one of a series of Late Ordovician gneiss exposures that define the trend of the Bronson Hill anticlinorium and has been included in the Oliverian Magma series of Billings (1937; 1956). It is predominantly a felsic biotite- or hornblende-bearing

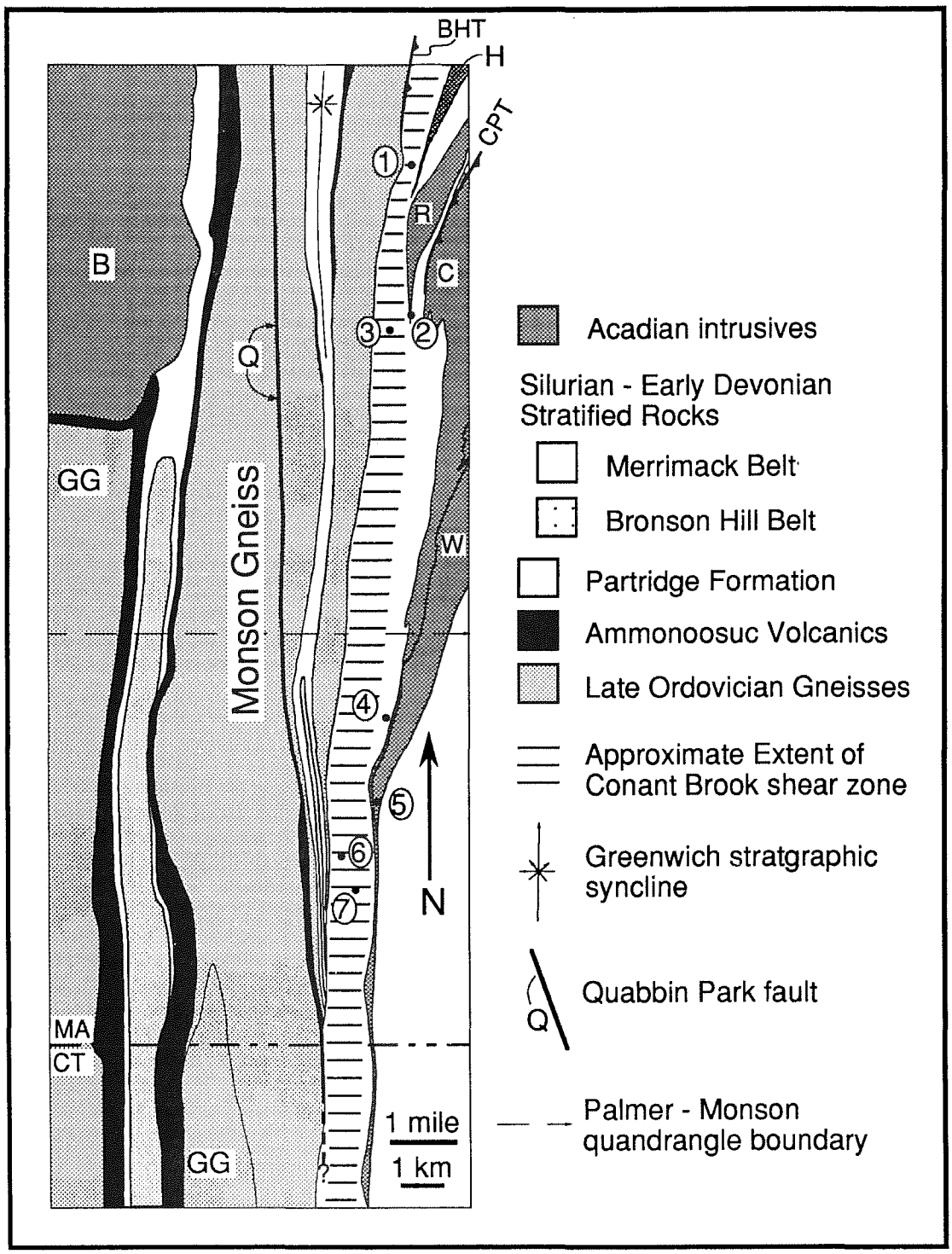

Figure 5. Bedrock geologic map of the Palmer and Monson quadrangles (see Fig. 2 for location) based on mapping of this study and compilation from Peper (1976; 1977). Stratified Silurian-Devonian rocks of the Bronson Hill belt on this map include the Silurian Clough Quartzite and the Devonian Littleton Formation in the Great Hill syncline. Those of the Merrimack belt ar primarily Silurian Rangeley Formation. Labelled units include: GG = Glastonbury Gneiss; B = Belchertown pluton; H = Hardwick pluton; R = Gneiss of Ragged Hill; C = Coys Hill Granite; W = Diorite at West Warren. Labelled structural features include: BHT = Brennan Hill thrust and CPT = Chesham Pond thrust. Numbered dots indicate stops on this trip.

tonalitic or granodioritic gneiss, locally interlayered with fine- to coarse-grained amphibolite. U-Pb ages of zircons from the Monson Gneiss lie between 454 +3/-2Ma. and 442 ±2 Ma. (Tucker and Robinson, 1990).

Merrimack Belt

Of the units in the Monadnock sequence of the Merrimack belt (Fig. 3), only the Rangeley Formation and possibly the Francestown and Warner Formations have been identified in the vicinity of the Conant Brook shear zone. The most common rock type within the Rangeley Formation is medium- to fine-grained, feldspathic to aluminous, gray- or rusty-weathering schist. The typical mineral assemblage within schists in this area is quartz-plagioclase-biotite-orthoclase-garnet- sillimanite-graphite-ilmenite-apatite-zircon and/or monazite±pyrrhotite. Cordierite is common in the granulite-facies rocks east of the shear zone. In southern New Hampshire and northern Massachusetts, Thompson (1985) and Springston (1990) identified a subtle internal stratigraphy within the Rangeley Formation, based largely on weathering character (ie. gray- or rusty-weathering) and the local presence or absence of conglomerates or calc-silicates. This internal stratigraphy was not identified within the rocks of this study, however the Rangeley was broadly divisible into gray- and rusty-weathering schist. These divisions were mappable in areas of good outcrop control (eg. Fig. 6), but difficult to correlate throughout the area. Other rock types characteristic of the Rangeley Formation (Thompson, 1985) and found in the area include quartz-feldspar-biotite-garnet granulites, calc-silicates, feldspathic quartzites, and sandy, gritty, or conglomeratic lenses or layers. Calc-silicates and granulites are common throughout the shear zone, but sandy quartzite or conglomerate localities are rare. Also present are abundant pegmatites, granitic veins, and melt patches. Within the shear zone, these are generally disaggregated and highly deformed (see below).

Limited exposures of the Francestown and possible Warner Formations are present in the Monson quadrangle (Stop 4). These discontinuous exposures lie along strike from similar belts of rock mapped to the north in the Ware area (Fig. 2) (Field, 1975). The Francestown in this area is a distinctive, very rusty-weathering graphite-pyrrhotite calc-silicate granulite (Field, 1975; Peterson, 1992). It is mica-poor and locally very quartz rich. Rare fresh surfaces have a white to blue-gray color. Outcrops are typically slabby or blocky. A belt of distinctive big garnet, sillimanite-rich gneiss/schist, east of the Francestown Formation is correlated with similar rocks along strike in the Big Garnet syncline (Field, 1975). Recently, Robinson and Goldsmith (1991) have suggested that this belt of rock may be correlative with the uppermost part of the Silurian Warner Formation, rather than the Littleton Formation.

Intrusive rocks

Three of the four Acadian intrusive bodies present in the eastern part of the Palmer and Monson quadrangles, the Hardwick tonalite, Gneiss of Ragged Hill, and Coys Hill granite, appear to be truncated or attenuated along the eastern margin of the Conant Brook shear zone. All four of these intrusive bodies were significantly deformed and metamorphosed during the Acadian Orogeny.

The Hardwick tonalite is dominantly a biotite- or hornblende-bearing tonalite (Shearer, 1985). It is identical to the Spaulding Quartz Diorite in New Hampshire (Fowler-Billings, 1949; Shearer, 1985; Thompson, 1985) which has a Devonian Rb-Sr whole rock age of 393±5 Ma. (Lyons and Livingston, 1977; Lyons et al., 1982). The gneiss of Ragged Hill (see Stop 2), named by Field (1975) for extensive exposures on Ragged Hill in the Ware area (Fig. 2), is typically a light-colored,weakly to strongly foliated plagioclase-quartz- K-feldspar-biotite gneiss with minor garnet, sillimanite, zircon or monazite, and oxides. The Coys Hill granite is a coarse-grained, weakly to strongly foliated granite gneiss with large tabular microcline phenocrysts, up to several centimeters long, in a plagioclase-quartz-biotite-garnet matrix. The Coys Hill Granite is lithologically similar to and has been correlated with the Kinsman Granite that makes up the Cardigan Pluton in New Hampshire (Field, 1975; Tucker, 1977; Thompson, 1985; Robinson et al., 1986). The Kinsman Granite is interpreted to be an early Acadian pluton (Clark, 1972; Barriero and Aleinikoff, 1985), intruded during the latest Silurian or earliest Devonian. The age of the granite is based on a 413 ±5 Ma. Sm-Nd whole rock and garnet age (Barriero and Aleinikoff, 1985) and a 402 ±19 Ma. Rb-Sr whole rock age (Lyons et al., 1982). Diorites and tonalites of the West Warren pluton, exposed to the east of the Coys Hill granite are typically dark-colored, strongly foliated, medium- to fine-grained gneisses. Intrusive relations between the two units in the West Warren quadrangle (Pomeroy, 1974) suggest an overlap in the age of different phases of emplacement of these two plutons. This suggests that it may be an early Acadian (latest Silurian to earliest Devonian) intrusive, similar in age to the Coys Hill Granite.

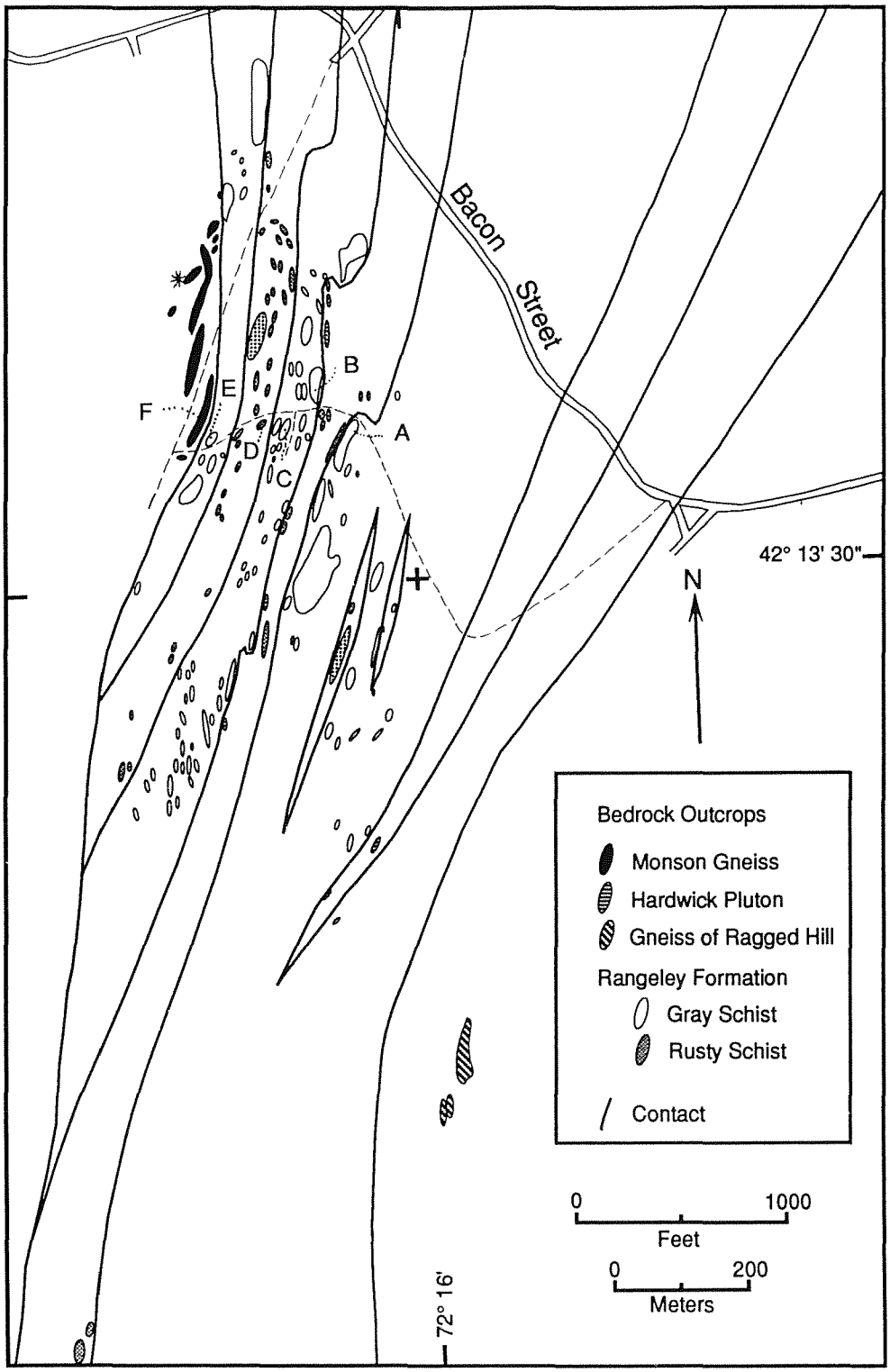

Figure 6. Detailed map of the Pattaquattic trail area (Stop 1) in northeast part of the Palmer quadrangle south of Bacon Street showing location of bedrock outcrop control. Different patterns are used for different rock types. Mapping in this area was done on a scale of 1:6250. The outcrop noted with an asterisk contains the fold for which data is plotted in Fig. 11. Dashed line is part of the Pattaquattic trail, followed for the traverse of Stop 1. Letters designate outcrops visited at Stop 1.

STRUCTURAL FEATURES ASSOCIATED WITH THE CONANT BROOK SHEAR ZONE

Most of the structural features examined in the area of Fig. 5 and associated with the Conant Brook shear zone were produced during the late Acadian backfold and dome stages. A few major nappe-stage features are interpreted to project south into the vicinity of the Conant Brook shear zone. These include the Brennan Hill and Chesham Pond thrusts and the North Orange nappe (Fig.2). Small-scale structural fabrics could not be tied to these structures, however map-scale features are indicative of their presence. We will see possible evidence of early (nappe-stage?) deformation along strike from the Chesham Pond thrust at Stop 4. Backfold- and dome-stage fabrics and structural features within the Conant Brook shear zone are present at all scales of observation.

The Conant Brook shear zone

Map-scale Structural Features. At map-scale, the Conant Brook shear zone can be identified by the attenuation or truncation of map units and earlier structural features (Figs. 2 and 5). One of the most striking features on maps of this area is the dramatic southward narrowing, attenuation, and truncation of stratified and intrusive rock bodies into the Conant Brook shear zone. This narrowing has a distinct asymmetry that may be related to deformation within the shear zone. As noted above, the eastern margin of the shear zone is partly defined as the locus of attenuation or truncation of intrusive rock bodies. Each of these bodies is characterized by dramatic southwestward attenuation and by the presence of high strain zones near or at their western margins, adjacent to the shear zone.

Along its west margin, the Conant Brook shear zone may truncate nappe stage structures, including the North Orange nappe and the Brennan Hill thrust, and dome-stage structures such as the Greenwich stratigraphic syncline (Fig.5). In addition, the thin belts of gray- and rusty-weathering schists, mapped locally within the Rangeley Formation in the shear zone, appears to have an orientation that is rotated clockwise from the more northerly strike of the shear zone. In the area of Stop 1 in the northern Palmer quadrangle (Fig. 6) at least one of these belts appears to be truncated along the eastern margin of the Monson Gneiss.

Planar Structural Features. Planar structural features observed within the map area include bedding, the dominant regional foliation, and the composite mylonitic foliation associated with formation of the shear zone. Exposures of well preserved bedding were not observed within the shear zone, however, compositional layering between mica-rich layers and more quartz-feldspar rich layers is common. In two localities, just east of the shear zone, possible graded bedding suggests that locally the rocks are right-side-up.

The regional foliation is defined primarily by the planar alignment of platy minerals and of linear fabric elements and generally has a steep to moderate west dip and north-northeast strike. The dip is quite steep (70 - 80° west) in the vicinity of the Monson Gneiss and typically becomes shallower eastward (Field, 1975; Berry,1989; Peper, 1976; 1977; Pomeroy, 1977; Seiders, 1976; Peper et al., 1975), where it averages 40° west in the eastern part of the Brimfield-Sturbridge area (Fig. 3) (Berry, 1989). Formation of this foliation has been attributed to nappe-stage deformation (eg. Tucker, 1977), however it is most likely a composite fabric.

Outside and east of the Conant Brook shear zone, smaller-scale discrete shear zones deform the regional foliation. The foliation external to these shear zones has a steeper west dip than the foliation defining the shear zone (see Stop 5). The character of the earlier foliation is not distinct from that of the later cross-cutting mylonitic foliation. Similar discrete shear zones with shallower dips than the dominant foliation have also been described by Berry (1989) in the Brimfield-Sturbridge area. These discrete shear zones are similar in style and kinematics to the backfold-stage shearing observed within the Conant Brook shear zone (see below). The earlier fabric, cut by these discrete shear zones, was not observed or could not be distinguished from the dominant mylonitic foliation within the Conant Brook shear zone. The geometry of these small-scale shear zones may reflect the overall geometry of the Conant Brook shear zone in this area.

Within the Conant Brook shear zone there is a dominant mylonitic foliation that dips steeply west. The attitude of foliation is relatively consistent throughout the shear zone with an average dip between 60° - 70°W (Fig. 7a). Deviations from this orientation are generally produced by minor folds in the foliation. From the northern to southern part of the area the strike of foliation changes slightly from NNE-SSW to N-S. This mylonitic foliation is defined by planar alignment of biotite grains and by the plane of alignment of linear fabric elements. At the outcrop and thin section scale, foliation is slightly anastomosing both along strike and down the dip. This curving nature of

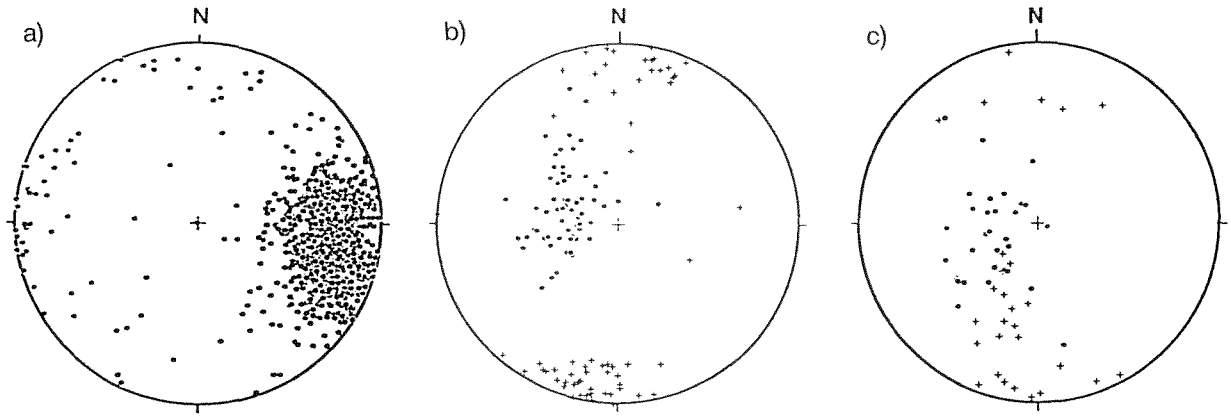

Figure 7. Equal area nets showing: (a) 819 poles to mylonitic foliation from shear zone rocks east of the Monson Gneiss, (b) and (c) the orientation of both L_W and L_S lineations from the Palmer and Monson quadrangles, respectively. Lineation data shown is only from outcrops where both lineations were observed, in most cases on the same foliation surface. The plots show a bimodal population for lineations measured throughout the main shear zone. The two plots differ mainly in the dominant orientation of L_S. Closed circles (•): L_W lineations; crosses (+): L_S lineations.

the mylonitic foliation is interpreted to be a product of partitioning of deformation into high and low strain domains (cf. Bell, 1981).

Linear Fabrics. Two distinct populations of mineral lineations, present within the Conant Brook shear zone, can be distinguished largely on the basis of their orientation. These have been correlated with the two regionally pervasive lineations (Fig. 4) (Peterson, 1992; Peterson and Robinson, in press). Within the shear zone, the backfold-stage lineations (L_W) have a steep west plunge and the dome-stage lineations (L_S) have a shallow north or south to southwest plunge. Each lineation appears to be genetically related to a set of mylonitic microstructures. Within the Conant Brook shear zone, both lineations lie within the plane of dominant mylonitic foliation, although in detail, individual folia may contain only one of the two lineations (see below).

The orientation relationship between the two lineations within the mylonitic foliation changes from the northern to southern parts of the shear zone. With few exceptions, L_W maintains a consistent, steep west plunge throughout the zone (Fig. 7b,c), whereas the orientation of L_S changes from the northern to the southern parts of the zone. In the north, L_S is generally orthogonal to L_W and has a shallow north or south plunge (Fig. 7b). The within-plane angular difference between the mean orientation of L_W and L_S in the north is approximately 82°. The orientation of L_S changes in the south to a more south-southwest trend and slightly steeper plunge (Fig. 7c). In the south L_S has two point maxima, so that the within plane angular differences with Lw are 38° and 70°. Despite this smaller angle between the two lineation directions in the south, the lineations still appear geometrically distinct.

Both lineations are defined by strong alignment of coarse prismatic sillimanite and the dimensional preferred orientation of dynamically recrystallized quartz and feldspar grains, in some cases with aspect ratios greater than 10:1. In foliation-parallel sections, coarse prismatic siliimanite and elongate quartz and feldspar crystals are parallel to each other and to the local elongation direction of porphyroclast tails. Not all of the prismatic sillimanite in the area is strongly oriented. Most rocks possessing a strong sillimanite lineation also have local zones, on the scale of millimeters to centimeters, of randomly oriented sillimanite. These apparently represent low strain domains surrounded by anastomosing higher strain domains of strongly oriented sillimanite (see Stop 6).

Both lineations are observed throughout the main shear zone, however their distribution is heterogeneous at all scales of observation. Locally, one lineation is particularly well developed to the exclusion of the other. However, in a large number of outcrops, the two lineations can be observed together, in many cases on the same foliation

surface. Outcrops like that at Stop 3, where the scale of partitioning between the two lineations was evident in the field, were particularly useful in comparing the lineations and related fabrics. Based on observations at these outcrops, the physical characteristics of the two lineations are indistinguishable and distinct differences between mylonitic fabrics associated with each are not evident.

Kinematic Indicators. Asymmetric recrystallized porphyroclast tails and strain shadows are the most prevalent kinematic indicators produced by shearing during both phases of lineation development, however, asymmetric folds and shear bands are also common (Hanmer and Passchier, 1991). These kinematic indicators are evident both in outcrop and thin section. In mylonites associated with development of the backfold-stage lineation, all kinematic indicators suggest a west-side-up (reverse-slip) sense of motion. All kinematic indicators associated with the dome-stage lineation have a west-side-north or dextral sense of motion. These kinematics indicators are pervasive and consistent throughout the Conant Brook shear zone.

Deformational Microstructures and Processes. Within the Conant Brook shear zone, most minerals in the schists have undergone intense grain-size reduction and dynamic recrystallization characteristic of mylonite formation (Tullis et al., 1982). Deformed phases within pelitic schists include quartz, orthoclase, plagioclase, biotite, and sillimanite. The schist matrix is comprised of fine- to medium-grained dynamically recrystallized quartz, feldspar, and biotite. Quartz ribbons are common. All of these phases retain evidence of dynamic processes with little evidence of annealing. Based on quartz and feldspar textures that indicate a predominance of recrystallization-accomodated dislocation creep, Peterson (1992) has suggested that deformation took place at high temperatures and generally high strain rates. Garnet, feldspar, or polymineralic aggregates of quartz, feldspar, mica, or other phases, make up most of the porphyroclasts in these rocks. The polymineralic porphyroclasts appear to have formed by extreme disaggregation and attenuation of pegmatite or granitoid bodies, suggesting that these rocks have undergone a relatively high degree of strain.

Crystallographic Preferred Orientation. Fabrics produced during both phases of lineation development show remarkable similarity at the macro and micro scale, but have discernable differences at the crystal lattice-scale. In order to characterize deformation at the crystal-lattice scale, the orientations of quartz c-axes were measured in quartz ribbons in samples collected from across the shear zone using a universal stage (Peterson and Robinson, in press). Quartz crystallographic fabric orientations produced by deformation reflect a combination of the initial orientation distribution, the deformation history and conditions of metamorphism during deformation, the active deformation mechanisms or slip systems, the shape of the finite strain ellipsoid (e.g., plane strain, etc.), and the post-deformation recrystallization history (Lister and Williams, 1979; Lister and Hobbs, 1980; Schmid and Casey, 1986). Quartz crystallographic fabrics were used here for preliminary determination of the active slip systems in quartz and for comparison with external mylonitic fabrics associated with L_W and L_S and not as kinematic indicators. In each of the samples measured, a single lineation, either L_W or L_S, was obvious in outcrop and thin section. Distinct differences were observed in the relationship between the c-axis orientations and the lineation orientation for L_W and L_S. These observations are discussed in detail in Peterson and Robinson (in press) and summarized briefly here. In sections where lineation L_W is exclusively developed, c-axes show an orientation maximum in the foliation plane, perpendicular to the lineation. There is also some spread of orientations in a girdle at a high angle, but not perpendicular to the foliation. This asymmetric spread of data may reflect a strong component of simple shear deformation (Simpson and Schmid, 1983; Lister and Hobbs, 1980). In contrast, samples dominated by L_S have quartz c-axis orientatations concentrated near or parallel to the lineation direction. The two sets of data have similar orientation maxima with respect to geographic coordinates, which is generally north-south, with a shallow plunge, but show differences in the geographic orientation of the data spread.

Implications of Fabric Observations. Many aspects of the mylonitic fabrics related to the two lineations within this shear zone are identical. Both lineations are mineral stretching lineations defined by the alignment of sillimanite needles, elongate quartz grains, and the stretching direction of porphyroclast tails. At the outcrop and thin-section scale, the two are differentiated mainly on the basis of orientation, in that L_W has a steep west plunge, whereas L_S has a shallow north-south or southwest plunge. The sense of shear associated with development of each lineation is consistent throughout the shear zone. L_W is associated with a west-side-up, reverse sense of motion. Fabrics associated with L_S indicate dextral, west-side-north motion.

The relationship between the two lineations and related mylonitic folia and the nature of partitioning between the two sets of mylonitic fabrics is complex. In detail, the orientation of the dominant foliation forms an enveloping surface to a composite foliation consisting of two sets of anastomosing folia (S_W and S_S in Fig. 8), related to the two lineations Lw and Ls, respectively. The angle between these two sets of folia is typically very

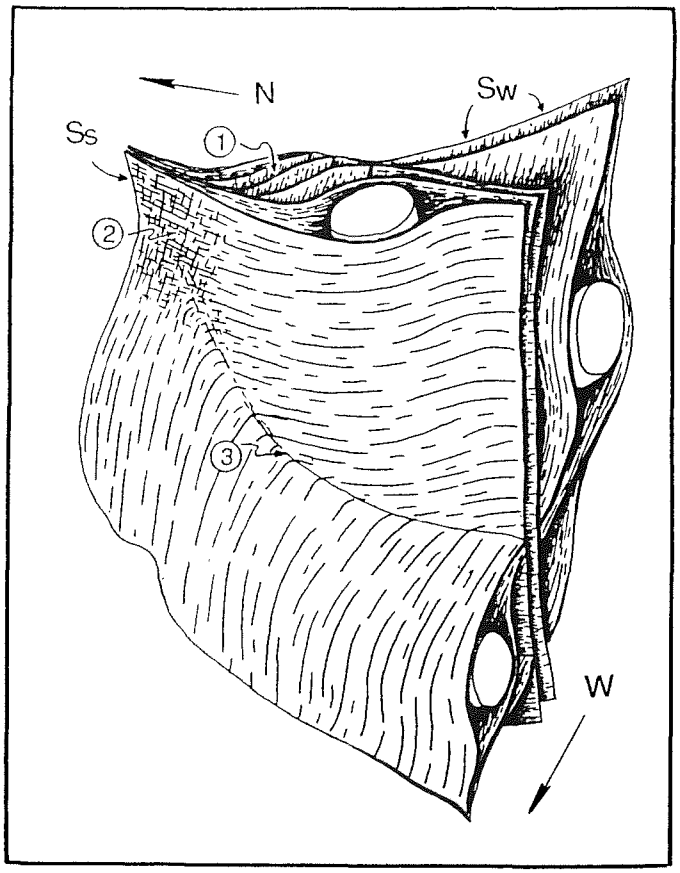

Figure 8. Schematic sketch showing the relationship between L_W, L_S, and the foliation surfaces that contain these lineations, referred to here as S_W and S_S respectively. The relationships illustrated are a composite of field and thin section observations. The scale of the diagram is variable and depends on the scale of partitioning of the later phase of deformation. The lines shown on foliation surfaces represent the orientation of lineations on that surface. The angle between the two sets of folia are exaggerated. In this sketch Ls and S_S form the younger cross-cutting fabrics. Numbered areas 1, 2, and 3 are referenced in the text.

small so that only one dominant foliation is observed in the field and only rarely can both be clearly distinguished in thin section. The two sets of folia have complex patterns of intersection due to nearly orthogonal axes of curvature, governed by the orientation of their associated lineation (Fig. 8) and a small, but variable angle of intersection.

The numerous places both in the field and in thin section, where both lineations are observed on essentially the same foliation surface, represent regions of intersection of the two sets of folia, S_W and S_S. Within these regions, one lineation might cross cut (Fig. 8 - location 2) or curve continuously into (Fig. 8 - locations 1 and 3) the orientation of the other lineation. The scale of these regions of intersection and the size of single lineation domains depends on the scale of partitioning between high and low strain domains of the later-formed fabric and on the local angle of intersection (smaller angle = larger region of intersection). A region of intersection might make up part of a thin section or could encompass several outcrops.

Timing relations between the two sets of fabrics were generally difficult to determine. The best places to make such determinations were in the regions of intersection of the two sets of folia. Where they could be determined with some reliability, a consistent age relationship between the two lineations, L_W and L_S, was not observed. Most fabrics indicate that L_W is older than L_S, some others suggest the opposite. The interpretation made here is that the deformation producing the backfold-stage lineation, L_W, is generally older than that producing the dome-stage lineation, L_S, with an overlapping transition. Regional observations are consistent with this, indicating that backfold-stage fabrics are deformed by, and thus older than dome-stage structures (Robinson et al., 1982; Peterson, 1984; Berry, 1989).

The most apparent difference between the two lineations, L_W and L_S, is the relationship between each lineation and the local maximum concentration of quartz c-axis orientations. Three possible interpretations for this difference have been suggested by Peterson and Robinson (in press). The first possibility is that the second deformation, associated with L_S, produced mylonitic fabrics and a complete change in dimensional preferred orientation of

sillimanite and quartz, without significantly changing the lattice preferred orientation of quartz. Alternatively, deformation during formation of L_S may have only locally reoriented the lineation but totally homogenized the quartz lattice orientations throughout the area. The preferred interpretation (Peterson and Robinson, in press; Peterson, 1992) is that the change in c-axis patterns resulted from a change in the dominant glide direction in quartz, from prism slip parallel to <a> to prism slip parallel to <c>. The relationship between quartz c-axis orientations and L_W are consistent with non-coaxial strain (Lister and Hobbs, 1980) with the dominant slip direction in a prism plane parallel to <a> (Lister and Dornsiepen, 1982) with a probable component of <a> slip in the basal plane. The relationship between L_S and the local c-axis orientation is more consistent with the dominant slip in a prism plane parallel to <c>. The earlier L_W-forming deformation may orient the quartz c-axes so that their maximum is near the transport direction for the later deformation. It is possible that this reorientation favors <c> slip enough to produce the observed change in dominant glide direction between the two phases of deformation. Both phases of deformation seem to favor prism slip in quartz during deformation.

Conditions of Deformation

The typical peak phase assemblage in pelitic schists within the main shear zone is garnet + biotite + K-feldspar + sillimanite + plagioclase + quartz. Typical granulite facies schists with coexisting garnet and cordierite are observed in rocks just to the east of the main shear zone. Preliminary estimates from garnet-biotite thermometry suggest that peak metamorphic temperatures of rocks involved in the shear zone were as high as 750° C. Garnets from this shear zone are chemically homogeneous, except where retrograde ion exchange has occurred near adjacent biotites (Peterson, 1992), suggesting the absence of late metamorphic fluid (Tracy, 1982).

Textural observations indicate that peak metamorphic minerals make up the mylonitic fabrics, so that both phases of shearing took place after the peak of metamorphism. The predominance of dynamic over static recrystallization textures may also reflect the continuation of deformation beyond the metamorphic peak. The scarcity of late metamorphic fluids during deformation may have been important in preventing both extensive retrograde reactions during shearing and extensive annealing after deformation ceased.

Some observations suggest that shearing deformation took place at relatively high temperatures. As mentioned above, the dominant deformation mechanism in quartz and feldspar within these rocks is dislocation creep. The mechanism of dislocation creep is active in naturally deformed feldspars at temperatures above 450° or 500° C (Tullis, 1983). The extensive recrystallization of feldspar under relatively dry conditions may suggest conditions of deformation that were comfortably above this limit. The apparent activity of prism slip mechanisms in quartz from these rocks gives further evidence for high temperature deformation conditions (see above discussion). Ave'Lallement and Carter (1971) showed that under laboratory conditions, prism slip is only active at temperatures greater than 800 °C. Lister and Dornsiepen (1982) suggest that this transition to prism slip under natural conditions takes place between 600°-700°C at 6 kbar. Preservation of retrograde ion exchange textures in garnets, only near adjacent biotite in fine-grained mylonites, may also suggest that deformation did not significantly alter the relative position of biotite with respect to garnet after attainment of garnet-biotite closure temperatures (≈500°C).}

The textural similarity between the two lineations and associated mylonitic fabrics suggests that the metamorphic conditions during both phases of deformation were essentially the same and possibly near peak temperature conditions for this area. This may indicate that the rocks in this zone remained near the peak of metamorphism for a relatively long time, or that the time between the two phases of deformation was relatively short.

Progressive Deformation

According to the criteria suggested by Tobisch and Paterson (1988), the fabric relations observed in the Conant Brook shear zone are consistent with a progressive deformation history from backfold to dome stage. The deformation style and prevailing metamorphic conditions for development of backfold and dome stage fabrics are virtually identical. The presence of only one dominant mylonitic foliation related to these two lineations within the shear zone suggests that the local X-Y plane of finite strain did not change significantly from backfold to dome stage. Thus, the change from backfold to dome stage requires only a change in transport direction in this area. This implies that the orientation of the principal finite strain axes remained relatively fixed, with the backfold-dome stage transition marked by a switch of the maximum and intermediate extension directions. The southwest orientation of dome-stage L_S lineations in the southern part of the shear zone may represent a transport direction intermediate between backfold-stage lineations and the final orientation of dome-stage lineations.

The Prescott Syncline, Greenwich Syncline, and Quabbin Park Fault

These structures will not be addressed on the field trip, however a brief description is useful for the discussion of the structural history below. The eastern margin of the Monson Gneiss has a complex geometry in the central Monson quadrangle. Between the thin easternmost belt of Monson Gneiss and the main belt of Monson Gneiss is a complex sequence of cover rocks. The eastern part of this sequence is interpreted as the southern extension of the Greenwich syncline, which appears to be truncated along the eastern margin of the Monson Gneiss (Fig. 5) (Peterson, 1992). The Greenwich syncline is defined as a stratigraphic syncline, however it may also be a structural anticline that exposes the overturned limb of a major backfold-stage fold in the eastern belt of Monson Gneiss (Robinson, personal communication). The Brennan Hill thrust may also be exposed in the southern part of the Greenwich syncline (Peterson, 1992). Truncation of the Greenwich syncline by the Conant Brook shear zone, which appears to have been active during both backfold and dome stage, suggests that formation of the Greenwich syncline and related folds was synchronous with and outlasted by dome-stage deformation within the shear zone.

A thin belt of Ammonoosuc Volcanics, west of the Greenwich syncline and essentially surrounded by Monson Gneiss, has been interpreted to represent the southern extension of the Prescott stratigraphic syncline, separated from rocks to the east by a fault, referred to here as the Quabbin Park fault (Fig. 5) (Peterson, 1992). The Prescott syncline (Fig. 2) is mapped along northwest margin of the Monson Gneiss and has been interpreted by Robinson (1963) to be an early backfold-stage structure. Identification of the Quabbin Park fault is based in part on offset of metamorphic isograds, map patterns in the Prescott syncline, and geomorphic features. It has not yet been identified in outcrop and much more work is needed to confirm its existence and character. Offset of metamorphic isograds across the Quabbin Park fault suggests that significant motion along this fault followed the peak of Acadian metamorphism.

Structural History and Tectonic Implications

A possible scenario for the development of structural features associated with development of the Conant Brook shear zone is illustrated schematically in the sequence of retro-deformed cross sections of Fig. 9. Constraints for construction of these cross sections are discussed in Peterson (1992). Initial Acadian deformation during the nappe stage produced large-scale, west-transported fold nappes. The easternmost nappe in Fig. 9a represents the North Orange band of Monson Gneiss. These fold nappes were cut by thrust nappes with continued westward transport (Fig. 9b). One of these, the Brennan Hill thrust, carried the North Orange fold nappe westward and, in some places, juxtaposed rocks of the Rangeley and Partridge Formations. During the early backfold stage, rocks in the Bronson Hill anticlinorium began to uplift, relative to the Merrimack synclinorium. At this time, the main body of Monson Gneiss was transported northward in the core of a large fold (Fig. 9c). This produced the Prescott stratigraphic syncline and the overturned stratigraphic section observed in the Mount Grace area (Robinson, 1963). Oblique collision might have caused the main body of Monson Gneiss rocks to "squeeze" north as rocks in the Bronson Hill anticlinorium began to uplift.

With continued uplift, the rocks began to overturn to the east and the Conant Brook shear zone was initiated along a zone of focused strain, on the overturned limb of this major structure (Fig. 9d). The west-side-up direction of transport recorded within the shear zone is consistent with this backfold-stage westward transport. During backfold-stage deformation, the Conant Brook shear zone may have truncated or attenuated the North Orange band of Monson Gneiss and Brennan Hill thrust along its west margin and the Acadian intrusive bodies along its east margin. Formation of the Quabbin Park fault (QPF) is shown at this time with a west-side-up motion. If the Quabbin Park fault is interpreted to be a later-formed structural feature, the sense of offset on the fault might be different, but the gross geometry shown in Fig. 9 would probably be similar.

Shearing within the Conant Brook shear zone continued from backfold to dome stage, but with a shift in the direction of transport from an east-west dip-slip motion to a north-south strike-slip motion, parallel to the trend of the orogen (Fig. 9d, e). This transition appears to be fairly rapid and progressive. Large-scale, upright, dome-stage folds, including the Greenwich stratigraphic syncline, may have formed during the early dome stage in response to dextral motion along the Conant Brook shear zone. Truncation of the Greenwich syncline by the Conant Brook shear zone suggests that shearing continued beyond this folding event (Fig. 9e).

The cross sections of Fig. 9 suggest that the backfold stage, not the dome stage, was primarily responsible for uplift of the basement gneisses of the Bronson Hill anticlinorium. The most intense deformation related to the

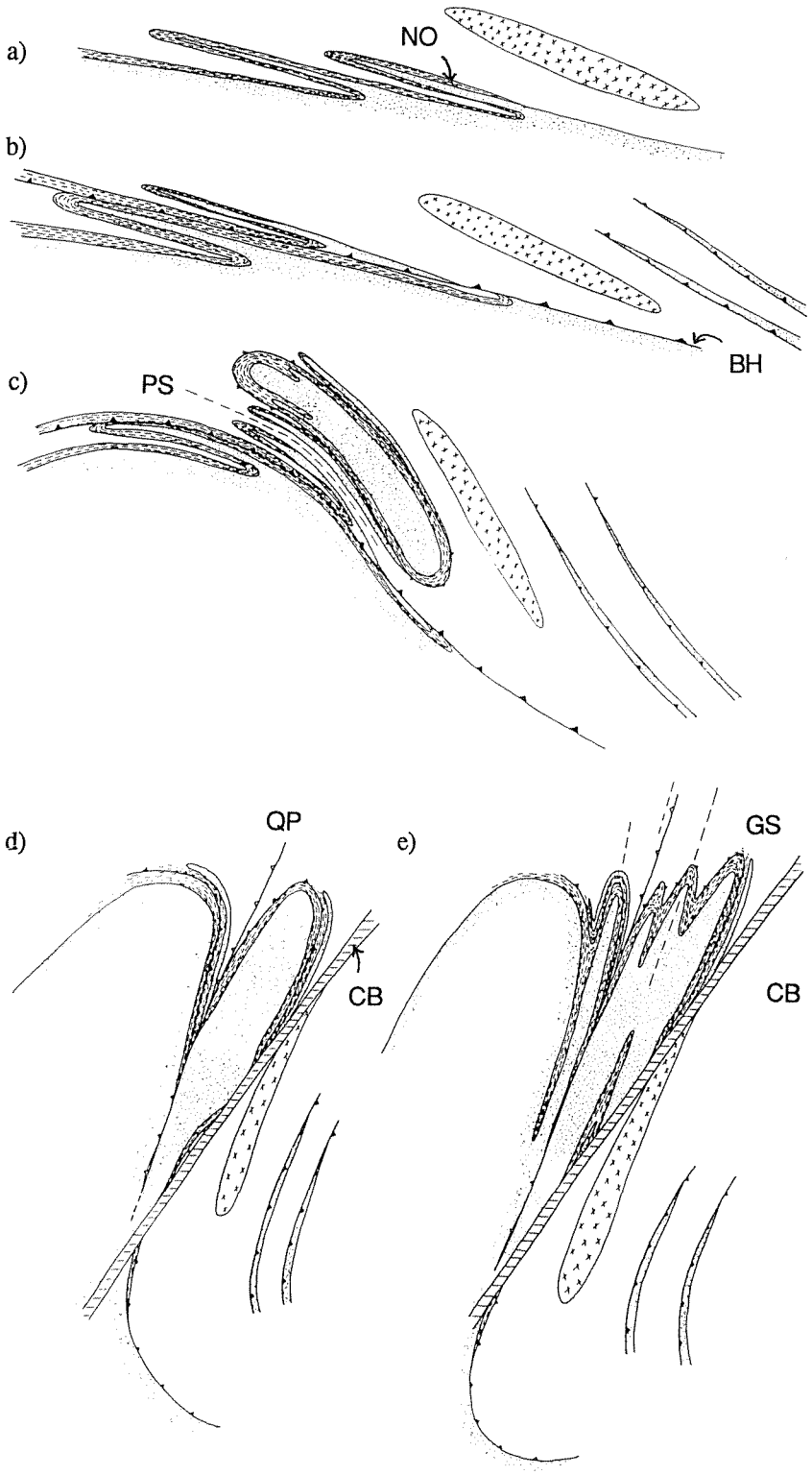

Figure 9. East-west retrodeformed cartoon cross sections through the south-central Massachusetts showing the possible sequence of development of major structures. a) Fold nappe stage. b) Thrust nappe stage. c) Early backfold stage. d) Late backfold stage. e) Dome stage. Monson Gneiss = stippled; Bronson Hill belt cover = dashed pattern; Merrimack belt strata = unpatterned; Acadian plutons = x's; NO = North Orange Nappe; BH = Brennan Hill thrust; PS = trace of Prescott stratigraphic syncline; CB = Conant Brook shear zone; QP = Quabbin Park fault; GS = Greenwich stratigraphic syncline.

backfold stage was concentrated in rocks of the Merrimack synclinorium as the Bronson Hill rocks rode up and over them (Fig. 9). Peterson and Robinson (in press) have suggested that the north-south, orogen-parallel lineations produced during the dome stage are a product of localized orogen-parallel collapse initiated by overthickening in the Bronson Hill and continuing oblique convergence in the late stages of the Acadian orogeny. In the Pelham dome area, this history appears to be complicated further by late Paleozoic deformation and metamorphism.

ACKNOWLEDGEMENTS

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ROAD LOG

Assemble at the large parking lot of the Crystal Springs Market Place, west of the intersection of Rte 9 and Rte 202 in Belchertown, north of the center of town.

Mileage

0.0 Road log begins at the traffic light at the intersection of Rte 9 and Rte 202. Head east on Rte 9 toward Ware.

10.4 Blinking light, intersection of Rte 9 with Rte 32, Town of Ware. Continue straight through light. Be careful - this is a dangerous intersection and all other traffic has the right of way.

10.5 Stop light. continue straight.

10.6 Stop light, intersection with West Warren Road. Turn right (south) onto West Warren Road.

13.3 Right turn onto Bacon Street (also marked by sign for Nenameseck Sportsman's Club). We will make an immediate right off Bacon Street to park in the lot of the Sportsman's Club.

STOP 1. PATTAQUATTIC TRAIL AREA - "THE SLOT" (≈ 2.5 HOURS) (Palmer quadrangle). The Pattaquattic trail area (Fig. 6), south and west of Bacon Street and northeast of Pattaquattic Hill was nicknamed "the slot" by Peter Robinson because a significant proportion of the belt of rocks mapped by Field (1975) in the Ware area, appears to squeeze south into this narrow, kilometer-wide belt or "slot" between the Monson Gneiss and the intrusive bodies to the east (Figs. 2 and 5). The "slot" area marks the northernmost extension of the focused deformation that characterizes the Conant Brook shear zone. The purpose of this stop is to traverse the Conant Brook shear zone from east to west in order to examine the nature of deformation within the zone and to begin to get a handle on the scale of deformation partitioning within this zone. Both lineations are well developed in this area (Fig. 10), however, one or the other is typically dominant in any one outcrop. We will also examine the character of the Monson Gneiss near its eastern contact with the shear zone and compare the styles of deformation in the gneiss and schists. The round-trip hiking time is about 40 minutes. We will make several short stops to look at the rocks as we traverse the shear zone from east to west. There is a lot of outcrop in this area.

FOOT LOG:

Head west on foot along Bacon street to a place where two dirt paths intersect Bacon Street. We will follow the Pattaquattic Trail (old Tavern Road), which takes us west (straight ahead) as Bacon Street veers northward.

Stop 1A. Stop at the first large outcrop just to left (south) of the trail. This is a gray-weathering, graphitic biotite-quartz-orthoclase-plagioclase-garnet-sillimanite schist. The underside of this outcrop offers a good view of the steeply west-dipping foliation surface. On this surface, a strong shallow, south-plunging stretching lineation is evident. This is defined by alignment of coarse, prismatic sillimanite and stretched quartz. A second steeply west-plunging lineation might be observed locally in this outcrop. Viewed parallel to the shallow lineation and perpendicular to foliation, particularly on the top surface of this outcrop, kinematic indicators, which include asymmetric tails on porphyroclasts and asymmetric melt patches, give a consistent dextral sense of shear. A lens of calc-silicate can be seen on the north side of the outcrop.

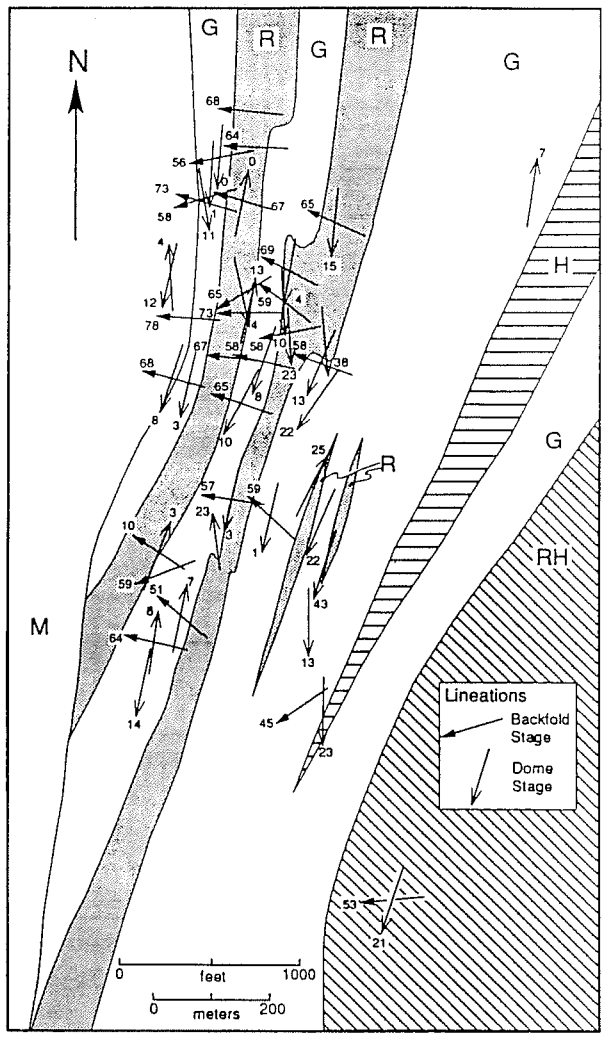

Figure 10. Map of area south of Bacon Street in the northeast part of the Palmer quadrangle showing the distribution of representative lineation orientations. This map covers approximately the same area as the map in Fig. 6. Note that the backfold and dome-stage lineations in this area are nearly orthogonal. Patterned rock units include: M - Monson Gneiss (dark shading); G - gray-weathering schist of the Rangeley Formation (no pattern); R - sulfidic rusty-weathering schist of the Rangeley Formation (light shading); H - Hardwick Pluton (horizontal rule); RH - Gneiss of Ragged Hill (diagonal rule).

Continue west along the trail. We will cross through a thin belt of sulfidic, rusty weathering schist of the Rangeley Formation. This is exposed in low outcrops to the north and south of the trail, however we will not stop to look at them on this trip.

Stop 1B. This is the first relatively large knob of rock on the right (north) side of the trail. It is predominantly held up by pegmatite, but has some interesting bits of schist clinging to its sides. On the near (east) side is a mylonitic schist cut by a thin pseudotachylite. In outcrop, the pseudotachylite appears as a dark green to black flinty, near-vertical layer with some thin offshooting veins. In thin section, it is extremely fine-grained and brittly deforms the mylonitic foliation in the rock. Injection veins are present locally. Pseudotachylites found ≈ 5 miles northwest of here in the Quabbin Reservoir spillway are thought to be Mesozoic in age. Along the back side of this pegmatite knob is a poorly exposed bit of very fine-grained highly deformed schist with few remaining porphyroclasts, classified as an ultramylonite. Ultramylonites are rare within the schists in the Conant Brook shear zone as continuous recrystallization of the micas appears to absorb much of the deformation. Here, the schist may have been buttressed against the stiffer pegmatite body to produce locally very intense deformation.

Just to the west is a second knob, dominated by a fine-grained feldspar-quartz-rich granulite. This rock contains small pink garnets and much less biotite than the surrounding schists. In outcrop it does not appear as deformed as the schists. In thin section quartz and feldspar appear highly deformed. Quartz is strongly recrystallized with serrate grain boundaries and plagioclase shows deformation twins. Deformation in these granulites is also strongly partitioned into thin biotite-richer layers. These layers are much finer grained than the quartz-felspar-richer layers and

locally display composite C/S fabrics or shear bands. This granulite layer is also found along strike to the north. A strong rodding lineation can be seen locally in these rocks.

Across the main trail from these two knobs is a secondary path that heads south. Follow this for a short way to the first large knob on the right (west).

Stop 1C. This is another outcrop of dominantly gray-weathering graphitic schist. There are two main points to this stop. First, the outcrop is dominated by a steeply west-plunging lineation. This can be seen on the foliation surfaces, defined by coarse sillimanite and stretched quartz. Kinematic indicators related to this lineation suggest a west-side-up sense of motion. Note that, other than orientation, the shear fabric in this rock is not noticeably different from that observed at the first stop.

Secondly, just up the hill to the west and south, is another outcrop that contains a layer of quartz pebble conglomerate and quartz grit. These conglomerates are characteristic of the Rangeley Formation further to the north (Thompson, 1985; Springston, 1990; Robinson et al., 1988). This is one of the few conglomerate exposures of the Rangeley formation found this far south. Please do not sample it. Return to the main trail and continue west. On the left is an outcrop of rusty-weathering sulfidic schist.

Stop 1D. This is a very brief stop to look at rocks in the westernmost thin belt of rusty-weathering schist mapped in this area. This rock is typical of much of the rusty-sulfidic schist of the Rangeley Formation in the region. Continue west about 100 meters.

Stop 1E. On the right are a series of low outcrops of gray-weathering graphitic schist. Notable in this series of outcrops is a dominant, shallow south-plunging lineation, a strong shear fabric, and a dextral sense of shear. The next low ridge across the small gully to the west is Monson Gneiss, so that the contact is constrained here within a few meters. Stop briefly at the outcrop of Monson Gneiss just south of the trail. This outcrop is dominated by pegmatite, but contains plagioclase gneiss with thinly interlayered amphibolite, typical of the Monson Gneiss in this area. Follow the ridge of Monson Gneiss north.

Stop 1F. There are a number of features to note as we walk along this ridge of Monson Gneiss. In the Pattaquatic trail area, the Monson Gneiss is generally thinly interlayered with fine-grained amphibolite. Garnet, which is generally uncommon in the Monson Gneiss, is locally abundant here. The style of deformation of the gneiss is very different from that in the schists, just to the east. The gneiss contains a strong foliation that has been folded by a series of tight to isoclinal, asymmetric similar folds. The folds commonly have steep, strongly curving hinge lines and a consistent asymmetry indicating a dextral, west-side-north sense of motion. Figure 11 shows data from one of these folds with a curving hinge line that folds a strong shallow stretching lineation. These folds apparently formed during late dome-stage deformation (Peterson, 1992). The north end of this ridge is a small pavement showing asymmetric folds and possibly refolded folds as well as localized zones of shearing. Pegmatites

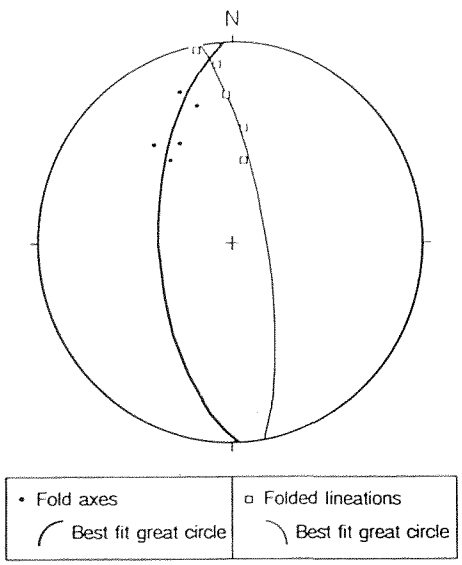

Figure 11. Equal area diagram showing fold axis and lineation data from one fold in the Monson Gneiss, just west of the contact with schist in the area south of Bacon Street (see Fig. 6 for location). Closed circles are fold hinge orientations for a single fold. Open squares are the orientations of a quartz-feldspar stretching lineation measured around this fold. The best fit great circle to the fold hinge data is the axial plane of the fold. The intersection of this plane with the plane containing the folded lineations gives a shallow north-plunging movement line (Trend/Plunge = 353, 09).

are common in this outcrop. In one of them, we can see strong shearing and development of an ultramylonite. These pegmatites appear to be relatively soft to the strain relative to those in the schists just to the east.

Return along the trail to the vehicles.

Return, via Bacon Street to West Warren Road and turn right.
15.0 Junction, Rte 67. Turn right.
16.2 Low outcrops on the right are Coys Hill granite gneiss.
16.4 Turn right on Old Warren Rd.
16.9 Large outcrop on right (west) side of road. Carefully cross the road and park on the grass on the left (east) side of the road.

STOP 2: GNEISS OF RAGGED HILL (≈ 20 MINUTES) (Palmer quadrangle). This outcrop marks the southernmost exposure of the Gneiss of Ragged Hill and the approximate eastern margin of the Conant Brook shear zone. The map pattern suggests that the Gneiss of Ragged Hill is apparently terminated southward by the shear zone. The outcrop also preserves a spectacular strain gradient from north to south. In general, the gneiss in this outcrop is fairly typical of the Gneiss of Ragged Hill, except that here it contains a much stronger foliation. Start at the northern end of the outcrop and work south. Note the garnet and locally sillimanite present in the gneiss. Both lineations are locally well developed in this outcrop, however, the shallow dome-stage lineation is dominant. At the south end of the outcrop, the foliation is a bit shallower and the strain intensity increases toward the base of the outcrop. Following this high strain zone south, it becomes a flinty ultramylonite with a few remaining porphyroclasts of garnet and feldspar. It is difficult to see a lineation within this ultramylonite in the field, but a strong shallow quartz-feldspar-sillimanite stretching lineation is obvious in foliation-parallel thin sections. Relationships are a bit complicated by a late brittle fault near the south end of the outcrop, but it is still possible to follow the transition from coarse gneiss to very fine ultramylonite.

Continue south and west along Old Warren Road.
17.6 Large outcrop both sides of the road. Park on the right (north) side of the road, beyond the first large outcrop.

STOP 3: HIGHLY STRAINED PELITIC SCHISTS, RANGELEY FORMATION (STATION P100) (APPROXIMATELY 30 MINUTES) (Palmer quadrangle). This outcrop offers a spectacular, fresh, 3-dimensional cross-strike exposure of both graphitic and sulfidic schists of the Rangeley Formation within the Conant Brook shear zone and preserves domains in which each of the two lineations and related fabrics are dominant. This has made it an important laboratory for exploring the character of these two fabrics, their relationship to each other in time and space, and the nature of partitioning between the two.

There is a contact in this outcrop between graphitic schists to the west and sulfidic schists to the east. Most of the fabric work was done on the graphitic schists because the samples tend to be fresher. Throughout the outcrop, these schists contain abundant inclusions of pegmatite, granite or granitic gneiss that have been extremely disaggregated and/or attenuated. Most of these now form strongly asymmetric lenses that suggest a sense of shear consistent with other kinematic indicators in the rock (discussed below). These disaggregated bodies emphasize the extreme deformation imposed on these schists during shearing. A few of the less deformed (younger?) pegmatites contain coarse muscovite, which is otherwise not present in these schists. In thin section, these coarse muscovite grains are randomly oriented and appear to have grown after deformation ceased in these rocks. In contrast, the quartz within these pegmatites is deformed, with serrate grain boundaries, suggesting that the pegmatites were intruded prior to the cessation of deformation. The schists surrounding these muscovite-bearing pegmatites is sericitized. These textures suggest that some of these more continuous and possibly late pegmatites may have served as conduits for late fluids, causing very localized retrograde hydration.

One of the most important features of this outcrop is the distribution of lineations and related shear fabrics. Approximately 8 - 10 meters from the west end of the outcrop the schists are dominated by the steeply west-plunging backfold-stage lineation. In this part of the outcrop, there is little or no evidence of the shallow lineation. This steep lineation is evident on foliation surfaces both in the field and in thin section and is defined by abundant coarse sillimanite, elongate quartz grains and the stretching direction of porphyroclast tails. On the front surface of the outcrop, parallel to this lineation, numerous kinematic indicators suggest a consistent west-side-up sense of motion. West of this, foliation surfaces commonly reveal both lineations. At the west end of the outcrop, a strong

shallow south-plunging mullion-like lineation is evident on the foliation surface. In detail, this lineation is also defined by aligned sillimanite and stretched quartz. Thin sections cut approximately perpendicular to each of the finite strain axes show no evidence in this part of the outcrop for the steep lineation. On top of the outcrop at the west end, numerous kinematic indicators show a dextral or west-side-north sense of shear.

Observations made in outcrop and thin section across this transition from a domain of steep lineations to a domain of shallow lineations suggest the following relationships:

1) In the field, both lineations share essentially the same steeply west-dipping anastomosing mylonitic foliation. In detail, however, the orientation of the dominant foliation forms an enveloping surface to a composite foliation consisting of two sets of anastomosing folia related to each lineation (Fig. 8).
2) The style of shearing deformation associated with the two lineations is indistinguishable.
3) There is no apparent difference in metamorphic conditions between these two phases of deformation.
4) It is very difficult to determine the age relationship between these two fabrics, however fabric relations suggest that the steep lineation is generally older with some overlap.

Quartz c-axis orientations were measured in several thin sections from this outcrop as well as in other parts of the shear zone, particularly where only one lineation could be found in thin section. The results of this study are discussed in the text.

	Continue west along Old Warren Rd.
19.8	Turn left onto Rte 32 (south).
19.9	Make a left turn onto Breckenridge Rd. where Rte 32 veers sharply right. PLEASE BE CAUTIOUS!! Visibility is poor around this curve and traffic is generally moving fast.
22.0	Right turn onto Rte 20 (east).
22.7	Stop light (Palmer, MA). Turn left (Rte 32 detour).
22.8	Turn left onto Rte 32.
25.5	Angle left onto Tobey Rd.
25.8	Turn left, up hill, just before factory gates (formerly Church toilet seat factory) onto Beebe Rd.
26.5	Turn right onto Town Farm Rd.
26.9	Turn left onto Brimfield Rd.
27.6	Turn right onto East Hill Rd.
28.6	Turn right into parking lot of Westview Farm Creamery. Driveway is between the house and the Creamery. Park in the rear part of the lot.

STOP 4: EAST HILL TRAVERSE, WESTVIEW FARM (≈ 2 HOURS) (Monson quadrangle). The purpose of this traverse is part stratigraphic, part structural, and part petrologic. The hill that we will traverse is one of the few places in the Palmer and Monson quadrangles where stratified rocks other than the Rangeley Formation have been identified within the Merrimack belt, west of the Coys Hill pluton. The rocks that we will see on this traverse are outside of the main part of the Conant Brook shear zone. Although a number of the rocks observed along this traverse are highly strained, the scale of partitioning between highly deformed and less deformed rocks is greater than that observed within the shear zone. In part this may be a reflection of the different rock types in this area. We also have the opportunity on this traverse to observe the deformation character of the Coys Hill Granite. Cordierite-garnet-bearing schists, typical of metamorphic zone VI are present in the eastern part of this traverse. These are the only cordierite-bearing schists that we will see on this trip.

We will conduct the traverse from west to east, across strike, beginning in the Rangeley Formation and ending in the main belt of Coys Hill Granite.

FOOT LOG:

Walk north from Creamery along East Hill Rd. and then cut east across the pasture to a path that leads eastward, approximately 100 meters north of the man-made pond that is partly hidden by the woods east of the pasture.

Stop 4A. Just after entering the woods, outcrops to the left (north) are 'typical' gray to reddish-weathering feldspathic schists of the Rangeley Formation. Note that these rocks are not as highly strained as those seen earlier in the day. Continue east along the path to a place where a narrow N-S-trending ridge juts out toward the path ≈ 500'.

Stop 4B. The rock that follows the ridge is an extremely rusty, sulfidic calc-silicate granulite with abundant graphite and pyrrhotite and very little mica. It typically has a slabby or blocky appearance and is locally quartz-rich. This rock is very similar to rocks described by Field (1975) to the north in the Ware area as part of the Fitch or Francestown Formation (Robinson and Goldsmith, 1991). Just to the east of the ridge is a foliated biotite tonalite. These two units can be traced up hill along strike to the north for several hundred meters, but are lost to lack of outcrop both to the north and south. Follow the ridge north along strike for a short distance to look at the character of the Francestown Formation and then head due east to a relatively large outcrop ≈ 150' across strike.

Stop 4C. More of the Rangeley Formation, very similar to that seen in the first outcrops of this traverse. This thin belt can be traced to the north and south from here. Continue east to an overhanging outcrop along the crest of the ridge.

Stop 4D. Enter the outcrop from the south end. This is a second and more continuous belt of Francestown Formation. The best exposures are found on the steep east slope of the ridge. Similar to the previous belt, these rocks are very graphite and pyrrhotite-rich and mica-poor and appear slabby or blocky. Some quartz-rich layers appear bluish on a fresh surface. This belt of Francestown Formation is approximately 80' (25 m) thick. Follow the ridge northward about 160' to a place where a large gray outcrop trends downhill to the east.

Stop 4E. The eastern contact of the Francestown Formation with a gray-weathering, graphitic, big garnet-sillimanite-rich pelitic schist to the east is complex. It is marked by a fine-grained mylonite, a few meters thick, which can be seen at the western, uphill end of this large outcrop. Continue north along strike for another 100'.

A thin belt of reddish-weathering feldspathic schist, similar to that observed in the Rangeley Formation, just to the west, is found locally between the mylonite and the Francestown Formation. The Francestown also appears to thin to nothing northward in this vicinity. A thin sliver of highly deformed Coys Hill granite is also found along the western contact of the mylonite in one or two places, just to the north. These complex relations may indicate that this contact may have been slivered by faulting during or prior to mylonite formation. Walk downhill to the east ≈ 350' through this gray-weathering schist.

Stop 4F. Large outcrop of garnet-sillimanite-rich schist. These rocks typically have abundant sillimanite and large garnets. In some places the garnets are rimmed by felsic melt patches. In this outcrop, the rock contains a strong sillimanite lineation with a steep west plunge. This aluminous garnet-rich schist is found along the western Coys Hill contact as far north as the Monson - Palmer quadrangle boundary and south to the Conant Brook "reservoir" area. This unit is tentatively correlated with the uppermost Silurian Warner Formation (see discussion in text) and is approximately 330' (100 m) thick in this area.

Stop 4G. The contact with the Coys Hill granite lies just below and east of Stop 4F. The Coys Hill granite is a coarse-grained, heterogeneously deformed garnet-biotite-microcline granite gneiss with large, distinctive K-feldspar megacrysts. In these outcrops it is relatively biotite-poor and might easily be confused with a deformed pegmatite, however the shape of the feldspar megacrysts and the abundant red garnets are distinctive. Where it is highly deformed, as in these outcrops, the feldspar megacrysts form porphyroclasts with finely recrystallized asymmetric tails. Garnets also form porphyroclasts, typically with mica-rich matrix tails. The matrix is dominated by finely recrystallized biotite and ribbony quartz. Very fine-grained mylonites or ultramylonites are rare within the Coys Hill granite gneiss. On this hill, the eastern contact of the Coys Hill granite with the Warner schists is decorated with pods of a medium- to coarse-grained tonalite or diorite. The Coys Hill is about 160' (50 m) thick here and is in contact to the west with the diorite of West Warren, which we will see at Stop 5.

Summary of Stop 4. The sequence of rock types exposed on this hill is similar, although not identical, to the sequence described by Field (1975), west of the Coys Hill granite, on Ragged Hill in the Ware area. The rock unit correlations made here are based on correlations made by Field (1975), Robinson et al. (1982), and Robinson and Goldsmith (1991). This sequence was not observed elsewhere in the Monson and Palmer quadrangles. As mentioned above, the garnet-rich, upper Warner-equivalent schists can be traced along strike for a few kilometers, however no other exposures of the Francestown formation were observed. This may be due to lack of exposure or discontinuity of the unit. The structure responsible for repetition of the Francestown Formation on this hill is not known, but the stratigraphic sequences and geometry are most consistent with some kind of fault repetition. The slivers of Coys Hill granite and Rangeley-type schists along the mylonite contact between the Francestown and Warner may be indicative of early faulting. Thompson (1985) has suggested that an early, nappe-stage thrust, the

Chesham Pond thrust, may follow the western contact of the Coys Hill granite south from New Hampshire. Perhaps these hints of early slivering or thrusting near the Coys Hill contact are clues to this early thrusting.

Follow the foot and horse path at the base of the hill south and then west back to the vans.

Turn right (south) from Westview Farm Creamery onto East Hill Rd..
29.5 Turn left onto Munn Rd.
30.1 Large outcrop on left side of road. Park on right side of the road, beyond the outcrop.

STOP 5: THE DIORITE AT WEST WARREN (≈ 20 MINUTES) (Monson quadrangle). This outcrop marks the southernmost exposure of the diorite at West Warren. We are very near its western contact, as the ridge that you can see in the woods to the west is held up by Coys Hill Granite. The dominant rock type in this outcrop is a medium- to fine-grained biotite-hornblende-plagioclase diorite gneiss, locally with garnet and hypersthene. At the south end of the outcrop a discrete, moderately west-dipping shear zone deforms the dominant steeply west-dipping foliation. The shear sense observed in outcrop and thin section is west-side-up. The dominant lineation in this rock is steeply west plunging, but a bit recalcitrant. This zone is interpreted to be a late backfold-stage shear zone, formed at the same time as shearing within the Conant Brook shear zone. The dominant foliation may be formed during the nappe stage, early backfold stage, or as a composite of the two. The shear zone is a very fine-grained biotite-rich mylonite to ultramylonite, locally with anastomosing quartz ribbons. The relative abundance of quartz within this shear zone suggests that high strain may have been partly accommodated by a layer or vein of tonalite. Other shear zones within this outcrop incorporate pegmatites. The relative softness of quartz to deformation may have helped to localize shearing. Three aspects of this outcrop are important here. First, the fine-grained nature of the sheared rocks in this outcrop are fairly typical of the character of shearing deformation within this unit. Secondly, this type of localized deformation is demonstrative of the larger scale of deformation partitioning observed outside of the Conant Brook shear zone. Thirdly, the geometry of a shallower shear zone cutting an earlier steeper foliation has been observed in a number of places regionally and may be characteristic of the overall geometric relationship of the Conant Brook shear zone to the external fabrics.

Turn around and head back west on Munn Rd.
30.7 Intersection with East Hill Rd. Continue straight ahead on Munn Rd.
31.7 Sharp left turn onto Wales Rd.
32.7 Conant Brook Dam entrance. Turn left.
33.0 Park in small lot by Conant Brook Dam.

Walk across the top of the Dam to the north side.

STOP 6: THE CONANT BROOK DAM SPILLWAY (≈ 1 HOUR) (Monson quadrangle). The Conant Brook Dam spillway provides a spectacular cross-strike exposure of highly strained schists within the Conant Brook shear zone. It is after these exposures that the shear zone is named. The spillway is dominated by an extremely rusty sulfidic schist with a very strong steep stretching lineation. In places, particularly in the upper part of the spillway, the rocks are essentially L-tectonites. Compared with the rest of the shear zone, tight to open folds, with axes parallel to this lineation are common in these outcrops. These folds produce a fairly complex local map pattern in the contact between the rusty schists to the east and the gray, graphitic schists that dominate the west end of the spillway. It is difficult to merge this local pattern with the regional map pattern due to generally poor bedrock exposure surrounding the spillway. Calc silicate boudins or lenses are common within the rusty schists in the spillway and a few lenses of felsic gneiss and amphibolite are present near the contact between gray and rusty schist in the central part of the spillway. There are parts of the spillway outcrop where the steep lineation is not as well developed and the shallower lineation can also be seen. Whereas the two lineations are nearly orthogonal in the northern part of the Conant Brook shear zone, they are closer in orientation in the southern part of the shear zone (Fig. 7b). In this and surrounding outcrops, the backfold-stage lineation has a steep west plunge and the dome-stage lineation has a moderate to shallow southwest plunge.

Walk across the dam and then take a right to start at the east end of the spillway. We will make a quick east-to-west traverse of the spillway to look at some specific characteristics. The eastern part of the spillway is dominated by extremely rusty-weathering sulfidic schists. Near the east end of the spillway on the south side, the steep lineation is very strong. Across the spillway, the schist is interlayered with calc-silicates. The steep lineation is still strong, but the foliation is also well developed. Folds with axes parallel to the lineation are also present and are best defined in the calc-silicate layers. On some surfaces in this area, domains of randomly oriented sillimanite,

probably representing low strain domains, can be seen to be surrounded by higher strain domains of strongly oriented sillimanite.

Continue west and climb over the spillway lip. With careful observation, a shallower, southwest-plunging lineation can be seen, along with the steep lineation, on foliation surfaces in some of the outcrops on the north side of the spillway, below the lip. Folded calc-silicate lenses or boudins are also present in these outcrops.

Continue west. The rusty-weathering outcrops that you pass to the north contain steeply plunging folds, with axes parallel to the steep lineation. The contact with gray-weathering, graphitic schists is encountered where the spillway begins to open up on the left (south). Near this contact are lenses or discontinuous layers of felsic gneiss and amphibolite. The stratigraphic and structural implications of these rocks is not clear.

Continue west to the first good exposure of gray-weathering schist on the north side of the spillway. The steep lineation is still well developed in these rocks and it is easier to see kinematic indicators, which suggest a west-side-up sense of shear. Similar to Stop 3, this schist contains disaggregated lenses of pegmatite, granite, and granitic gneiss, suggesting very high strain.

Return to the vans.

Return to Wales Rd.
33.3 Turn left onto Wales Rd.
34.1 Outcrops on both sides of the road. Park on the right side of the road.

STOP 7: WALES ROAD MYLONITE (Optional) (≈ 20 MINUTES) (Monson quadrangle). The best exposures here are on the north side of the road. This roadcut includes both gray-weathering graphitic schists and rusty-weathering sulfidic schists of the Silurian Rangeley Formation Garnet is less abundant in the rusty weathering schist. Garnet and feldspar form porphyroclasts in this highly deformed schist. In thin section, these have rounded shapes and asymmetric matrix tails. Coarse sillimanite and stretched quartz define a strong steeply west-plunging lineation in the outcrop and in thin section. Biotite is deep red brown. Quartz most commonly forms finely recrystallized ribbons. Composite asymmetric porphyroclasts and lenses of quartz and feldspar appear to be attenuated and disaggregated pieces of pegmatite. Granitic veins are are also disaggregated during deformation and in one place, on top of the outcrop, a vein is folded and truncated. The pegmatite at the east end of the outcrop contains quartz, plagioclase, K-feldspar, sillimanite and minor biotite. In thin section, the feldspar is strongly recrystallized and pulled apart into small boudins.

END OF TRIP. To return to Amherst or the Massachusetts Turnpike, reverse direction on Wales Rd and follow it to Rte 32. Take Rte 32 north to Palmer, where you can pick up the turnpike or follow signs north to Amherst.

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THE BERNARDSTON NAPPE AND BRENNAN HILL THRUST IN THE CONNECTICUT VALLEY AND THEIR ROOT ZONES EAST OF THE BRONSON HILL ANTICLINORIUM

by
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PURPOSE OF TRIP

Robinson, Thompson and Elbert (1991) summarized the development of the nappe theory in the Connecticut Valley region since Jim Thompson's first proposal in 1954. A recent development of this theory is that of thrust nappes, which depends on the stratigraphic and structural interpretation of three areas, the Mount Monadnock area, New Hampshire, (P. J. Thompson, 1985, 1988a, 1988b), the Bernardston-Hinsdale area, Massachusetts and New Hampshire (Trask, 1964; Trask and Thompson, 1967; Elbert, 1988), and the Mount Grace area, Massachusetts (Robinson et al., 1988) (see Figure 1). The work by P. J. Thompson in the Monadnock area established the Monadnock sequence stratigraphy and its correlation with strata in central New Hampshire and northwestern Maine (Hatch et al, 1983). The work by Elbert in the Bernardston-Hinsdale area resulted in the discovery of lowest Devonian conodonts (Elbert et al., 1988), extended the Monadnock stratigraphy to the west side of the Bronson Hill anticlinorium, and demonstrated the stratigraphic position of magnetite-garnet iron formation in the upper part of the Silurian Perry Mountain Formation. The work in both these areas provided a key for reinterpretation of the complex geology of the Mount Grace area by Robinson.

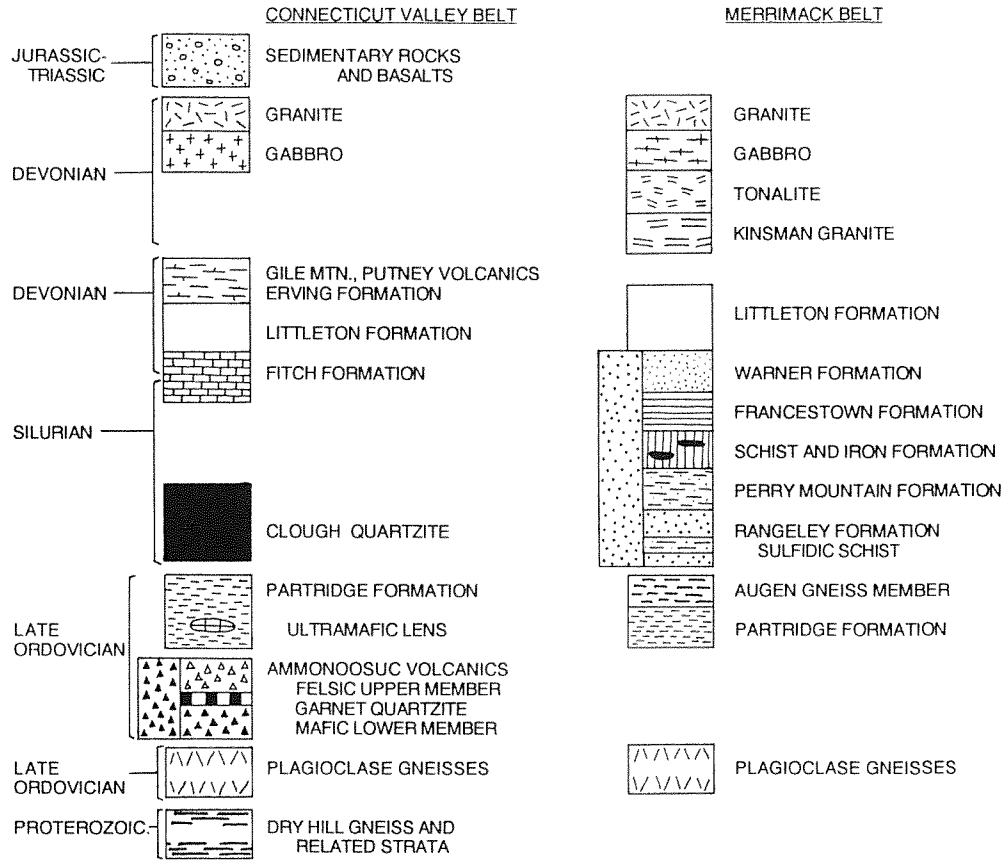
The purpose of this trip is to examine the stratigraphic and structural basis for direct linkage between the Bernardston-Hinsdale and Mount Grace areas, on opposite sides of the Mesozoic Connecticut Valley border fault (Figure 1). Emphasis will be placed on identification of the Bernardston nappe in its type locality (Stop 1), including recently discovered lowest Devonian conodonts, and in its root zone in the Mount Grace and Orange areas (Stops 3 and 6), and on strata of the Brennan Hill thrust sheet in both areas (Stops 2, 4, 5). There will also be an examination of some of the evidence for a basement-cored fold nappe within the Brennan Hill thrust sheet (Stops 8, 9). This is an amalgamation of selected stops from two separate trips run previously (Elbert, 1988; Robinson et al., 1988) and the reader is referred to these guidebook articles for more extensive details. Reference will be made here to figures previously published in Elbert et al. (1988) and Robinson et al. (1991) and a limited number of reprints will be available on the field trip.

STRATIGRAPHY

The major stratigraphic packages of significance to this field trip (see Robinson et al., 1991, Fig. 2) are the Proterozoic and Ordovician gneissic and related rocks in the cores of gneiss domes of the Bronson Hill anticlinorium, the Ammonoosuc Volcanics and Partridge Formations of the Ordovician cover sequence, and two sequences of Silurian-Lower Devonian strata, the Bronson Hill sequence and the Monadnock sequence. The Bronson Hill or Connecticut-Valley-belt sequence consists of thin Lower Silurian Clough Quartzite, metamorphosed sandstone and conglomerate, and Upper Silurian-Lower Devonian Fitch Formation, calc-silicate and marble, overlain by thick Lower Devonian Littleton Formation, metamorphosed marine turbidites. The Monadnock or Merrimack-belt sequence consists of a thicker metamorphosed Silurian sequence of Lower Silurian Rangeley Formation, shaley turbidites with conglomerates; Lower Silurian Perry Mountain, well bedded sandstones and shales with lenses of magnetite-garnet iron formation; Middle Silurian Francestown Formation, sulfidic calcareous shale with interbedded aluminous shale; and Upper Silurian Warner Formation, feldspathic and calcareous siltstone, again overlain by Littleton Formation. The Bronson Hill and Monadnock sequences of Silurian strata are considered to have been deposited on opposite sides of a "tectonic hinge" (Moench and Boudette, 1970) separating regions of slow and rapid deposition on the northwest margin of a major depositional apron or trough. The distinction between the Bronson Hill- and Monadnock-sequence strata is crucial to the development of the theory of thrust nappes.

The history of the Bernardston fossil locality, as well as the results of conodont studies, are summarized by Elbert (1988) and Elbert at al. (1988), and only briefly recounted here. Fossil remains were first discovered in the marble beds at the Williams Farm in Bernardston by Edward Hitchcock in 1833 (Hitchcock, 1833, page 295). Hitchcock's (1833) discovery was probably the first of pre-Carboniferous fossils in New England. Since then, the locality has been the subject of more than a dozen published contributions. Hitchcock later noted (1851) that James

Key to patterns in Figures 1, 4, 5, 6, 8.



Hall had correlated the Bernardston marbles with the Onondaga Limestone (Devonian) of New York based on crinoid fragments. Dana visited the locality and concurred with the Devonian age assignment (Dana, 1873, 1877). He also reported the first discovery, by B.K. Emerson in 1877, of fossils in the quartzites immediately above the marbles. This was followed by studies by Whitfield (1883), Clarke (in Emerson, 1898), Schuchert and Longwell (1932), Balk (1941) and Cooper and others (1942). Boucot and others (1958) restudied the specimens of Whitfield, Clarke, and Balk. They demonstrated that the fossils reported by Balk from a magnetite bed were not from Bernardston but probably from Lower Devonian rocks in Nova Scotia. They collected new samples of marble from outcrops and calcareous quartzite from the spoil piles of the small, open-pit iron mines at this locality (A.J. Boucot, oral commun., 1983). The marble produced favositids and crinoids, only indicating a post-Early Ordovician age. The calcareous quartzite yielded the brachiopod *Eospirifer* cf. *E. radiatus* (Sowerby) indicating a late Llandoverian to early Ludlovian age. Although the fossiliferous calcareous quartzite bed is no longer exposed, earlier studies clearly established that it was immediately above (stratigraphically below) the marble. Trask's (1964) mapping firmly established the correlation of the quartzite with the Clough Quartzite and the marble with the Fitch Formation even though the marble was included as part of the Clough Quartzite in a subsequent field trip guidebook (Trask and Thompson, 1967). Most recently, Elbert et al. (1988) have described an assemblage of Lochkovian (Early Devonian) conodonts from the marbles of the Fitch Formation at this locality.

The conodonts appear to be the world's largest, as well as highest grade (color alteration index = 8), collection of regionally metamorphosed conodonts. Over 1000 recognizable conodonts were recovered from 129 kg of dominantly quartzose marble (see Figure 3). They are all relatively poorly preserved, having reached at least 500°C. They are generally recrystallized, virtually all incomplete, but few specimens are significantly deformed. The fragmentation of conodont elements in these samples was primarily an effect of deposition in a high-energy environment and not a result of tectonism. The conodonts are indicative of the earliest Devonian *woschmidti* to *eurekaensis* Zones. They are Lochkovian (earliest Devonian), younger than the Pridoli conodonts reported by Harris et al. (1983) from the Fitch near Littleton, New Hampshire. These results plus the work of Boucot et al. (1958) establishing a Llandoverian to Wenlockian age for the calcareous quartzite which outcropped physically above the marbles, independently support the decades-old structural interpretation that the section at Bernardston is structurally inverted. They also strongly imply significant erosional intervals both before and after deposition of the Fitch in this location. On a lighter note, we call attention to new paleontological work suggesting conodonts were the earth's first vertebrates (Browne, 1992).

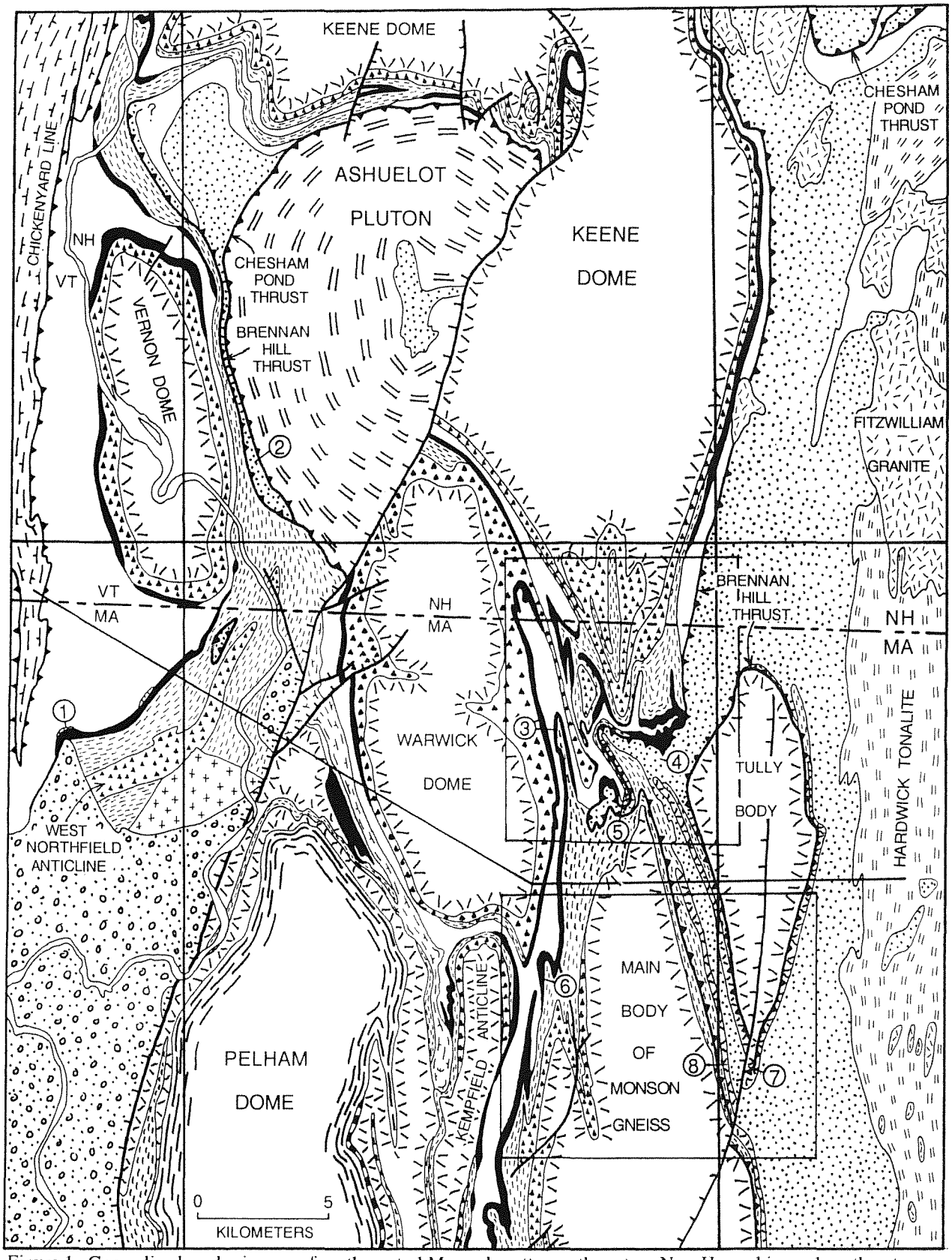


Figure 1. Generalized geologic map of north-central Massachusetts, southwestern New Hampshire and southeastern Vermont showing locations of stops for trip A-3 and areas of detailed Figures 5 and 8. Line of section of Figure 27 of Robinson et al. (1991) is indicated. Hachures are on the down-thrown side of Mesozoic normal faults; teeth are on original upper sides of Brennan Hill and Chesham Pond thrusts.

A particularly important aspect of this field trip is the stratigraphic sequence exposed at Biscuit Hill, Hinsdale, New Hampshire, and the key it provides for interpretations elsewhere. The sequence is all upside down and immediately beneath the Ashuelot pluton (Elbert, 1986). The section is extremely thin; estimated thicknesses of the individual stratigraphic units range from a few inches to ten meters. The characteristic succession of distinctive rock types, coupled with primary facing information from relict graded bedding, has led to the correlation of this stratigraphy with that of the Monadnock area (Elbert, 1986). The sequence begins with interbedded gray-weathering and rusty-weathering schist with schist-matrix and quartzite-matrix conglomerates and rare calc-silicate pods of the Rangeley Formation. The stratigraphic top of the Rangeley Formation contains a few thin quartzite beds and grades into the cyclically interbedded quartzites and gray schists of the Perry Mountain Formation. Graded beds in both the Rangeley and Perry Mountain confirm the topping direction. The top of the Perry Mountain contains biotite-rich, massive gray schist with beds and boudins up to 8 meters thick of fine-grained coticule (garnet granulite), coticule conglomerate, and magnetite-grunerite-garnet-apatite-quartz-graphite iron formation. Stratigraphically higher rocks at Biscuit Hill are well bedded, rusty-weathering, sulfidic graphitic calc-silicate granulites and interbedded sulfidic Mg-biotite schists of the Francestown Formation. This is stratigraphically overlain by well bedded, clean, actinolite-garnet-calcite calc-silicate gneisses, garnet granulites and interbedded purplish biotite granulites of the Warner Formation. The lower part of the Warner is generally richer in calc-silicate rocks and the upper part in biotite granulites. Numerous metamorphosed gabbro sills intrude the strata and are presumed to be related to the Mt. Hermon Hornblende Gabbro and the associated dikes and sills in the the Bernardston-Northfield area (Elbert, 1988).

The distinctive Perry Mountain iron formation is extremely important both to local mapping and regional interpretations. It is not significant in the Monadnock area itself, but here can be very precisely positioned within the Monadnock sequence It can be correlated with strata which had previously been assigned to the Littleton Formation in the Mt. Grace quadrangle (Robinson, 1967; Huntington, 1975), just above the root zone of the Bernardston nappe (Stop 5). Such a long-range correlation of such a thin units might seem questionable, but, when the Mesozoic listric motion on the Connecticut Valley border fault is restored, the two locations are only a few kilometers apart and at a comparable structural level.

STRUCTURAL GEOLOGY

The structural geology has been discussed in detail elsewhere and what is given here is only a thumbnail outline. The earliest structural features are a series of large-scale west-directed fold nappes developed in both the Bronson Hill sequence and in the Monadnock sequence. Those in the Bronson Hill sequence were first identified during the pioneering work of J. B. Thompson, Jr. , those in the Monadnock sequence by P. J. Thompson (Robinson et al., 1991). Subsequently, the recumbent-folded sequence was cut by several thrusts at an angle to the axial surfaces of the earlier fold nappes. The two most important thrusts, both first identified in the Monadnock area, are the lower Brennan Hill thrust and the higher Chesham Pond thrust (Figure 1). As a result of this circumstance, strata above or below the Brennan Hill thrust at any given point may be either right-side-up or inverted. This circumstance produces great difficulty in interpretation.

The next major phase of structural development, referred to loosely as backfolding, was a complex series of events that deformed all pre-existing features. In the area of this trip, the most striking feature of this episode was the northward and northwestward overturning of the main and Tully bodies of Monson Gneiss, so that the younger strata surrounding these bodies are commonly overturned. This northward overturning is displayed in the Mount Grace and Orange quadrangles (Figure 1), where all of the strata surrounding the main body of Monson Gneiss dip toward it, so that its northward termination has the three-dimensional form of the prow of a ship. Less obvious is the fact that the West Northfield anticline of Ammonoosuc Volcanics in the Bernardston area (Figure 1) is unrelated to the Bernardston nappe, but is a distal northwestward flap of this complex structure. This belt of Ammonoosuc was originally connected to the Ammonoosuc Volcanics peripheral to the west margin of the Monson Gneiss, and it is perhaps only an accident of erosion level that Monson Gneiss is not exposed in the center of the West Northfield anticline (see Robinson et al., 1991, Figure 27). The result of this complex process was that the axial surfaces of the fold nappes, as well as the Brennan Hill thrust, change across the axial surface of a tight syncline, from their normal upward-facing tectonic orientation on the Keene dome and the Vernon dome to an inverted orientation north of the main body of Monson Gneiss . This same syncline is displayed in the Bernardston area by the isolated patch of Rangeley Formation encircled by the thrust trace.

The third major phase of folding was the dome stage, characterized by tight north-south trending anticlines and synclines related in time to the formation of the gneiss domes, which appears to have been Acadian east of the Warwick dome. In the map pattern of the Mt. Grace area it is easy to see that these folds severely deform the axial

surface of the syncline produced during northward overfolding of the Monson Gneiss. In the Mt. Grace and Orange areas the dome-stage minor folds and mineral lineations consistently plunge south at angles from 20° up to 80°. Fabrics of conglomerates strongly indicate rock elongation parallel to the folds, suggesting they may have formed during an episode of orogen-parallel shear (see Peterson, this volume).

METAMORPHISM

The rocks observed on this field trip were subjected to Acadian regional metamorphism ranging in intensity from the garnet zone at Stop 1 to the sillimanite-staurolite-muscovite zone at Stops 2-8. Presently available evidence indicates that the peak metamorphic assemblages at all of the stops are Acadian and most formed synchronously with the Acadian dome stage about 360 Ma. None of the locations appears to lie inside the area of intense Pennsylvanian deformation and recrystallization at about 295 Ma that is documented by Robinson et al., elsewhere in this volume. Problems and details of the metamorphism near Stop 2 are discussed by Elbert, 1988, and Robinson et al., 1991; and characteristics of the metamorphism in the Mount Grace area are summarized by Hall (1970), Tracy et al.(1976); Schumacher and Robinson (1986); and Robinson et al., (1986).

ACKNOWLEDGEMENTS

The research related to this paper benefitted from the help and suggestions of a wide variety of persons too numerous to mention fully. For the first author it began at the inspiration of J. B. Thompson, Jr. and under the strong guidance of Marland P. Billings, and in interaction with Newell Trask. The original search for conodonts at Stop 1 was inspired by Norman Hatch, who provided the connection with Anita Harris and Kurt Denkler. The present stratigraphic interpretation of the rocks at Stop 5 was wisely foreseen by Norman Hatch and Gary Boone in 1976, though then rejected by the first author. Field research was supported by grants from the Earth Sciences Section of NSF, including most recently grants EAR-8410370, 8608762, 8804852 and 9105438(to Robinson). Preparation of the road log was assisted by Jon Bull.

ROAD LOG

Starting Point is Streeter's Barber Shop at 8:30 A.M., on Route 10 immediately west of the intersection with Interstate 91 (Exit 28, Route 10 South) in Bernardston, Massachusetts. This is approximately 23 miles and 30 minutes drive from Amherst. All the stops on this trip are on private property. Please respect both the outcrops and the rights of the landowners. Those following this guide in the future are responsible for securing permission before entering private property.

Mileage

0.0 Proceed west on Route 10 toward the intersection with Route 5.

0.1 Bernardston Auto Exchange-Streeter's General Store on left.

0.5 Stop sign, junction Route 5. Turn right and proceed north on Route 5.

0.9 Outcrop on left of conglomeratic Upper Triassic Sugarloaf Arkose at the northern extreme of the Deerfield basin. Continue northward on Route 5.

1.1 Pull off road onto wide grassy shoulder on right (east) side of road and park. Cross road (CAREFUL...Route 5 is heavily travelled at times). Walk up short dirt track, through gate into pasture. CLOSE GATE BEHIND YOU!!

STOP 1. BERNARDSTON FOSSIL LOCALITY (APPROXIMATELY 1 HOUR) Outcrops in the lowest part of the pasture (Figure 2) are part of the lower member of the Clough Quartzite. The rocks are gray weathering phyllites containing quartz-muscovite-chlorite-biotite-garnet-graphite-leucoxene. Typical of the lower- to middle-garnet zone in the region, these rocks are fine grained and virtually slates. Bedding is difficult to discern in this unit. The prominent schistosity is folded by a local open map-scale dome-stage fold (Figure 2). Outcrops approximately halfway up the pasture are metamorphosed conglomerates of the middle member of the Clough Quartzite. Here the conglomerates include a few schistose clasts as well as an impure, somewhat schistose matrix.

Continue to traverse towards the woods at the northwest corner of the lower pasture. Cross the barbed-wire fence carefully and continue northwards along the logging road. Abundant outcrops, especially on the west side of the logging road, are clean, quartz-pebble conglomerates with quartzite matrix. The conspicuous pebbles are uniformly vein quartz. This rock is typical of the Clough Quartzite as mapped by most workers and is representative of the middle member of the Clough Quartzite throughout the Bernardston-Hinsdale area.

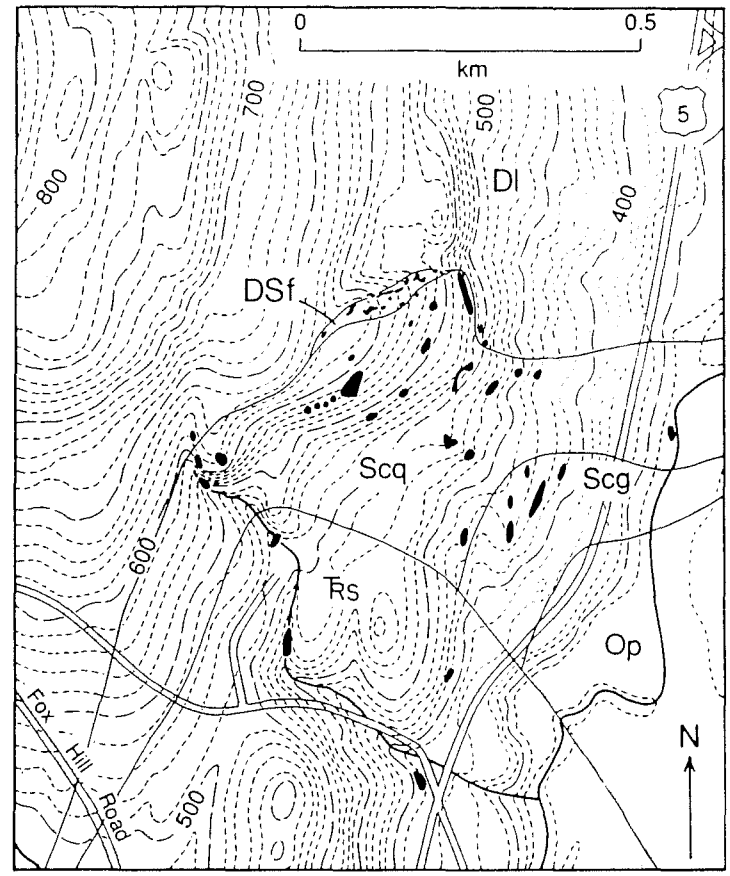


Figure 2. Detailed contact and outcrop map in the vicinity of the Bernardston fossil locality, Stop 1, from Elbert et al., 1988. Unit symbols: Op - Ordovician Partridge Formation; Scg - gray garnet-mica phyllite member of Lower Silurian Clough Quartzite; Scq - Silurian Clough Quartzite including pebble conglomerate; DSf - Lower Devonian - Silurian Fitch Formation marble; Dl - Lower Devonian Littleton Formation gray garnet-mica phyllite; Trs - Upper Triassic Sugarloaf Arkose.

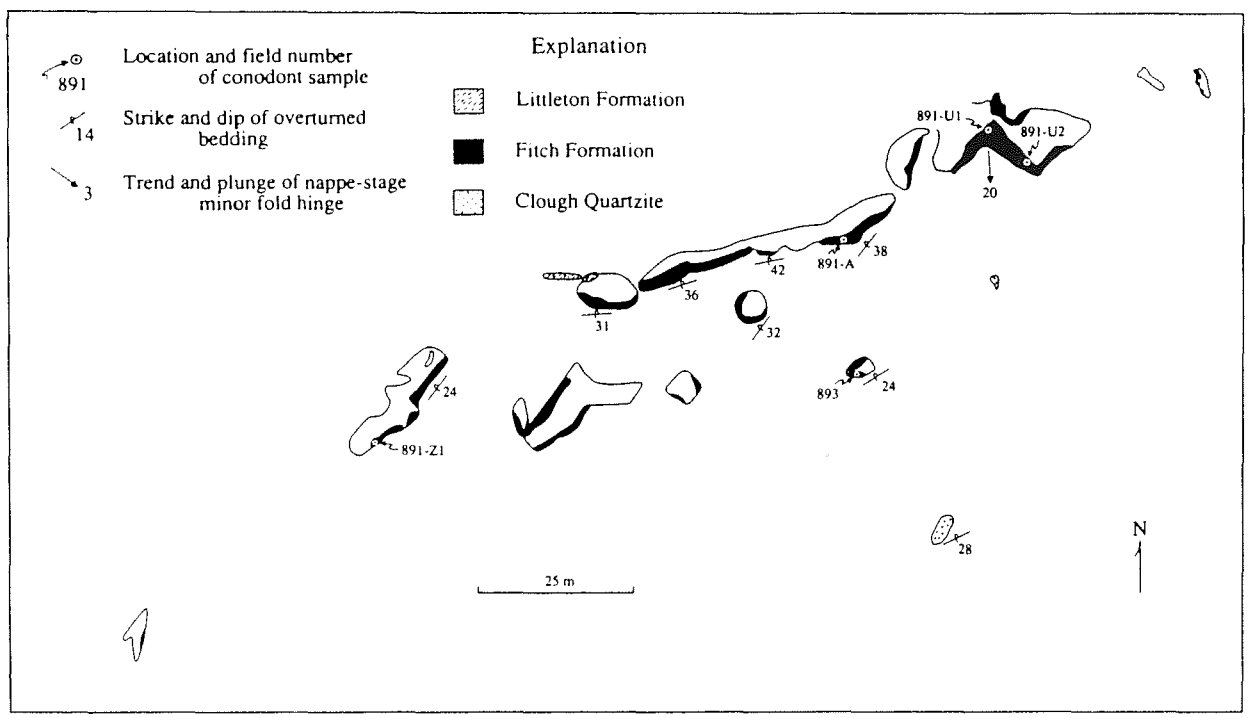


Figure 3. Tape and compass map from Elbert et al. (1988) showing conodont-sample localities and stratigraphic units exposed (outcrops patterned) in pits (outlined) in the Fitch Formation marble at Bernardston (see Figure 2). Sample location 893 lies in pit a few meters southwest of right-angle bend in logging road. Other marble-bearing pits are north of the road, mostly under large hemlocks. Structure symbols apply to closest outcrops.

Proceed northward along the logging road to where the road takes a nearly right-angle bend to the west. Marbles of the Fitch Formation are exposed in several small pits dug during mining of a small magnetite bed in the eighteenth and nineteenth centuries (Figure 3). Most of these pits are under the hemlocks to the north of the road. Due to the limited amount of conodont-bearing material and the unfortunately rapid disappearance of these important outcrops please **DO NOT HAMMER OUTCROPS** at this locality. There is abundant loose material for collection. The Fitch here is chiefly white to light-gray calcite marble containing scattered crinoid ossicles and minor (typically 1-3 percent modally) quartz, epidote, and pyrite. The prominent bed of magnetite-chlorite-quartz-garnet granulite is present in several pits as much as 0.8 m thick. The fossiliferous section of the stratigraphically highest Clough is interpreted to be a metamorphosed marine deposit which contained stenohaline, marine invertebrates (Boucot and others, 1958). Although this stratigraphically highest part of the Clough has been completely removed from the outcrops, its position was clearly documented by geologists in the nineteenth century. The map pattern of the Fitch, both locally and regionally, suggests that the Fitch unconformably overlies the Clough. The Fitch is unconformably(?) overlain by gray schist and quartzite of the Littleton Formation. One small, ground outcrop of Littleton Formation is shown on the map in figure 4. The outcrop is unimpressive but important as the best constraint on the stratigraphic top of the Fitch at this location.

Within the marbles lensoid bedforms, relict crossbedding(?), abrupt variations in grain size, and a wide range in the size of bioclasts indicate deposition in a near-shore, shallow-water, variable, but generally high-energy, marine environment. Megafossils include crinoid ossicles (as large as 3 cm in diameter), columnals (as much as 5 cm long), and large recrystallized coralline fragments. Taken together, these data indicate that the Fitch near Bernardston is chiefly a carbonate shoestring sand that formed as a channel filling, bar, or beach. The overall transition from Clough to Fitch to Littleton represents a grossly deepening upward sequence.

The southerly dip of the units at this location, coupled with the clear-cut stratigraphic distinctions and paleontological age constraints, presents straightforward evidence that the rocks in the region have been structurally inverted. This inverted belt of rocks continues for tens of kilometers along strike to the north and makes up the inverted limb of the Bernardston nappe. The Bernardston nappe is the structurally lowest of the nappe-stage Acadian folds at this latitude. Although outcrop-scale structural features unequivocally associated with the nappe-stage are rare, a small, south plunging nappe-stage fold in bedding can be seen on the north facing outcrop near the sample locality labeled 891-U1 in Figure 3.

Please retrace your path back to your vehicles being careful to leave all fences and gates as you found them.

Continue northward on Route 5.

1.7 Turn right onto Burke Flat Road.
1.9 Bridge over Interstate-91.
2.0 Intersection with Bald Mountain Road, bear right.
3.2 Junction Route 10, turn left (Route 10 north).
4.7 Historic marker - Lt. Ebenezer Sheldon's Fort.
5.8 Entrance on right to Mount Hermon Campus, Northfield Mount Hermon School.
6.1 Outcrops on Route 10 are rusty weathering pelitic schists and amphibolites of the Partridge Formation. Those on the northern side of the road include a Mesozoic shear zone.)
6.8 Bennett Meadow Bridge over Connecticut River. Skyline view of Notch Mountain, in Ammonoosuc Volcanics in the mantle of the Warwick dome..
7.3 Cross railroad bridge. Outcrop of Lower Jurassic, conglomerate member of the Turners Falls Sandstone in the Northfield Mesozoic basin is located down the embankment on the northeast side of the bridge.
7.4 Flashing light and stop sign at junction of Route 63, turn left onto routes 10 and 63 north.
8.1 Flashing light in center of Northfield, IGA supermarket on left.
9.3 Northfield-Mt. Hermon School, Northfield Campus on right.
9.8 Turn left on Route 63 north toward Hinsdale, N.H.
10.4 Entering Winchester, N.H. Begin Dartmouth College Road: *"Over this Route Eleazer Wheelock passed to found Dartmouth College, 1770."*
11.7 Old R.R. abutments.
12.7 Second old R. R. abutments.
12.8 Outcrop on right of rusty-weathering Partridge Formation staurolite-garnet-biotite schist with meter-thick felsic-volcanic bed.
13.8 Pass outcrops of Pauchaug Gneiss in core of Vernon dome on right.
14.4 Turn right onto Tower Hill Road (the turn is somewhat obscured when approached from the south).

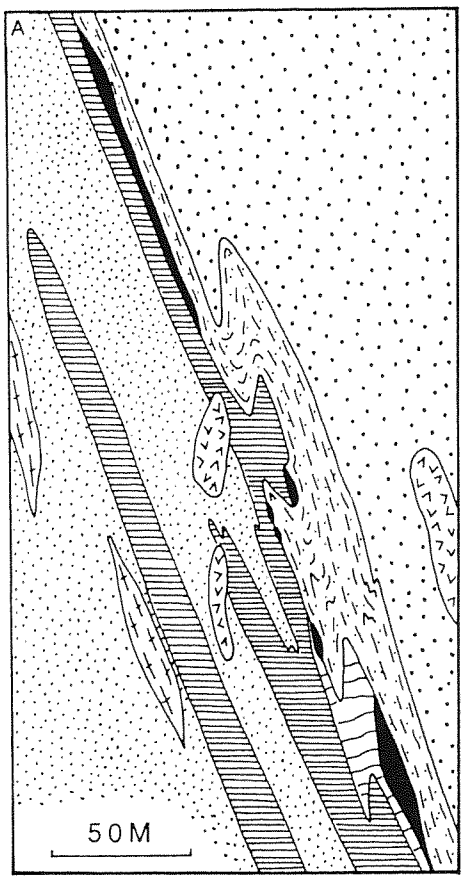


Figure 4. Detailed tape and compass map of part of Biscuit Hill, Hinsdale, New Hampshire, from Robinson et al., 1988. The multiply folded strata of the Monadnock sequence dip steeply northeast, and hence face downward toward the Brennan Hill thrust which lies beyond the edge of the figure to the lower left. The map-scale folds formed during backfold and dome stages.

14.6 Former site of railroad bridge.
14.8 Pass Depot Street on left, outcrops behind house northwest of intersection are parautochthonous Ammonoosuc Volcanics on the Vernon dome, small outcrop at northeast corner of intersection is Clough Quartzite on the dome. Continue up Tower Hill Road.
14.9 Pass Old Northfield Road on right, continue up Tower Hill Road.
14.95 Cross trace of Clough Quartzite on overturned limb of Bernardston nappe, continue up Tower Hill Road.
15.05 Outcrops of Partridge Formation in the core of the Bernardston nappe in brook to left of road.
15.6 Turn vehicles around in dirt road on left, drive back down Tower Hill Road.
15.7 Pull off road on right and park. Walk northwestward (follow flagging) to the western summit of Biscuit Hill.

STOP 2. MONADNOCK-SEQUENCE STRATA AT BISCUIT HILL (APPROXIMATELY 2 HOURS) Biscuit Hill presents an exceptional opportunity to see the rock types of the Monadnock-western Maine sequence in order in an area that has been mapped at 1/1000 (Figure 4). This trip's traverse has been designed to proceed from the top to the bottom of the hill, thereby working ***structurally downwards*** and ***stratigraphically upwards***. **Please be careful, the hillside is deceptively steep and there is a lot of loose rock to slip on or to knock onto other participants. Please hammer with forethought. Important folds, graded beds, and rare rock types are sometimes hard to see and should not be destroyed. If you are not absolutely sure you're sampling benignly ask for help!** There is quite a bit of collectable material already loose. A description of the important stratigraphic, structural, and petrologic aspects of this stop is given in the body of this chapter or by Elbert (1988). The following is meant to be a specific reminder of what to look for on the hill.

Most of the details of the Biscuit Hill stratigraphic sequence can be seen on this hillside traverse. They include the following: Interbedded gray-weathering schist with schist-matrix conglomerates and rare calc-silicate pods of the Rangeley Formation. A few thin interbedded quartzites mark the stratigraphic top of the Rangeley Formation and the gradation into the cyclically interbedded quartzites and gray schists of the Perry Mountain Formation. Although the contact between the Rangeley and Perry Mountain formations is gradational, exposure is sufficient in this region to map it consistently using the criteria that all conglomerate, calc-silicate pods, and epidote-bearing quartzites are mapped within the Rangeley Formation, while the base of the Perry Mountain Formation is marked by the first, continuous quartzite bed stratigraphically above schists and quartzites containing these distinctive Rangeley rock types. Graded beds near the Perry Mountain-Rangeley contact confirm the topping direction. The Perry Mountain contains biotite-rich, massive gray schist with beds and boudins of fine-grained coticule (garnet granulite) and magnetite-grunerite-garnet-apatite-quartz-graphite iron formation. Stratigraphically higher rocks at Biscuit Hill are well bedded, rusty-weathering, sulfidic, graphitic calc-silicate granulites and interbedded sulfidic Mg-biotite schists of the Francestown Formation. These are stratigraphically overlain by well bedded, clean actinolite-garnet-calcite calc-silicate gneisses, garnet granulites and interbedded purplish biotite granulites of the Warner Formation. A pair of large sills of metamorphosed gabbroic rock occur on the side of the hill. Although these are substantially altered and largely composed of biotite rather than hornblende (one also contains coarse garnet porphyroblasts), I have mapped them as sills of Mount Hermon Hornblende Gabbro, the only gabbroic intrusive known in the vicinity.

Backfold- and dome-stage deformations are recorded in outcrop-scale folds, rock fabrics, and map pattern. These two phases have virtually parallel axial surfaces as well as fold axes that plunge moderately towards the southeast, parallel to the mineral and pebble lineations, and are difficult to separate. However, some minor folds can be shown to have axial planes parallel to the prominent foliation and belong to the backfold stage, while others fold that same foliation and belong to the dome stage. Quantitative structural analysis is impeded by the presence of abundant magnetite-bearing iron formation in both outcrop and float. These two phases were preceded by early isoclinal folding and then thrusting which first produced the overall inversion of the Biscuit Hill sequence of rocks and then juxtaposition of the uppermost Silurian Warner Formation directly against the Late Ordovician Partridge Formation in the core of the Bernardston nappe just south of the base of the hill. Field trip participants may want to argue about what we feel to be the remnants of a fabric which may have been associated with the early isoclinal folding and can be seen in the beautifully graded beds in the Perry Mountain Formation.

Walk out (south) along logging road at base of Biscuit Hill, then ~0.25 miles up (east) Tower Hill Road. Follow flagging southward a few hundred feet into the woods to an exposure of the Brennan Hill thrust. The Brennan Hill thrust itself is not exposed on Biscuit Hill. It is exposed here, a short distance to the south, placing Upper Silurian Warner Formation against Late Ordovician Partridge Formation. One of the most interesting features of this fault is that it must be identified by stratigraphic mapping, because it is not marked by obvious structural or metamorphic differences.

Return to Tower Hill Road and walk up (east) to vehicles. Drive down Tower Hill Road.

16.0 Pass logging-road exit from Biscuit Hill.

16.3 Turn right onto Depot Street.

16.7 Across defunct railroad crossing, continue down Depot Street.

16.8 Cross bridge over Ashuelot River.

17.0 Junction routes 119 and 63, turn right on Route 119 toward Winchester.

18.6 Small cut in Kinsman Granite on left.

20.5 Flashing light and covered bridge (do not cross) at Village of Ashuelot.

22.3 Bridge over Ashuelot River.

22.4 Junction with lights of Routes 119, 10 and 78. Go straight at intersection on Route 78.

22.7 Kulick's Mall on left side of Route 78. Proceed east, then south, on Route 78.

24.0 Broad view of topographic basin eroded in granitoid core of north lobe of Warwick dome. Folded cover sequence at north end of dome is exposed at Meetinghouse Hill in Winchester village and shows that north end of dome plunges due east. Mt. Grace (1617') looms to the south, held up by the Lower Member of the Ammonoosuc Volcanics in a major northeast-plunging cross fold separating the dome into north and south lobes. Reconstructions across the Connecticut Valley border fault suggest that the Vernon dome is "rooted" in the northern part of the Warwick dome.

25.8 Massachusetts State Line.

29.1.1 Bear left (east) off Route 78 at Warwick Public Library in center of Warwick and proceed east on main road (Figure 5). Northeast and north of the church are outcrops of the mafic Lower Member of the Ammonoosuc Volcanics cut by folded pegmatite dikes. The dominant rock type is hornblende amphibolite, but there are also epidote-rich layers, and plagioclase gneisses with cummingtonite, garnet, biotite, and secondary chlorite. These outcrops are typical of the Lower Member of the Ammonoosuc Volcanics near Mt. Grace where the whole formation is 4000 feet thick. In these outcrops minor folds and lineations plunge steeply northeast and these outcrops lie west of the lineation "swirl"described by Robinson (1963), whereas the outcrops at STOP 1 lie east of it.

29.6 Park on right near small cut and outcrops of quartzite a few feet into woods. Small new road cut in gray schist is visible ahead on both sides.

STOP 3A. CLOUGH QUARTZITE, EAST LIMB OF WARWICK DOME (15 MINUTES) The purpose of STOP 3 in total (Figure 5) is to demonstrate the Clough Quartzite in autochthonous position on the east limb of the Warwick dome, the Littleton Formation overlying it, the Clough Quartzite repeated in inverted position above the Littleton Formation in the inverted limb of the Bernardston nappe, and the Partridge Formation structurally above the Clough in the anticlinal core of the nappe. This doubled sequence produced by the nappe-stage recumbent syncline beneath the Bernardston fold nappe is isoclinally refolded by the dome-stage Rum Brook syncline. A cross section, in which movement on the Mesozoic Connecticut Valley border fault is restored

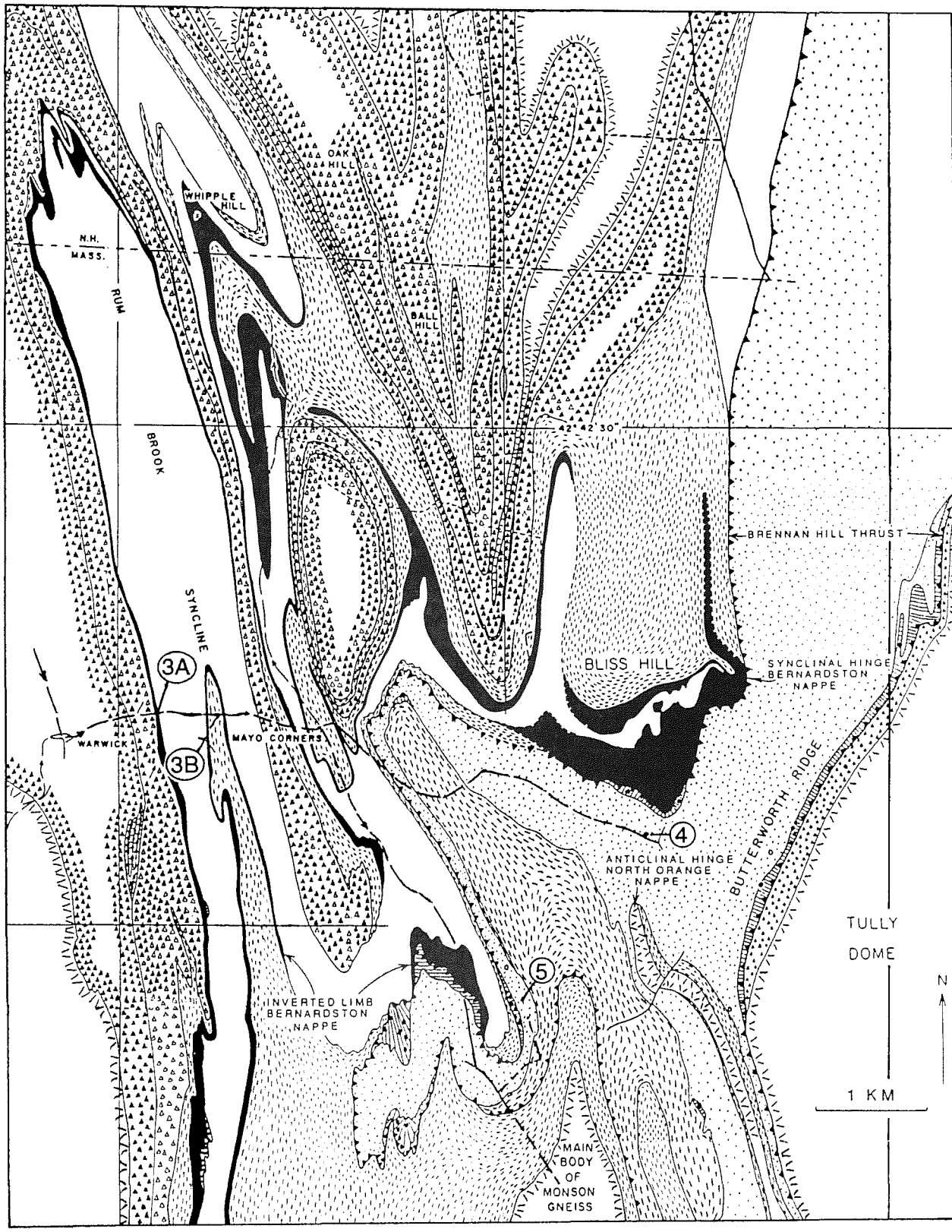


Figure 5. Geologic map of the central part of the Mt. Grace quadrangle and adjacent part of the Royalston quadrangle showing the locations of Stops 3A, 3B, 4, and 5. For key to symbols see explanation for Figure 1.

(Robinson et al., 1991, Figure 27), suggests that the inverted Clough described here connects directly in space with the lower limb of the nappe exposed at Bernardston.

Proceed east past small new road cut and larger road cut in Littleton Formation on both sides. These are good looking sillimanite-muscovite-staurolite schists that can be seen to greater advantage in natural exposures at STOP 3B. Recently R. D. Tucker (Robinson et al., this volume) has obtained a concordant U-Pb age on monazite from this outcrop of 359 Ma, indicating that the last significant metamorphic recrystallization in these low-sillimanite-zone rocks was Acadian.

30.5 Mayo Corners, go straight through.

32.6 Orange Town Line and outcrop to left which is Stop 4. Just before Town Line, turn left (northwest) into entrance for Sheomet Lake. Travel about 0.2 mile and park in lot southeast of outlet dam or on east shore of lake

LUNCH STOP (40 MINUTES) If weather is suitable, lunch will be held east of lake outlet where there are outcrops and blocks of Rangeley Formation sillimanite schist cut by south-dipping tourmaline veins. East of the lake is Bliss Hill, underlain by complexly folded Clough Quartzite in a synclinal hinge of the Bernardston nappe. The Brennan Hill thrust lies along the contact between the inverted limb of Clough Quartzite and Rangeley Formation. The contact is locally decorated by lenses of Perry Mountain garnet-cummingtonite rock (Huntington, 1975). On a small island in the lake is an outcrop of biotite marble that may correlate with Fitch Formation. After lunch, drive about 0.2 mile back to Orange Town Line and Stop 4 to resume road log..

STOP 4. RANGELEY FORMATION SILLIMANITE-STAUROLITE SCHIST (10 MINUTES). Walk east from Town Line on north side of road to large road exposure. This outcrop has been pronounced by Ben Harte of Edinburgh University as the most beautiful sillimanite-staurolite schist he has ever seen. Unfortunately, no electron probe analyses have yet been completed from this location although there is fairly abundant unpublished data from the surrounding region. The massive fibrolite sillimanite veins (described by B.K. Emerson, 1895, as "bucholzite") are here studded with euhedral staurolites.

The rock unit at this location was originally assigned by Hadley (1949) to the Lower Devonian Littleton Formation, then by Robinson (1963) to his Gray Member of the Middle Ordovician Partridge Formation, then assigned back to the Littleton Formation by Robinson (1967), and now, on the basis of new work in the Monadnock quadrangle (P.J. Thompson, 1985), is assigned to the Lower Silurian Rangeley Formation! The steeply south-plunging open folds with parallel mineral lineation are typical of dome-stage folds near the south-plunging end of the Keene gneiss dome. One or two delicate graded beds suggest tops north indicating the strata are here upside down. This facing sense happens to be consistent with the appearance of lenses of iron formation of the Perry Mountain to the north, apparently in fault contact with Clough Quartzite across the Brennan Hill thrust.

This outcrop lies about one mile west of the staurolite-out isograd, although it has not yet been proved that the loss of staurolite is due to a prograde reaction or a change of bulk composition. Many schists in this vicinity contain only a muscovite-sillimanite-biotite-garnet assemblage. Probe analyses of staurolite assemblages in gray schists from this general vicinity give the following information:

Staurolite	Fe/(Fe+Mg) = 0.83-0.84,	ZnO 0.27-0.54 wt %;	
Biotite	Fe/(Fe+Mg) = 0.53-0.59,	Ti/11 Ox. = 0.09-0.10;	
Muscovite	K/(K+Na) = 0.74-0.78,	Ti/11 Ox. = 0.01-0.02,	
Garnet Rims:	Alm 78-82,	Pyr 10-13,	Spess 3-6, Gros 2-5.

A few garnets show growth zoning with decreasing spessartine from 13 in cores to 3 in rims, all at nearly constant pyrope content. Opaque minerals are uniformly ilmenite and graphite. Garnet-biotite geothermometry suggests, with considerable uncertainty, temperatures of 550-630 °C, and garnet compositions give estimates of minimum pressure of 5.4-6.5 kbar using the calibration of A.B. Thompson (1976b) and Tracy et al. (1976), respectively.

Proceed west on Athol Road, retracing previous route to Mayo Corner and beyond.

35.2 Turn right (south) on Gale Road proceed to partial road block at about 0.15 miles. Do U-turn and park on right side of Gale Road facing north. .

STOP 3B. STRATA IN INVERTED LIMB OF BERNARDSTON NAPPE. At this stop will be seen Littleton Formation on the west limb of the Rum Brook syncline, Partridge Formation of the Bernardston nappe in the center of the Rum Brook syncline, inverted Clough Quartzite on the east limb of the Rum Brook syncline, and finally Littleton Formation on the east limb of the Rum Brook syncline (Figure 5).

Walk south and west past road block in abandoned paved road to outcrops on north side of U-shaped bend. These are superb sillimanite-muscovite-staurolite schists of the Littleton typical of the lowest grade part of the sillimanite zone where quartz-sillimanite knots are just beginning to develop. Foliation dips 50-60° east. Dome-stage mineral lineation plunges steeply southeast in this location just on the east side of the dome-stage lineation swirl described above.

Return to position of parked vehicles, crossing unexposed inverted Clough Quartzite on west limb of Rum Brook syncline. Proceed east into woods just south of new house. Here there are several exposures of rusty-weathering sillimanite schist and amphibolite of the Partridge Formation in the center of the Rum Brook syncline. Proceed east through woods to steep east-facing ledges of Clough Quartzite, including conglomerate, overlooking swamp. Foliation dips about 60°. These are in the inverted limb of the nappe and on the east limb of the Rum Brook syncline. Follow Clough ledges north along strike and continue north to main E-W road. Just across road to northeast is a small outcrop of Littleton Formation on the east limb of the Rum Brook syncline. Walk short distance west to Gale Road, then south to vehicles. Drive north to main E-W road.

35.5 Turn right (east) off Gale Road onto E-W road and recross allochthonous Clough. Swamp ahead is underlain by Littleton, large outcrops (not visible) north of road.
35.7 Littleton-Ammonoosuc contact on east limb of Rum Brook syncline.
35.8 Small cut in sulfidic Partridge schist on left. This lies on east limb of dome-stage anticline of Ammonoosuc Volcanics. Between here and Mayo Corners is another narrow syncline of Littleton Formation with Clough on both sides, here separating autochthonous sequences peripheral to the Warwick dome to the west and the Keene dome to the northeast.
40.0 Mayo Corners yet a third time. Turn sharp right (south) on Hastings Heights (hastingsites?) Road.
40.8 Outcrops on both sides at sharp bend in road. West side of road is formed of Clough Quartzite that is autochthonous on a tight anticline connected to the Warwick dome. Littleton Formation is exposed in a large knob east of road. On the Littleton outcrop, projecting quartz-sillimanite knots are elongated parallel to steeply plunging dome-stage folds in foliation. The rock consists of quartz 68%, biotite 15%, sillimanite 12%, muscovite 5%, minor garnet, ilmenite, and graphite, and a trace of zircon. Many Littleton schists in this vicinity contain staurolite as an additional phase. This schist is typical of the Littleton that lies in the syncline beneath the Bernardston nappe.
41.8 Small outcrops of Clough Quartzite on left. This is part of a thin remanent at this position of the inverted limb of the Bernardston nappe that is cut off immediately above by the Brennan Hill thrust.
41.9 Warwick-Orange Town Line.
42.2 Junction 985' with Gale Road. Bear left.
42.3 Sharp left turn off pavement onto Town Farm Road.
42.6 Blue house on left and private road toward Johnsonian Pond. Depending on number of vehicles, walking tour for STOP 5 will begin here or if number of vehicles is small, it may be possible to drive 0.2 mile in. Route beyond this point is described in walking log (Figure 6).

STOP 5 INVOLUTED ROOT ZONE OF BERNARDSTON NAPPE AND MONADNOCK SEQUENCE ROCKS IN BRENNAN HILL THRUST SHEET (2 HOURS). The purpose of this stop is to show some of the stratigraphic and structural features that led to the new interpretation of the Mount Grace area. Most of the detailed mapping was completed in 1966 and 1967, but stratigraphic reinterpretation is based on new data and interpretations in the Mt. Monadnock area by P. J. Thompson (1985) and the Bernardston-Hinsdale area by D. C. Elbert (1984, 1986, 1987, 1988). The stop is designed as a traverse in a tectonically upward direction to see a series of features in sequence, all on the nearly vertical west limb of the dome-stage Williams Pond syncline. All lineations and minor fold axes, except at one outcrop, plunge 40-60° south-southeast parallel to the axis of the Williams Pond syncline. The one exception is the group of minor folds contained within a boudin of iron formation at **station f** that plunge 35-40° north and appear to belong to an earlier stage of folding. The features to be seen in sequence are:

1) Typical Littleton Formation from beneath the Bernardston nappe.
2) Sulfidic schist of the Partridge Formation with minor amphibolite boudins in a belt 100 feet wide that constitutes what remains of the anticlinal core of the Bernardston nappe. This is in direct contact with Littleton to the west, indicating absence or shearing out of Clough Quartzite in this location on the nappe.
3) Boudins of quartz pebble conglomerate with pitted matrix along contact between Partridge and gray feldspathic schist of Rangeley Formation to east. This is the inferred location of the Brennan Hill thrust. Along this

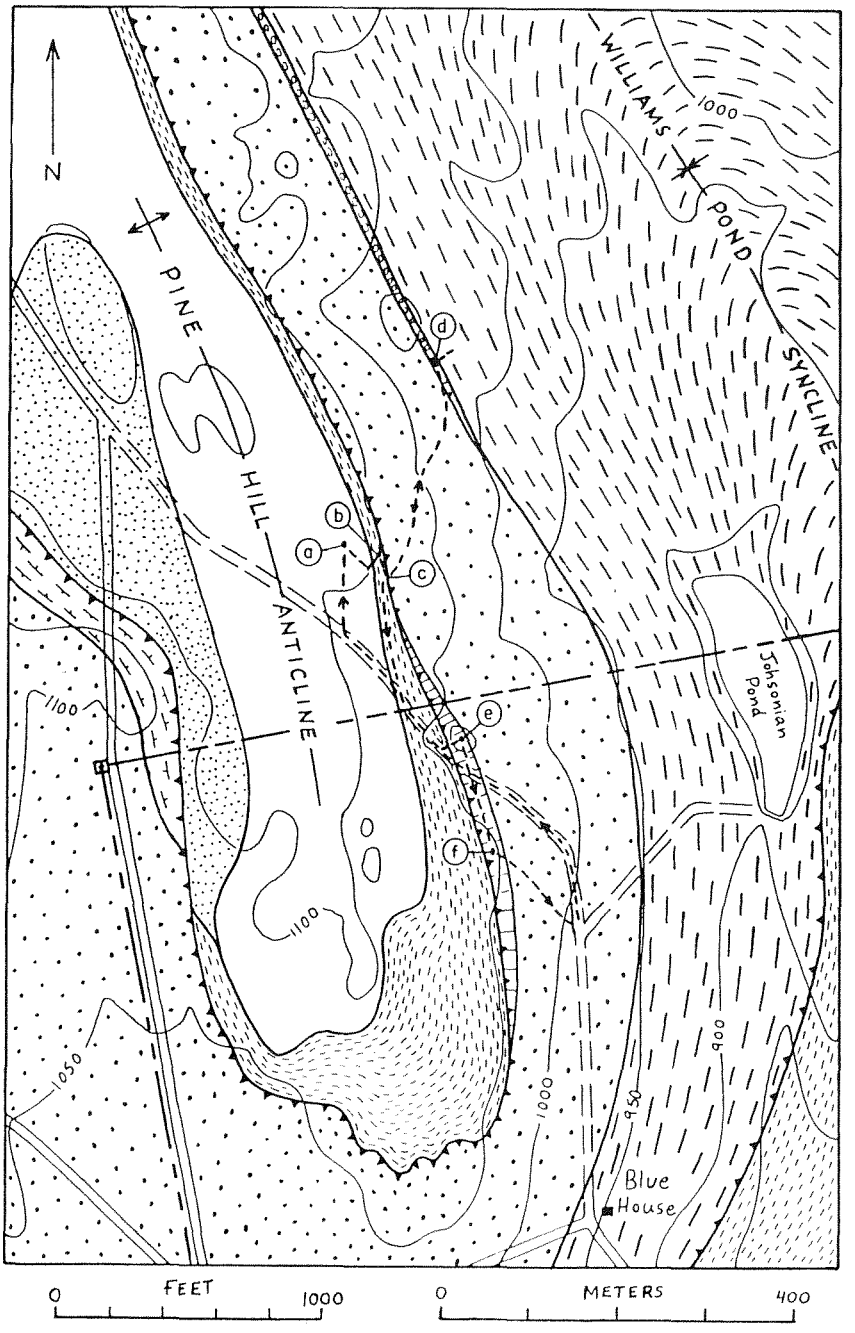


Figure 6. Detailed map of Stop 5 showing the Brennan Hill thrust above an attenuated root zone of the Bernardston nappe. Teeth on original upper plate of Brennan Hill thrust. These features have been involuted during backfolding and then refolded about the dome-stage Pine Hill anticline and Williams Pond syncline. Map units may be identified by comparison with Figure 5.

contact to the south, large boudins of iron formation assigned to the Perry Mountain formation will be seen, indicating the stratigraphic sequence above the thrust is inverted.

4) A belt of gray feldspathic and sillimanitic schist and granulite assigned to the Rangeley Formation. This belt occupies an isoclinal syncline believed to have formed in the backfold stage during northward overfolding of the main body of Monson Gneiss. About in the middle of this belt is a superb outcrop of granule-matrix conglomerate typical of the Rangeley.
5) A narrow belt of feldspathic granulite-matrix conglomerate assigned to the lower member of the Rangeley Formation. This is in contact to the east with the Augen Gneiss Member of the Partridge Formation along a contact tentatively believed to be an unconformity.

From the blue house (Figure 6) walk or drive north to fork in wood road. Where more-travelled fork bears right bear to left (limited parking here). Continue on foot on left fork past old stone post marking town line between Orange and Warwick. Continue on wood road until evergreen trees end and continuous grass appears in middle of road. Turn right (north) across stone wall and walk about 200 feet north to large ridge-top outcrops.

Station a Gray homogeneous sillimanite-rich schist of the Littleton Formation. Quartz-sillimanite knots are evenly distributed and give a "braille" effect to the outcrops. From these outcrops move diagonally down hill about S35E to east-facing outcrops overlooking small valley.

Station b This is a series of east-facing outcrops on west side of small valley to be traversed diagonally across strike in a direction N40E. Southernmost outcrop shows contact of gray Littleton schist to west and sulfidic

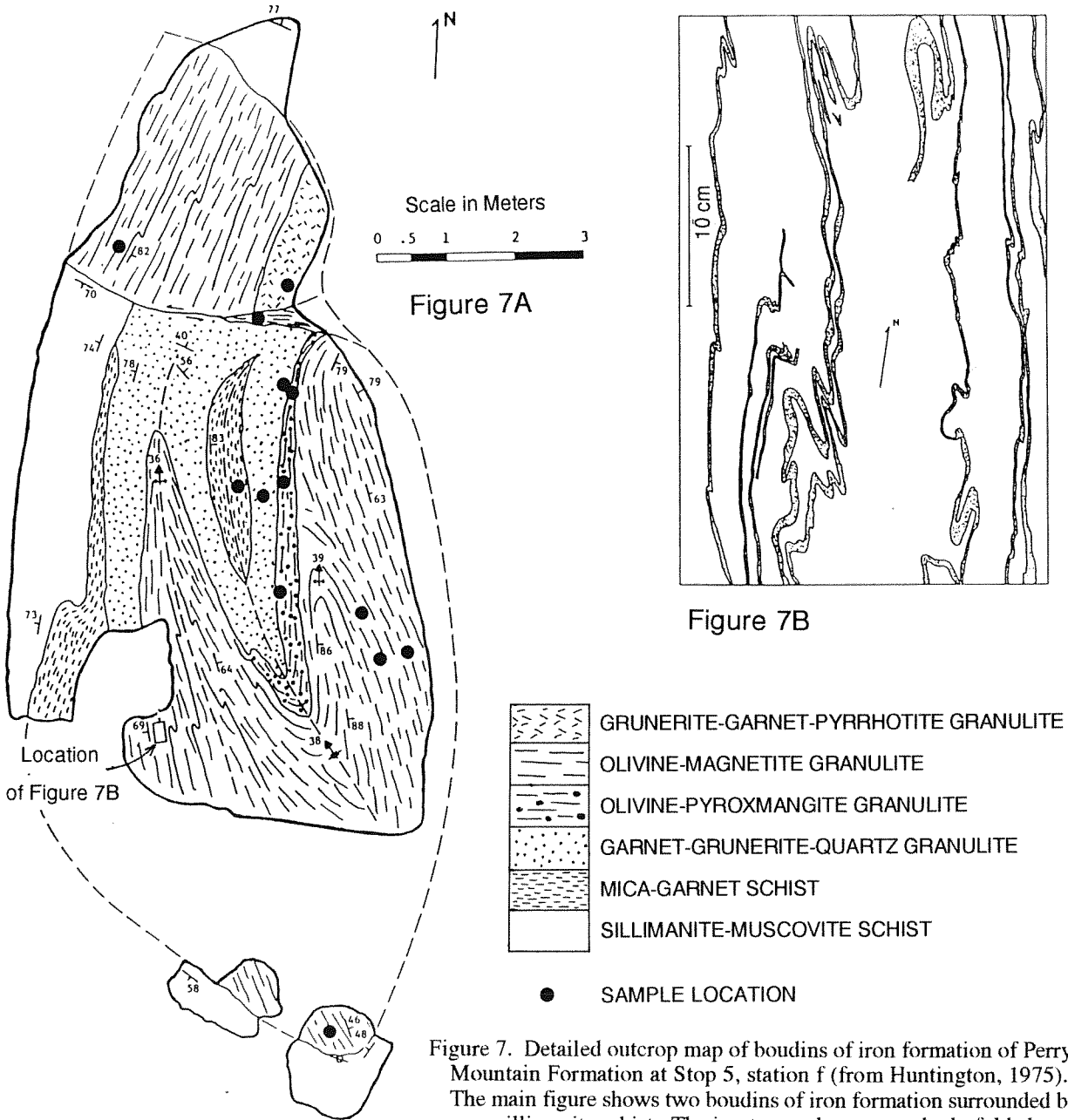


Figure 7. Detailed outcrop map of boudins of iron formation of Perry Mountain Formation at Stop 5, station f (from Huntington, 1975). The main figure shows two boudins of iron formation surrounded by gray sillimanite schist. The inset map shows complexly folded apatite beds in a small area.

Partridge schist of the Bernardston root zone to the east. At this location the root zone is 105 feet wide. The next several outcrops show sulfidic Partridge schist with small boudins of biotite amphibolite. About 200 feet to the northeast where the outcrop is steepest, sulfidic schist is in contact with a 1- to 2-foot lens of pebble conglomerate beyond which gray schist and granulite of the Rangeley Formation is exposed. A 10-inch pod of biotite amphibolite occurs in sulfidic schist about ten feet west of the conglomerate lens. The east edge of the sulfidic schist is the postulated location of the Brennan Hill thrust. From base of highest outcrop walk south about 50 feet across shallow valley to north-facing outcrop.

Station c A thin layer of pitted conglomerate is in sharp contact with sulfidic schist to west and in contact with gray schist and feldspathic granulite to the east. This was originally interpreted as Clough Quartzite between Littleton and Partridge on the upper limb of the Bernardston nappe. However, the conglomerate and the gray schist

and granulite are much more like rocks in the Rangeley Formation as defined in the Monadnock area. Traverse east across north slope and then northeast and downhill through a series of large outcrops of gray schist and feldspathic granulite of the Rangeley Formation, heading for a small southeast-trending stream gully shown on topographic map. During this traverse we will cross a northeast-trending swale and on the northside of this there is a superb peeled outcrop of Rangeley granule-matrix conglomerate, about equidistant between stations c and d.

Station d The contact between gray Rangeley schist to the west, and rusty schist and augen gneiss of the Partridge to the east can be located just on the north side of the gully. From here for a distance of about 250 feet north there are small outcrops and numerous loose blocks of granulite-matrix conglomerate with a probable thickness of 20-40 feet. These resemble the granulite-matrix conglomerate describe in the the lower member of the Rangeley in the Monadnock area (Thompson, 1985) and here may be resting unconformably on the Partridge Formation. Walk east 100 to 200 feet to view typical exposures of the Augen Gneiss Member. Retrace walking route to **station c** and follow shallow valley south to wood road. Proceed southeast on wood road past Warwick-Orange Town Line marker.

Station e Top of obvious small knob about 100 feet northeast of wood road. This shows the northernmost of several boudins of iron formation assigned to the Perry Mountain Formation based on the section at Biscuit Hill . It lies exactly along strike from the contact at **station c**. Walk southward parallel to strike across wood road to small steep outcrop on south side. Here is exposed an unusual lens of conglomerate tentatively assigned to the Perry Mountain. It consists of dark pebbles in a quartz-rich matrix. The dark pebbles consist of garnet, grunerite, apatite, dark green ferric chlorite and a trace of magnetite. The matrix consists of quartz, apatite, grunerite, hornblende, and minor chlorite and garnet. The pebbles appear to be redeposited fragments from the iron formation. Elbert also found evidence of redeposited iron formation at Biscuit Hill. Continue south at same elevation across woods trail to high-standing knob held up by two large boudins of iron formation.

Station f This large outcrop was discovered by Robinson in 1966 and mapped in detail following appropriate cleaning by Huntington (1975). Figure 7, based on his Figure 2 shows the distribution of rock types in the outcrop, including gently plunging early folds and the steeply plunging boudin neck line separating the two parts of the outcrop. The inset shows details of delicately folded apatite-rich beds in the southern part of the outcrop. Except for the fact that the rocks are now assigned to the Perry Mountain Formation rather than the Littleton, there is little to add to Huntington's detailed mineralogical and petrological analysis. Yet to be accomplished is an analysis of the sedimentary environment and paleogeography at the time of deposition of these unusual rocks during the late early Silurian. The low area west of the knob contains small outcrops of sulfidic schist and amphibolite of the Partridge Formation in the Bernardston root zone.

From the knob walk east on woods trail over pavement outcrops of feldspathic granulite of Rangeley Formation. Trail connects with fork in road. From there return south (right) back to blue house on Town Farm Road.

Following Stop 4, return right (west) on Town Farm Road toward Athol Road.

42.8 T junction with Athol Road. Turn left (southeast) and descend into basin eroded in Monson Gneiss.

43.7 Bottom of hill. Turn right (south) on North Main Street. Bottom of valley coincides almost exactly with position of synclinal keel of the main body of Monson Gneiss.

43.9 Low outcrops of Monson Gneiss on right. On left is Williams Pond, for which the dome-stage structural syncline is named, here expressed as a south-plunging keel of inverted Monson Gneiss. Proceed south through basin eroded in Monson Gneiss. As one approaches northern outskirts of Orange there are fine views to the southwest showing the prominent topographic ridge of Partridge Formation and Ammonoosuc Volcanics that bounds the Monson Gneiss to the west (Figure 8).

47.9 Stop lights in center of Orange. Turn right (west) on Route 2A.

48.1 Large exposure of Partridge sulfidic schist on right.

49.2 Junction Routes 2A and 78. Park on grass on right at entrance to gravel pit.

STOP 6. PARTRIDGE, CLOUGH AND FITCH FORMATIONS ON INVERTED LIMB OF BERNARDSTON NAPPE (30 MINUTES) This contact region is mapped continuously to here from the area of Stop 3B. Here the Clough-Partridge contact is exposed within 1m (Figure 8). The Clough is about 180 feet thick, of which the lower half is almost perfectly exposed. The Fitch is about 800 feet thick, the thickest anywhere in the Orange area, though not completely exposed. The bedding is essentially vertical. This location is about one kilmometer east of the first appearance of fibrolitic sillimanite in pelitic schists, which occurs as fibres within muscovite.

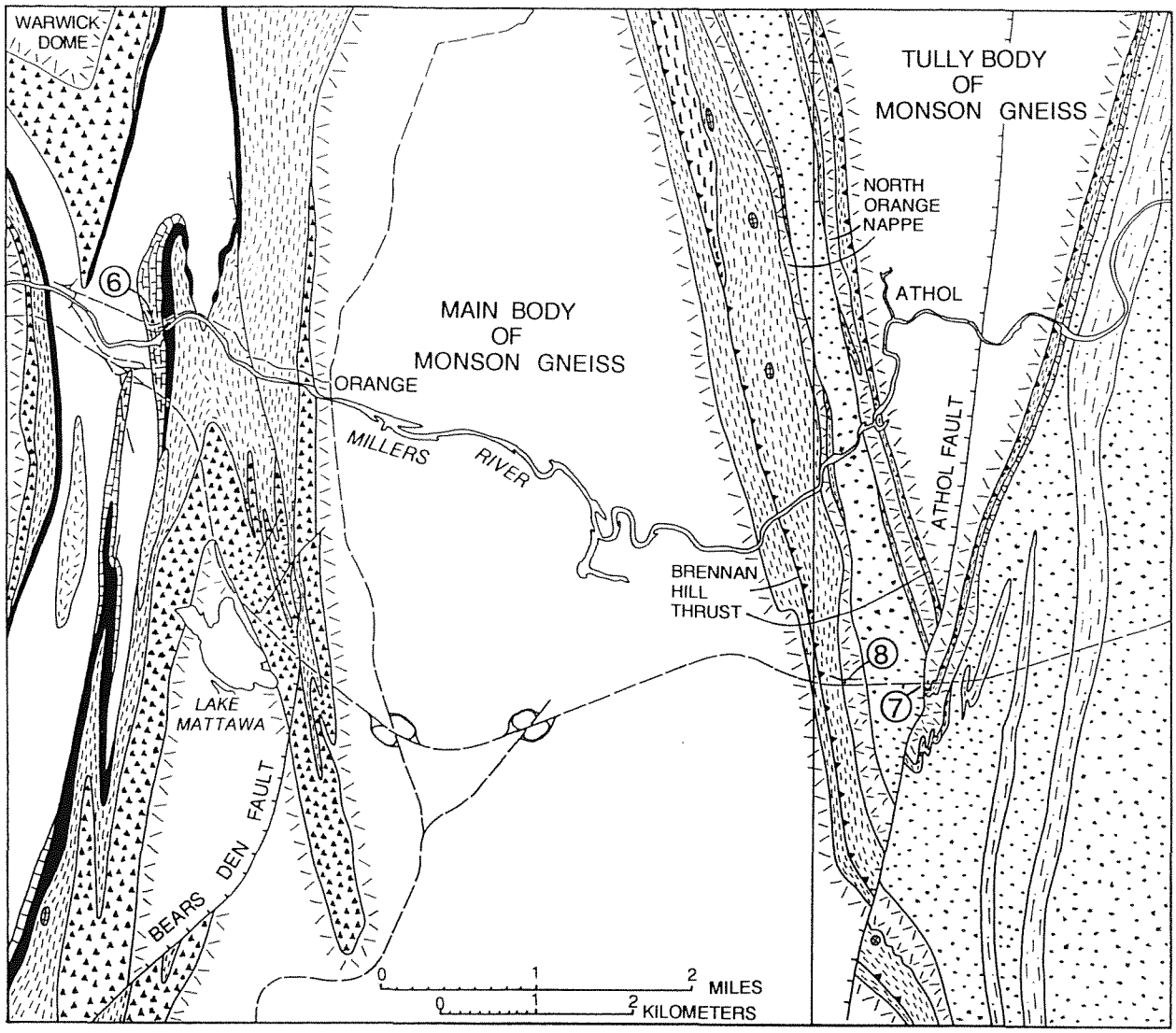


Figure 8. Generalized geologic map of the central part of the Orange quadrangle and the western edge of the Athol quadrangle showing locations of stops 6, 7 and 8. All of the obvious folds on the map plunge 10-30° south and formed during the dome stage. In the western part of the map these folds are in a previously inverted stratigraphic sequence, so that younger rocks appear in anticlines, older rocks in synclines. The main body of Monson Gneiss itself lies along the Williams Pond syncline. In the eastern part of the map, the North Orange nappe and the Brennan Hill thrust are repeated across the Temple Hill syncline, believed to have formed during the backfold stage, because there is no change in asymmetry of minor folds across this major fold. The south end of the Tully body of Monson Gneiss is a south-plunging anticline. For key to map symbols, see explanation for Figure 1.

Walk to knob in back of gravel pit where there are outcrops of Fitch, then northeast along low ridge and a short distance up and to right to Clough and Partridge outcrops on hill beyond. The Partridge is typical sulfidic muscovite-biotite-plagioclase-garnet schist, usually with minor staurolite and fibrolitic sillimanite. The Clough is a gray weathering schist near the base with more than 80% quartz plus muscovite, biotite, garnet, sillianite and minor staurolite and ilmenite. About 45 feet up from the base there are surprisingly feldspathic beds of quartz-plagioclase (An45)-mica-garnet granulite with minor tourmaline; at 60 feet there is more quartz-rich mica-garnet schist with up to 8% sillimanite forming a distinct dome-stage lineation that plunges gently south parallel to regional folds. The best Clough conglomerate is at the west end (top) of the good exposure about 80 feet above the base, where small pebbles can be recognized. Monazite separated from the quartz-sillimanite schist at this outcrop has given a concordant U-Pb age of 360 Ma, indicating the last significant metamorphic recyrstallization was Acadian (Robinson et al., this volume).

The Fitch Formation exposures, which are best just behind the gravel pit or in one slightly moved block within the gravel pit, consist of poorly bedded, fine-grained, gray weathering calc-silicate granulite containing various combinations of quartz, plagioclase, microcline, zoisite or clinozoisite, diopside, acrtinolite,calcite, scapolite, biotite, sphene and garnet. Interbedded with these are thin coarse layers with diopside, actinolite, zoisite or clinozoisite, garnet, or pits indicating the former presence of calcite. Minor rock types in nearby outcrop areas include sulfidic mica-silimanite-staurolite-garnet schist and actinolite-biotite schist. Marbles such as those exposed at Bernardston are unknown, but one outcrop of massive calcite marble with minor quartz has been located in the Fitch, about 5 miles south along strike from here.

Re-enter Route 2A and proceed west.

49.7 Sharp left turn (south) for Lake Mattawa through railroad underpass and across Millers River.

50.4 T intersection. Turn right and then bear left toward Route 2 East.

50.5 Another T junction. Left toward Route 2 East. First prograde fibrolite can be seen in road cut of Littleton about 0.2 mile west of here.

50.6 Left on entrance ramp for Route 2 East.

51.2 Large road cut in sulfidic schist of Partridge Formation low in sillimanite zone (see Figure 8).

51.7 Exposure in trees high on embankment to left. Coarse gedrite gneiss of Lower Member of Ammonoosuc.

52.1 View of Lake Mattawa on right. The low ground occupied by the lake and surroundings is mainly underlain by Monson Gneiss. This is in a gently-south-plunging dome-stage structural syncline in the stratigraphic sequence that had been totally inverted by the northward overturning of the main body of Monson Gneiss (Figure 8).

52.8 Partridge schist and amphibolite in core of Walnut Hill anticline, a dome-stage anticline in the same inverted sequence. Most of this ridge is held up by Ammonoosuc Volcanics with a narrow belt of Partridge along the anticlinal axis.

53.1 Road cut on right (bushes) shows contact between Monson Gneiss and Ammonoosuc Volcanics.

53.5 Overpass over Route 122.

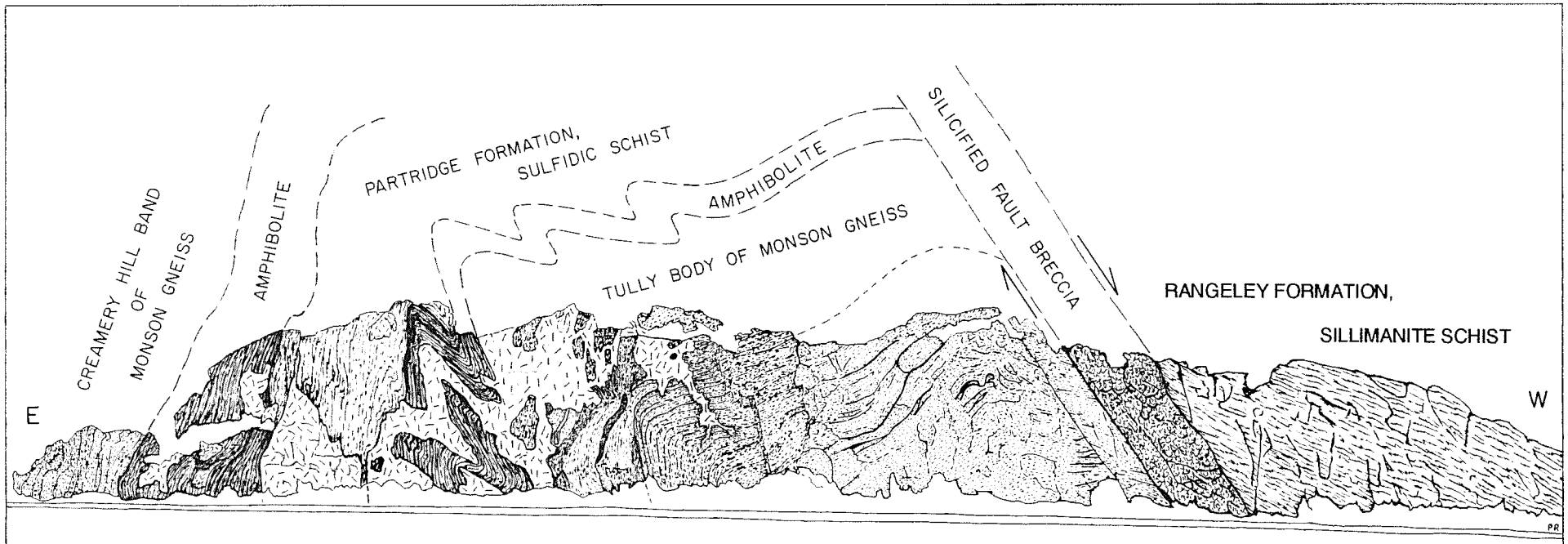
54.6 Junction with Route 202. Continue east on Route 2.

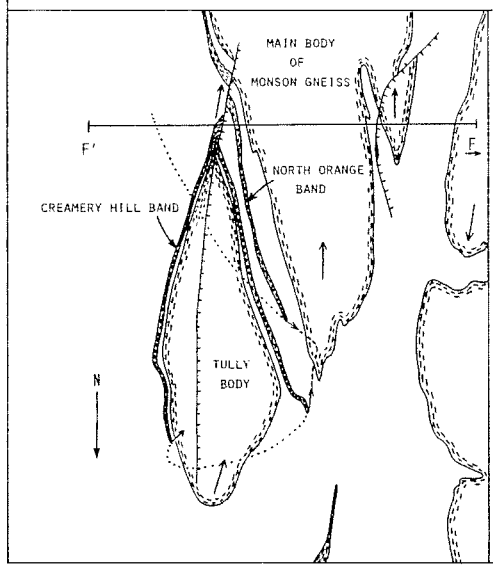
56.3 Bushed-over road cut on right. Gneisses in eastern part of main body of Monson Gneiss.

57.1 Athol road cut. Large, long trench cut, both sides. If traffic permits, particularly if you intend to proceed west after this stop, **bear left across highway to broad and firm grass strip on north side.** This area is much safer and better for viewing the geology than the narrow area on the south side. Since Figure 9 was drafted in 1979 (Robinson, 1979) the south side of the Athol cut has been blasted back about ten feet. This has changed some details, but has not altered any essential features.

STOP 7. SOUTH END OF TULLY BODY OF MONSON GNEISS, BRENNAN HILL THRUST, AND NORTH ORANGE FOLD NAPPE, CUT BY MESOZOIC NORMAL FAULT (30 MINUTES) The south end of the Tully body is a simple anticline overturned to the east and plunging gently south-southwest. Dome-stage normal asymmetric folds occur on both limbs. The exposures at Stop 7 are the southernmost ones of the gneiss in the core of the Tully body. At this point the anticline is cut just west of its crest by a west-dipping normal fault, the Athol fault, bringing gray schists of the Rangeley Formation on the west limb down about 1300 feet into contact with Monson Gneiss of the core. The actual fault zone is about ten feet wide and contains gouge zones, hematite-stained and cemented breccia, intense silicification, and vuggy quartz veins, all features characteristic of known Mesozoic faults. Secondary alteration extends into the metamorphic rocks on both sides for a considerable distance. Here the fault strikes N4W, but regionally it strikes N15E.

On the east limb of the anticline, east of the gneiss of the core, is a thin layer of rusty Partridge schist which has been traced for many miles. Well bedded amphibolites at both contacts probably belong to the Partridge, although it is tempting to consider that they might represent the Ammonoosuc Volcanics highly attenuated. Next east of the schist and amphibolite is a layer of Monson Gneiss (Creamery Hill band) that has been traced entirely around the southern part of the Tully body (see Figure 9 inset) and is interpreted as an early anticlinal fold nappe of gneiss that formed contemporaneously with the Bernardston nappe. A similar band of Monson Gneiss (North Orange band) east of the main body of Monson Gneiss is interpreted as the same recumbent anticline repeated by folding about the Temple Hill syncline that separates the two bodies. The North Orange band has been traced from North Orange at least as far south as the northern edge of the Palmer quadrangle in southern Massachusetts. Three anticlinal hinges for this early fold nappe, now named the North Orange nappe, can be seen in Figure 9 inset. This early fold nappe has been severely involuted, both by dome-stage folding and by the complex backfolding that





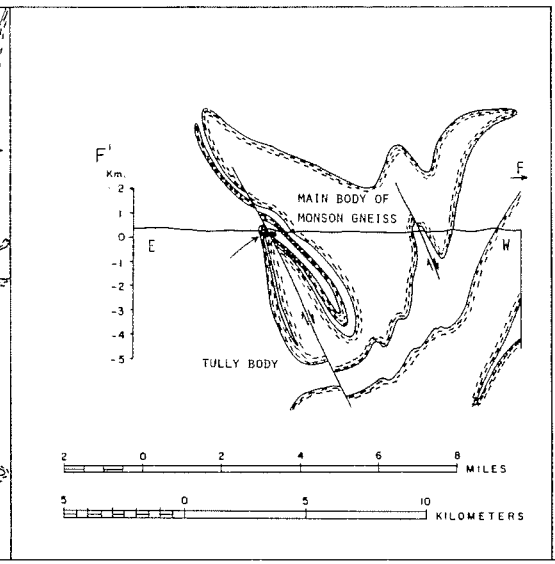


Figure 9. Sketch of south wall of rock cut at stop 8. Mesozoic Athol fault cutting crest of Tully body of Monson Gneiss. Approximate location of the cut is shown by nearly invisible stippled rectangle indicated by arrow in inset cross section. The major anticlinal or dome axis and satellitic folds are parallel to a strong Beta maximum (10% per 1% area) with trend N22E and plunge 18SW (Robinson, 1963). The Creamery Hill band of Monson Gneiss is interpreted (see insets) as an extremely attenuated basement nappe, the North Orange nappe, separated from the main and Tully bodies by an extremely attenuated isoclinal syncline and by the Brennan Hill thrust. The North Orange nappe is interpreted as a fold of the same generation as the Bernardston nappe but lying tectonically higher and more easterly (see Robinson et al., 1991, Figure 30). Drawing does not have constant scale. Outcrop is approximately 50 feet (15 meters) high at highest point.

included northward transport of the main and Tully bodies and the formation of the Temple Hill syncline itself (see Robinson et al., 1991, Figure 9). Nevertheless, a rational unwinding of these later deformations leads to the conclusion that the original transport direction for this nappe was from east to west. Further, if one accepts the tentative conclusion that the North Orange nappe lies in rocks structurally above the Brennan Hill thrust, then the trace of the Brennan Hill thrust must also lie in this outcrop, probably at or close to the contact between the narrow belt of Partridge Formation and the gneiss in the core of the Tully body.

The next outcrops to the east on Route 2 (visible from here) contain both gray and rusty schists of the Rangeley Formation on the east limb of the dome-stage anticline. The gray schists are similar to those on the west limb. These rocks are typical of a wide range of rocks above the Brennan Hill thrust. Where the schist has not suffered secondary alteration it consists of quartz 30-50%, biotite 20-40%, muscovite 15-30%, garnet 2-5%, and minor sillimanite and graphite. The assemblage is thus characteristic of the zone above the breakdown of staurolite, but below the first occurrence of sillimanite plus K-feldspar. The abundant pegmatite segregations that may be a product of partial melting, consist of about equal proportions of quartz and sodic oligoclase with minor muscovite, biotite, and garnet. The schist is commonly rich in biotite at contacts of segregations.

A curious pegmatite dike occurs on the eastern contact of the gneiss of the Tully body. It has the appearance of having been generated by partial melting in the gneiss of the Tully body, and has been injected through the contact amphibolite (note discordant contacts) into the overlying Partridge Formation. Since intrusion the pegmatite and its country rock have been folded in a series of large normal asymmetric folds.

Return to pavement westbound on Route 2 for a short distance.

57.6 Go to far end of grass strip on right beyond one section of guard rail. This is firm in all weather and is the parking spot for STOP 8, which is in a large road cut on Bachelder Road north of Route 2 where there is essentially no traffic. Trail leads from northwest end of grass strip 100 feet to low point of fence. In clear weather WATCH OUT FOR FALLING PARACHUTISTS!

STOP 8 MONSON GNEISS OF THE NORTH ORANGE BAND INTERPRETED AS THE BASEMENT CORE OF THE NORTH ORANGE FOLD NAPPE (30 MINUTES) See Figure 8. In this cut is exposed the contact between the Partridge Formation next to the main body of Monson Gneiss and the North Orange band of Monson Gneiss to the east, which is interpreted to occupy the core of the North Orange fold nappe. If the interpretation given under Stop 8 is correct, then all of these rocks lie above the Brennan Hill thrust, though the rocks near the west end of the cut may lie fairly close to its trace.. The Partridge consists of purplish, rusty-weathering plagioclase-biotite schist with minor garnet and sillimanite and abundant amphibolite layers. The Monson consists of interlayered plagioclase-biotite gneiss and hornblende amphibolite. A few layers contain garnet and anthophyllite that are not characteristic of the Monson elsewhere.

The dominant structural features are gently plunging, open to isoclinal anticlines and synclines parallel to the regional trend of minor folds and lineations believed to have formed during the dome stage. The folds show a normal sense of asymmetry indicating a major synclinal structure to the west, in this case the Williams Pond syncline that runs down the center of the main body of Monson Gneiss. Prominent amphibolite boudins in gneiss and schist have neck lines normal to fold axes suggesting dome-stage north-south extension. In the schist there are several examples of earlier folds with folded axial surfaces.

In the gneiss there are several examples of earlier lineations and folds folded across the later folds. The most spectacular example is at the top of the third gneiss outcrop on the north side of the road. Here a gently plunging late anticline has a biotite lineation parallel to its axis. A prominent earlier lineation is folded across the crest and a related isoclinal fold, plunging southeast, can be seen on the east limb. This earlier fold is clearly a fold in foliation and closely resembles 2nd-stage folds assigned to the **backfold stage** in large outcrops in the Quabbin Reservoir area. Even the movement sense of this early fold can be worked out and indicates that structurally higher rocks moved north. This is at least conceptually consistent with the postulated northward flowage of the main body of Monson Gneiss during the backfold stage, which is believed to have involuted the earlier North Orange nappe. This is the best exposure showing fold interference in three dimensions that has been seen in central Massachusetts outside the Quabbin Reservoir area.

END OF TRIP. To reach Amherst follow Route 2 west to exit for Route 202 south and follow to yellow flashing light for Amherst Road in Pelham which becomes Main Street in Amherst. Distance 21.6 miles, driving time about 30 minutes.

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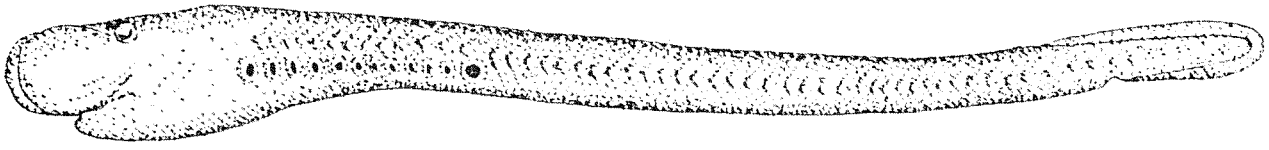

Figure 10. Reconstruction of the worm-like conodont recently identified as the first vertebrate. About 4.5 times actual size. From a drawing by Richard J. Alrdridge in Natural History as reproduced by Browne (1992).

CONTRASTING P-T-t PATHS: THERMOCHRONOLOGY APPLIED TO THE IDENTIFICATION OF TERRANES AND TO THE HISTORY OF TERRANE ASSEMBLY, SOUTHEASTERN NEW ENGLAND

by
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INTRODUCTION

One of the more persistent questions relating to the tectonic evolution of eastern New England concerns the identification and time of assembly of its lithotectonic zones: the Avalon, Putnam-Nashoba, Merrimack Trough, Central Maine, and Bronson Hill (Fig. 1). Most attention has been given to the Avalon zone, with some arguing that its arrival in the Middle Ordovician caused the Taconic orogeny, some favoring an early Devonian arrival causing the Acadian Orogeny, and some holding that at least the last stages of assembly occurred during late Paleozoic Alleghanian orogeny. Stockmal et al. (1990) and Van der Pluijm et al. (1990) have proposed that initial collision occurred in the Silurian, but continued in a protracted way into the late Paleozoic. Discrimination among these hypotheses of punctuated or protracted assembly has been stymied by the high grades of metamorphism of rocks of all zones, the lack of temporal control of metamorphism(s), and the scarcity of fossils. On this field trip we explore some of the methods and evidence for contrasting pressure-temperature-time (PTT) paths for the rocks of the various zones, and their significance in unraveling the tectonic history.

THERMOCHRONOLOGY

Thermochronologic data have contributed greatly to resolving ages of metamorphism of the various zones. Mineral ages are well known to be reset by later heating events, and in fact the ages of many minerals in igneous or high grade metamorphic rocks reflect the time of cooling, rather than the time of crystallization. Closure temperatures, or the temperatures below which minerals effectively retain radiogenic daughter isotopes, are for argon: hornblende, 500 $\pm$ 50°; muscovite, 350 $\pm$ 25°C; biotite, 300 $\pm$ 25°C; and K-feldspar, 200 $\pm$ 50°C (McDougall and Harrison, 1988), and for lead: sphene 575 $\pm$ 50°C, and monazite 720 $\pm$ 20°C (Cliff and Cohen, 1980; Copeland et al., 1988). Crystals of these minerals that grew below their respective closure temperatures would "freeze in" the time of crystallization because temperatures would not have been high enough to allow the diffusive loss of the radiogenic daughter isotope. However, the apparent ages of crystals that were present when the rocks were hotter that these closure temperatures would reflect the time of cooling of the rocks to temperatures less than the closure temperature, regardless of their petrologic origin.

LIMITS OF ALLEGHANIAN METAMORPHISM

One approach to constraining the time of final assembly of these lithotectonic zones is to establish the extent of Alleghanian metamorphism in eastern New England. If Alleghanian metamorphism were later than important motion on terrane boundaries, then Permian isograds would cut across zone boundaries, and the cooling histories of the rocks would be uniform across terrane boundaries. High grade Alleghanian metamorphism is well documented in the Avalon zone, with its Permian peak metamorphism and relatively rapid Permian-Triassic cooling. In spite of the commonly high grades of metamorphism reached, Alleghanian overprint in the rocks west of Avalon only reaches greenschist facies (Zartman et. al., 1970). This late Paleozoic metamorphism contrasts strongly with metamorphism in the Putnam-Nashoba zone, with its slow cooling from high-grade Silurian or older peak metamorphic conditions (Wintsch et al., 1992). Preliminary data from the Central Maine zone also indicates pre-Permian cooling, showing that it too, escaped Alleghanian metamorphism. The major differences in cooling histories among rocks of these zones require the final juxtaposition of rocks of the Avalon, Putnam-Nashoba, and Central Maine zones to younger than peak-Alleghanian metamorphism.

This field trip is designed to show the relevance of thermochronology to understanding the history of metamorphism of rocks at the outcrop scale and to belts of rocks on the terrane scale. It will emphasize that much metamorphic history can be extracted from a metamorphic rock in spite of a low variance mineral assemblage that does not constrain narrowly the metamorphic conditions. The contrasting PTT paths reflected

by the mineral ages of the rocks in the three lithotectonic zones we will visit (Avalon, Putnam-Nashoba, and Central Maine zones) require that significant assembly of these zones post-dates middle to late Paleozoic metamorphism in each of these zones. Much of the data upon which the trip is based has been presented in Wintsch et al. (1991; 1992), and prior reference to these is suggested. Much of the text that follows has been drawn from these references. I follow Zartman's (1988) suggestion of using the neutral term 'zone' to refer to the various belts of rocks with common sedimentary, metamorphic, and/or plutonic histories. Subzones of the Avalon zone are identified as Esmond-Dedham, Hope Valley, Massabesic, and Pelham, with Avalonian rocks exposed in the Willimantic window correlated with the Hope Valley zone (Wintsch et al., 1990). I follow Zartman (1988) in referring to all rocks structurally overlying the Putnam-Nashoba zone as Central Maine zone, while recognizing that rocks in the Merrimack trough extent south into Connecticut (Stop 5; Robinson and Goldsmith, 1991).

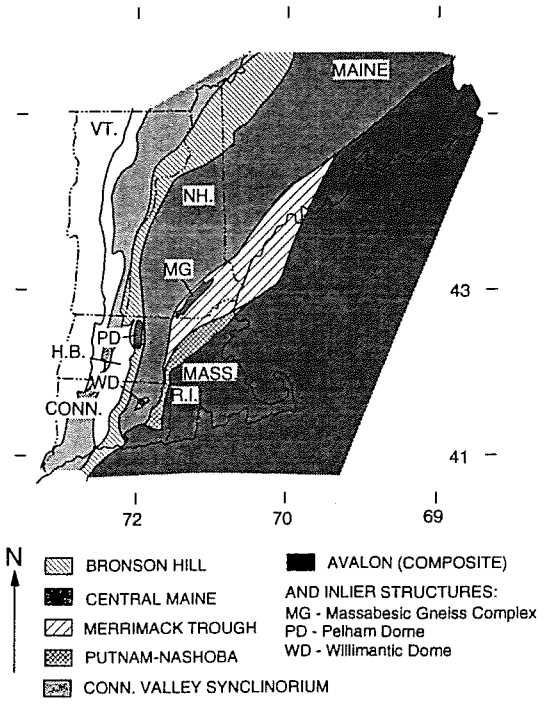

Figure 1. Generalized geologic map of eastern New England, showing the distribution of lithotectonic zones (after Zartman, 1988; Hutchinson, et al., 1988). The locations of inliers and windows exposing rocks of the Avalon composite terrane are included. H.b., Hartford basin.

GEOLOGIC AND METAMORPHIC SETTING

Avalon Zone. The Avalon zone consists of late Proterozoic rocks of coastal New England (Fig. 1), but also includes three inliers of rocks in windows or domes in central New England west of the main outcrop area. These three structures: Willimantic window, Pelham dome, and Massabesic Gneiss complex, share with the main body of the Avalon zone a Late Proterozoic crystallization age and a late Paleozoic metamorphism, and are thus correlated with it. The presence of these structures indicates that Avalonian rocks structurally underlie the zones to the west at least as far as the Hartford basin. This common history further suggests that faults that cut this terrane (e.g. Hunts Brook, Hope Valley, Beaverhead, Fig. 2) do not accommodate km scale displacement.

In the area of this field trip, the Avalon zone, including the eastern Esmond-Dedham and the western Hope Valley subzones (O'Hara and Gromet, 1985), contains primarily Late Proterozoic felsic plutonic and volcanic, and minor metasedimentary rocks (Skehan and Rast, 1990). These are cut by Ordovician, Devonian, and late Paleozoic intrusive rocks, and several pre-, syn-, and post-metamorphic faults. East of the Hope Valley fault the grade of metamorphism varies from unmetamorphosed Late Proterozoic igneous rocks, and Cambrian and Pennsylvanian sedimentary rocks in eastern Massachusetts to upper amphibolite facies in southwestern Rhode Island. Permian isograds (Fig. 2) have a general northwest strike, with the Ponaganset Gneiss in northwestern Rhode Island (see STOP 1) lying approximately between the staurolite and sillimanite isograds.

In the Hope Valley subzone, this metamorphic gradient persists from northeastern Connecticut where muscovite + biotite ± garnet-bearing rocks conspicuously lack migmatites and feldspar-bearing veins (Dixon, 1974; Moore, 1983; STOPS 2, 4), to the south and west, where anatectic conditions prevailed (Lundgren, 1966; Wintsch and Aleinikoff, 1987; Dipple et al., 1990). Amphiboles in the northern part of the Hope Valley zone are fine grained and acicular (near Stop 2; Stop 4), but south of Stop 4 they are coarse grained, nearly equant,

show straight extinction, and generally share smooth, straight boundaries with all adjacent grains. The time of metamorphism of these rocks is quite uniform; hornblende cooling ages are uniformly late Paleozoic (Table 1; Fig.2), which precludes any km scale tilting since late Permian times. Timing of peak metamorphism is about 275 Ma, in the Esmond-Dedham (Dallmeyer et al, 1990) and the Hope Valley and Willimantic subzones (Wintsch et al., 1992; Table 1; Fig. 3).

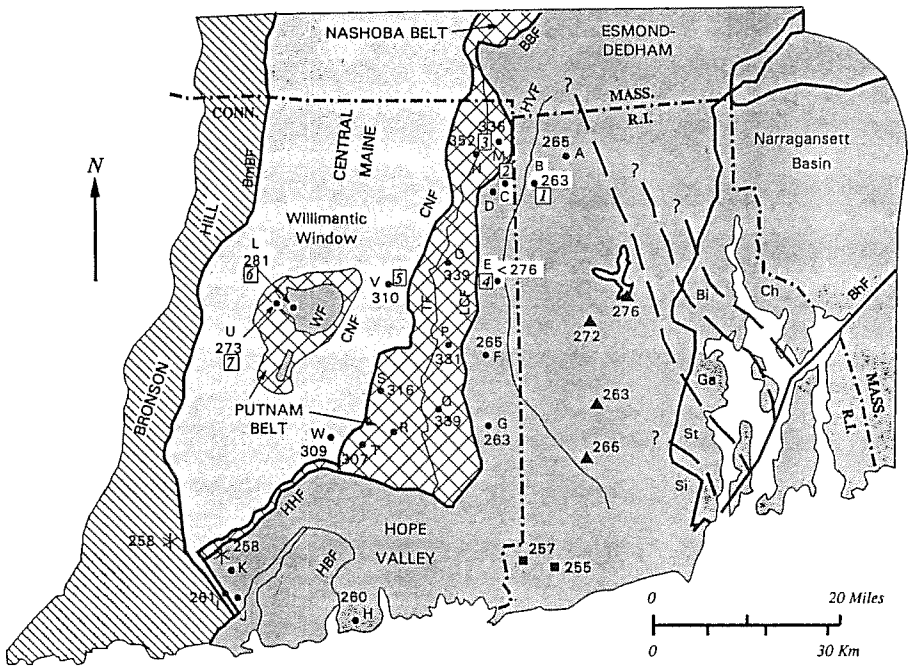

Figure 2. Generalized geologic map of southeastern New England, compiled from Wintsch et al. (1992) showing the distribution of lithotectonic zones, selected bounding and internal faults, and Alleghanian isograds in the Pennsylvanian Narragansett basin and adjacent crystalline rocks. Hornblende cooling ages by the $^{40}Ar/^{39}Ar$ method are from Dallmeyer (1982, ■), Dallmeyer et al. (1990, ▲), Wintsch and Sutter (1986,*), and this study (●). Letters record locations of samples keyed to Table 1. For other explanation, see text. Abbreviations are: BBF, Bloody Bluff fault; BhF, Beaverhead fault; BmBF, Bonemill Brook fault; CNF, Clinton-Newbury fault; HBF, Hunts Brook fault; HHF, Honey Hill fault; HVF, Hope Valley fault; LCF, Lake Char fault; TF, Tatnic fault; WF, Willimantic fault; Ch, Bi, Ga, St, Si, chlorite, biotite, garnet, staurolite, and sillimanite zones, respectively. Patterns of lithotectonic zones are given in Fig. 1.

Putnam-Nashoba Zone. The Putnam-Nashoba zone contains Late Proterozoic (?) pelitic metasedimentary and mafic metavolcanic rocks (Goldsmith, 1991) all of which are metamorphosed to upper amphibolite facies conditions (STOP 3). These metamorphic rocks were intruded by several Silurian mafic rocks (Zartman and Naylor, 1984) and cut by several ductile faults (Dixon and Lundgren, 1968; Zen et al., 1983), including the Tatnic fault (Fig. 2). Available thermochronology suggests very slow cooling (3°C/m.y.) from a Silurian or older upper amphibolite facies metamorphism to the Triassic (Table 1; Fig. 3). Data from Putnam-Nashoba rocks of both the Putnam belt and the Willimantic window define parallel cooling curves of nearly identical age that support the lithotectonic correlation of the two bodies of rock. In the eastern Putnam belt (Fig. 2) Alleghanian resetting was minimal (Stop 3); even muscovite apparent ages are consistent with slow cooling (Fig. 3). However, in the Willimantic window (Stop 7) hornblende has been reset, while sphene has not.

TABLE 1. SUMMARY OF ARGON ISOTOPIC RESULTS

SAMPLE Field	Fig. 2	MIN,	AGE, ±(Ma)	%^{39}Ar	NO. STEPS / TOTAL	MSWD	^{40}Ar/^{36}Ar, ±	COMMENT
Avalon Zone, East of Hope Valley Fault: Esmond-Dedham Subzone.								
88-101A	A	Hb	265,1	75.3	5/14		atmos	plateau age
88-102	B	Hb	263,1	70.6	4/11		atmos	plateau
		Bi	249,1	100	1/1		atmos	total fusion
Avalon Zone, West of Hope Valley fault: Hope Valley Subzone.								
88-106	C	Mu	249,1	60.3	4/8		atmos	plateau age
		Bi	250,1	100	1/1		atmos	total fusion
88-213	E	Hb	276	13.8	1/10		atmos	minimum age
		Hb	277,8	98.3	9/10	1.00	850,103	isochron
		Mu	250,1	80.0	7/10		atmos	plateau age
		Bi	244,1	100	1/1		atmos	total fusion
88-218	F	Hb	265,1	90.1	6/10		atmos	plateau age
88-306	G	Hb	263,1	69.7	7/10		atmos	plateau age
88-401	H	Hb	260,1	60.2	7/11		atmos	plateau age
84-413	I	Mu	250,1	90.2	6/7		atmos	plateau age
		Hb	261,1	74.5	5/9		atmos	plateau age
		Bi	256	100	7/7		atmos	total fusion
Avalon Zone: Willimantic Window								
88-226A	L	Hb	287	6.3	1/9		atmos	minimum age
		Hb	281,3	88	8/9	1.8	566,23	isochron
88-226B		Bi	243,1	100	1/1		atmos	total fusion
		Kf	228	100	15/15		atmos	total gas
88-226C		Mu	247,1	82.7	6/9		atmos	plateau age
Putnam-Nashoba Zone (Eastern Belt): Quinebaug, Tatnic Hill Formations								
88-112	M	Hb	370	23.5	1/8		atmos	minimum age
			336,5	100	8/8	1.11	2616,135	isochron age
88-116	N	Hb	361	15.6	1/10		atmos	minimum age
			348,2	77.4	10/10	0.12	1533,286	isochron age
88-203	O	Hb	339,2	47.5	2/9		atmos	minimum age
88-221	P	Hb	352	24	1/11		atmos	minimum age
			331,28	100	11/11	2.4	2646,1060	isochron age
88-313a	Q	Hb	339,2	69.7	6/12		atmos	plateau age
			334,11	30	6/12	0.62	3915,110	isochron age
88-329	R	Mu	271	100	11/11		atmos	total gas
			275,2	36.2	4/11		atmos	preferred age
Putnam-Nashoba Zone (Tatnic Hill Formation): Willimantic Window								
88-227	U	Hb	306	24.1	2/11		atmos	minimum age
			279,16	95.2	7/11	0.64	1329,295	isochron age
Central Maine Zone (Oakdale [Hebron] Formation)								
88-201	V	Hb	310,2	53.8	4/12		atmos	plateau age
88-202	(V)	Mu	252,1	67.8	8/14		atmos	plateau age
88-318	W	Hb	311,2	50.3	4/13	0.47	455,46	isochron
Central Maine Zone? (Yantic Member, Tatnic Hill Formation)								
88-225	S	Hb	316,4	80.0	5/11		atmos	minimum age
88-317	T	Hb	323	22.0	1/8		atmos	minimum age
			307,9	98.9	7/8	1.4	1085,193	isochron age

Note: Hb, hornblende; Mu, muscovite; Kf, K-feldspar.

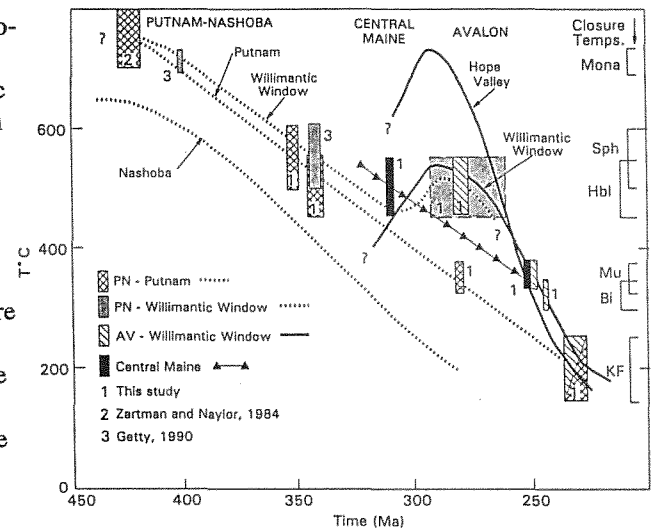

Figure 3. Summary of the thermal histories of selected blocks in the Avalon, Putnam-Nashoba, and Central Maine zones in southeastern New England (adapted from Wintsch et al., 1992); closure temperatures are indicated along the right axis. Curves of the Hope Valley and Willimantic subzones of the Avalon zone show peak Permian metamorphism; earlier prograde metamorphic history is inferred from rocks to the east. The thermal histories of the Massachusetts Nashoba and Connecticut Putnam belts contrast strongly with those in the Willimantic window, where hornblende (large, shaded box) but not sphene are reset. This documents a ~550 °C Alleghanian overprint on these rocks that is not present in the eastern belts. Preliminary data on the Central Maine zone suggest cooling rates similar to those of the Putnam-Nashoba zone. Alleghanian overprint here is unlikely because the structurally lower Putnam-Nashoba rocks do not show this overprint, and these rocks may indeed reflect slow cooling from Acadian metamorphism.

Central Maine Zone. Rocks of the Central Maine zone as used here include all rocks structurally overlying Putnam-Nashoba rocks, and all rocks east of the Bronson Hill zone (Fig. 1). Rocks of particular concern to this trip are those initially mapped as the Hebron Formation (Dixon and Pessl, 1966; Stop 5), but are now correlated with the Oakdale and Paxton Formations in Massachusetts (Robinson and Goldsmith, 1991) in Zartman's (1988) Merrimack Trough, of potential terrane status. In eastern Connecticut the ubiquitous calcarious metasiltstones (Oakdale [Stop 5] and Paxton) and argillaceous metasediments (Scotland Schist member) are metamorphosed to the staurolite-kyanite zone of the epidote-amphibolite facies conditions. Preliminary thermochronology in these rocks shows that hornblendes have not been affected by Alleghanian metamorphism; they show plateau or isochron ages of about 310 Ma (Fig. 3). Muscovite from the Scotland Schist member, however, does show a ~250 Ma age, consistent with both slow cooling and with Alleghanian resetting (Fig. 3; Table 1).

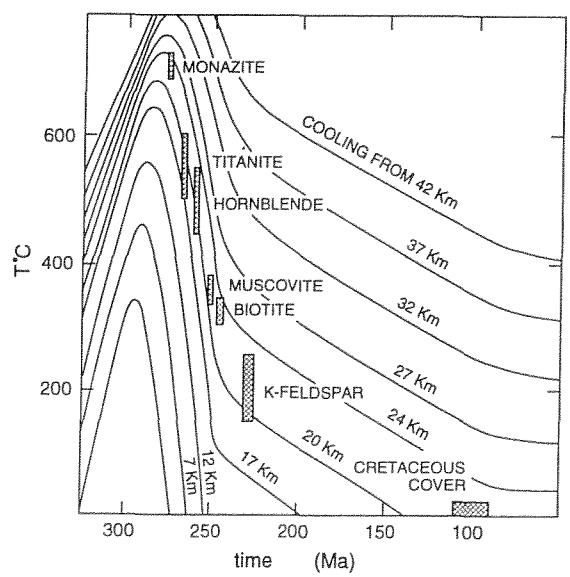

Figure 4. Summary of mineral cooling ages from the southwestern Avalon terrane, south-central Connecticut (Table 1) superimposed on model PTT paths for these rocks. Shaded box at 100 Ma is the age of Cretaceous cover and records the latest possible time of exhumation. Isopleths labeled in km contour temperature-time paths modeled following Haugerud (1986). Rocks presently exposed at the surface are monitored with the 22 km isopleth. Other isopleths predict the metamorphic histories of rocks above the present erosion surface, and now exhumed (<20 km) and those still buried below the surface (>24 km). These calculations show that in the early Permian rocks within a few km above those now exposed would have cooled less than 5 m.y. before those dated.

TECTONIC SIGNIFICANCE OF THERMOCHRONOLOGY

AVALON 'COMPOSITE' ZONE

The Avalon zone is commonly referred to as a composite terrane, composed of several 'subzones,' some of unique metamorphic history. However, available thermochronologic data do not support this idea, because the metamorphosed portion of the Esmond-Dedham subzone reflects a metamorphic history analytically indistinguishable from the Hope Valley zone. Moreover, this metamorphic history is very similar to that of Alleghanian inlier rocks (Fig. 1), all reflecting nearly simultaneous late Paleozoic metamorphism and Permo-Triassic cooling of the entire metamorphosed portion of the Avalon 'composite' terrane. Thus the rocks of this zone behaved as a single thermal block throughout the Late Paleozoic, and any independent history of the subzones must predate apparently uniform Carboniferous metamorphism.

HISTORY OF FAULT MOTION: LAKE CHAR FAULT

The cooling history of the rocks from the Putnam-Nashoba zone contrasts strongly with those of the Avalon zone. Peak metamorphic conditions in these rocks were Silurian or older, while in the underlying Avalonian rocks they were late Paleozoic (Fig. 4). The 80 m.y. difference between the hornblende cooling ages (Fig. 3) thus precludes any possibility of thermal equilibrium between Putnam-Nashoba and Avalon zone rocks during the Alleghanian heating of the latter (Wintsch et al., 1992). On the contrary, it requires significant post-metamorphic displacement along the Honey Hill-Lake Char-Bloody-Bluff fault zone. The truncation of isograds at the western edge of the Avalon zone by the Lake Char and Bloody Bluff faults further indicates important motion on this fault zone that post-dates peak metamorphic conditions in both the hanging wall and foot wall rocks. The data require that: (1) Alleghanian metamorphism in the Avalon composite terrane occurred elsewhere, remote from rocks now exposed in the Putnam-Nashoba zone, and (2) peak metamorphism in the Avalon terrane predates significant movement on the Honey Hill fault system (contrary to O'Hara, 1986).

HISTORY OF FAULT MOTION: CLINTON-NEWBURY FAULT

The Clinton-Newbury fault is an important fault in Massachusetts, but is poorly recognized in Connecticut, perhaps because the change of metamorphic grade across the potential fault zone is not large as it is in Massachusetts. Preliminary data form the Paxton (Hebron) Formation (Stop 5) show that these hornblendes have not been reset by Alleghanian metamorphism, but reflect Pennsylvanian cooling. This age is younger than that in the Putnam-Nashoba rocks. Because rocks structurally overlying and in thermal equilibrium with Putnam-Nashoba rocks would have cooled *before* the rocks below them, hornblendes from the Paxton Formation would be expected to predate the Putnam-Nashoba rocks by approximately 5-10 m.y. They could not have cooled after the rocks of the Putnam-Nashoba zone, if they had been overlying them during the cooling of both. Therefore, the rocks of the Central Maine zone must be allochthonous, and have overridden the Putnam-Nashoba rocks since ~310 Ma, probably in the early Permian, along an extension of the Clinton-Newbury fault zone.

The location of the Clinton-Newbury fault in Connecticut is not known in detail. Pease (1989) places in an the Canterbury Gneiss, interpreted to be a syntectonic intrusion. However, Paxton (Hebron) Formation rocks are present on both sides of the gneiss, suggesting that it is internal to this unit. The upper unit of the Tatnic Hill Formation (Yantic Member) is everywhere blastomylonitic, and would be a possible location. However, in the southwestern part of the Putnam belt, hornblendes (samples S and T, Fig. 2) exhibit discordant ages in agreement with the ~310 Ma age of hornblende in the Central Maine zone suggests that the boundary between these rocks must lie below the Yantic Member, or about 3 km east of the boundary indicated on Fig 2.

ASSEMBLY OF LITHOTECTONIC ZONES IN SOUTHEASTERN NEW ENGLAND

The implication of the thermochronologic data from the Avalon, Putnam-Nashoba, and Central Maine zones is that the Alleghanian metamorphism did not regionally affect eastern New England in its present geometry. On the contrary, this high-grade metamorphism was apparently restricted to the Avalon zone, and was transported

with the Avalon zone to New England. The direction of motion was apparently from the east, with Putnam-Nashoba rocks thrust eastward over Avalonian rocks. This is consistent with the occurrence of Central Maine rocks overthrusting Putnam-Nashoba rocks, also from the east.

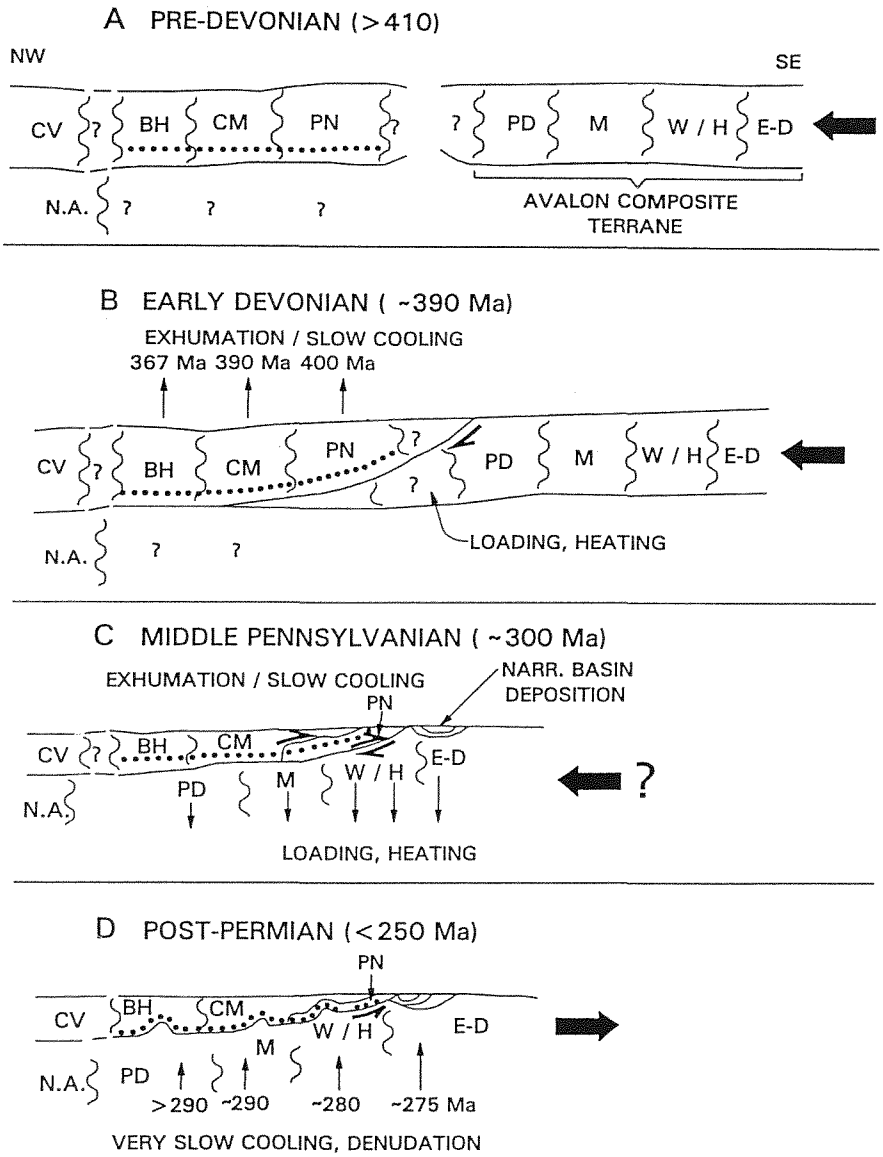

Figure 5. A set of schematic cross sections showing the constraints placed on terrane assembly by thermochronologic data. Cross sections group lithotectonic zones into belts with common cooling histories. Wavy vertical lines emphasize uncertainty on the timing and nature of the possible displacement on terrane or zone boundaries. **A.** Possible configuration of zones in pre-Devonian times. The Avalon zone is shown as composite. North American (N. A.) Grenvillian rocks are shown under Connecticut Valley synclinorium (CV) rocks; other lower crustal rocks are unknown. A west-northwest relative motion of Avalon is implied. **B.** The onset of cooling of the Putnam-Nashoba (PN), Central Maine (CM), and Bronson Hill (BH) zone rocks as reflected by monazite cooling ages. **C.** In the mid Pennsylvanian Avalon zone rocks were being loaded and heated, while the structurally higher rocks were still undergoing slow, continuous exhumation. **D.** By post-Permian time Avalonian rocks of the Pelham dome (P), Massabesic Gneiss (M), Hope Valley (H), and Esmond-Dedham (E) zone had reached peak metamorphic temperatures, and had begun to cool (at the ages indicated). The major loss of section (~10 km in the east) required by the disparity of amphibole cooling ages between the Avalon and Putnam-Nashoba zone rocks is indicated by the dotted line in the Putnam-Nashoba zone. In post-Permian time, Mesozoic extension was important, followed by very slow cooling and denudation, caused by isostatic not tectonic forces.

These thermal relationships allow reconstruction of the history of assembly of these zones to each other (Wintsch et al., 1992; Fig. 5). In Figure 5a lithotectonic zones are grouped together based on common thermal histories; their physical relationships to one another in space and time is not known. In pre-Devonian times, rocks of the Connecticut Valley Synclinorium, Bronson Hill, Central Maine, and Putnam-Nashoba zones were present east of North America, but their underlying basement rocks are not identified. To the south and east (present coordinates) lay the rocks of the Avalon zone. In the early Devonian (Fig. 5b), monazite cooling ages from 400 to 367 Ma document the onset of slow and continuous cooling and exhumation of the Putnam-Nashoba, Central Maine, and Bronson Hill zones. This cooling is interpreted to reflect underplating and heating of the underthrusted rocks, but probably not of the Avalonian rocks because they do not show thermo-

chronologic evidence of Acadian metamorphism. The timing of this early stage of collision is consistent with this event being at least partially responsible for the Acadian orogeny.

The slow, constant cooling of Central Maine and Putnam-Nashoba rocks through the Carboniferous implies sustained underplating of these rocks from the east (Fig. 5c). In contrast, data from Avalonian rocks show cooling did not begin until the latest Pennsylvanian, implying heating and loading earlier in the Pennsylvanian. The overlap in time of cooling of Central Maine and Putnam-Nashoba rocks, and heating of the Avalonian rocks implies that the underthrusted rocks at this stage were Avalonian. The direction of shortening at this time is less clear: oblique motion is quite possible. In post Permian time (Fig. 5d), all three zones appear to have cooled together, indicating < km-scale motion. Late mylonites on along the Avalon boundary record extensional motion (Goldstein, 1989), consistent with other Mesozoic structures along the eastern margin of North America.

A significant result of the data presented above is that at about 280 Ma, Avalonian rocks were undergoing high-grade metamorphism, while Putnam-Nashoba rocks were at lower greenschist facies conditions. Given normal geothermal gradients, this approximately 400°C temperature difference requires the presence of at least 10 km of rock between the Putnam-Nashoba and Avalonian rocks in the Late Pennsylvanian (time of metamorphism). This thickness of rock must have been removed in the Permian, because by Triassic time Avalon and cover rocks were adjacent, and cooling together. There is no way of identifying this 10+ km thick package. The Honey Hill-Lake Char fault system cuts both Avalonian and Putnam-Nashoba zone stratigraphy, and some loss of section on both hanging and foot walls is likely. However, it is possible that an entirely different, unidentified terrane occupied this structural level. For graphical simplicity, we show all this loss of section schematically in the Putnam-Nashoba zone (section below dotted line in Figure 5a-c; section has disappeared in Figure 5d). An analogous loss of section is implied for the boundaries of the Avalon composite terrane around the inlier structures (Wintsch et al., 1992) and these are also indicated by the dotted line. The Clinton-Newbury fault in Connecticut must also have significant displacement on it, but the nature of this displacement is still under investigation. Section may have been cut out, but it is also possible that the thrust stayed in a single structural layer, and moved deeper, later cooling rocks from the west over Putnam-Nashoba rocks to the east.

CONCLUSIONS

1) The thermal disequilibrium between Avalonian basement and cover requires that the Alleghanian thermal event in Avalonian rocks occurred under rocks other than those now structurally above it; Avalon, and its high-grade metamorphism must have been transported to its present configuration within a few 10s of m.y. after peak metamorphism, in the late Permian.

2) Rocks of the Central Maine zone were thrust east over Putnam-Nashoba rocks. The timing of this thrusting must have been younger than the younger of the two ages (mid Pennsylvanian). The fault marking this boundary probably lies structurally below the upper member of the Tatnic Hill Formation.

The case for a westward directed underplating of Putnam-Nashoba rocks by Avalonian rocks in the late Permian is strong (Wintsch et al., 1992). With these preliminary data showing Pennsylvanian cooling of Central Maine rocks, thermal equilibrium between these rocks and Putnam-Nashoba rocks is also precluded. Central Maine rocks probably were thrust over Putnam-Nashoba rocks from the west, also during the latest Paleozoic. The style of assembly of these rocks seems to be following one of an accretionary prism, where North American craton acted as a buttress, while Putnam-Nashoba and then Central Maine rocks rose over Avalon and Putnam-Nashoba rocks, respectively, all during the final assembly of Pangea during the late Alleghanian.

ACKNOWLEDGMENTS

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benefitted from conversations with H. R. Dixon, R. Goldsmith, M. Kunk, L. Lundgren, S. Quarrier, M. H. Pease, and J. Sutter. I thank Norm Gray and the Department of Geology and Geophysics, University of Connecticut, for their hospitality while preparing the final manuscript. This research was partially supported by the U. S. Geological Survey, Indiana University, and NSF Grants EAR-8618305, 8803540, and 9104495.

ROAD LOG

Assemble at Ma's Frosty Restaurant at the intersection of US 44 and SR 21. This can be reached by following US 44 west from its interchange with I 395 near Putnam, Ct.

Mileage

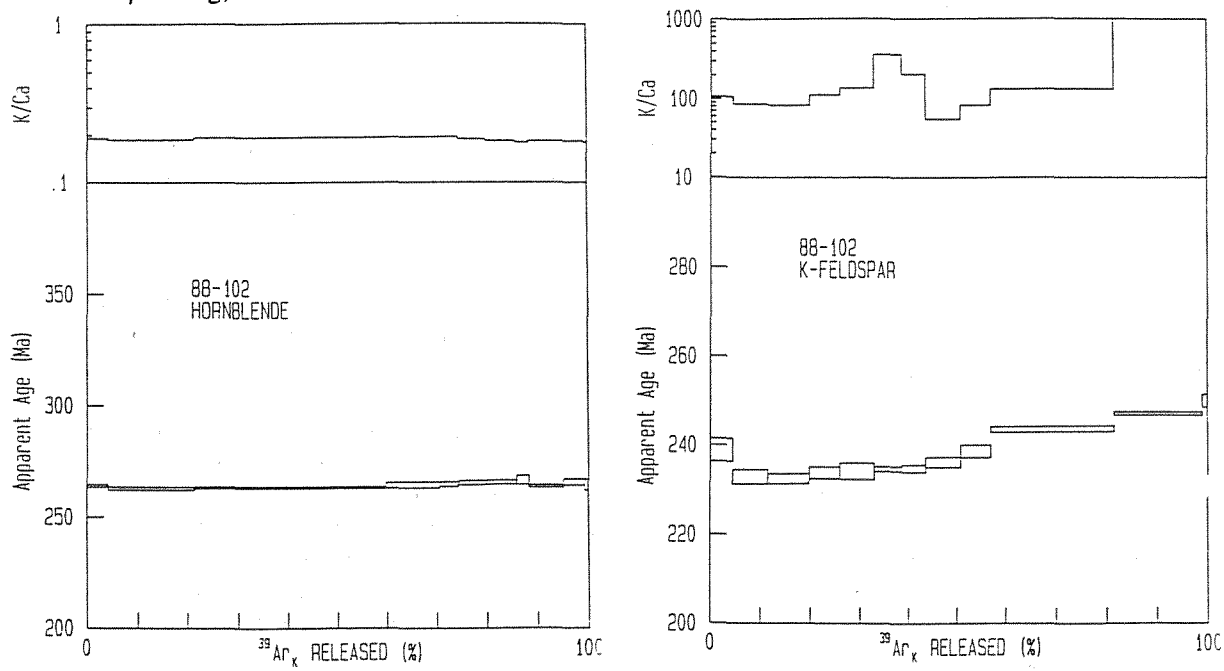
0.0 Ma's Frosty Restaurant. Follow US 44 east.

2.5 Approximate trace of the Lake Char fault, separating Avalon and Putnam-Nashoba rocks.

3.6 Rhode Island state line

5.6 Low road cuts on the left (N) side of the highway. Carefully make U-turn ark on north side of road.

Stop 1. Avalon Zone, Esmond-Dedham Subzone (45 min). This is a two m high road cut, on the north side of U.S. Rt. 44 between Bowdish Reservoir and Lake Washington; 350 m east of small bridge over Bowdish Reservoir, Rhode Island. The rock was mapped as Ponaganset gneiss (Late Proterozoic) by Dixon (1974) in the Thompson quadrangle, Connecticut. The rock is a grayish-tan, massive weathering well foliated, plagioclase, quartz, microcline, biotite, epidote, hornblende, sphene gneiss. Foliation and a conspicuous lineation (30°N15E) is defined by the parallel alignment of disseminated biotite flakes, but biotite-rich folia also give the rock a locally banded appearance. Hornblende (100 x 200 μm) is only ~2 % of the rock, occurring as anhedral equant, and as elongate grains, always surrounded by and locally embayed by biotite and epidote. These textures suggest the retrograde replacement reaction: hornblende + K-feldspar + H^+ = biotite + epidote + Mg^{++} + Ca^{++}. Equant to elongate grains of K-feldspar are 200 to 500 μm long, and show weak cross-hatch twinning. Plagioclase exsolution lamellae occur at a spacing of 50 μm, and patch perthite is rare. Biotite flakes are 200-500 μm long, and contain rare inclusions.

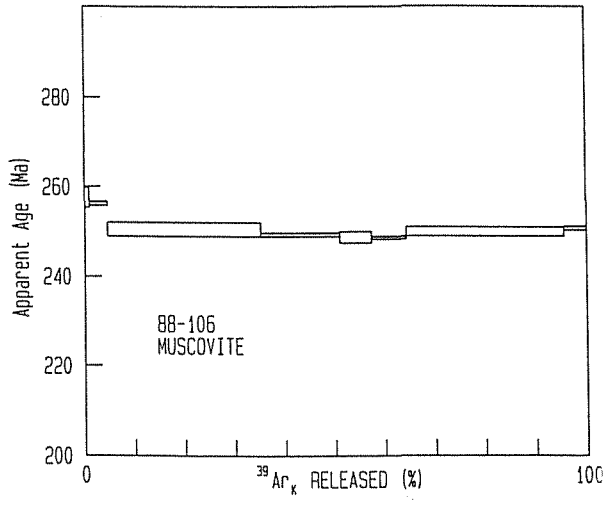


This outcrop contains a strong record of cooling from Alleghanian metamorphism. Hornblende from this outcrop plateau at 263 Ma, biotite total fusion age is 249 Ma, and K-feldspar gives early Triassic ages (Table 1). However, sphene from this rock gives a concordant age of 602 Ma (J. Aleinikoff, pers. comm.). Thus the temperature of Alleghanian metamorphism in this rock exceeded the closure temperature of hornblende, but not of sphene, and must have been ~ 550°C. In fact the age of 602 Ma is very easily interpreted to be a cooling

age from original Precambrian emplacement. This rock, then, was heated to about first sillimanite isograd conditions in the Permian, but last experienced second sillimanite grade temperatures in the late Precambrian. Given this cooling history, and a temperature estimate for hornblende replacement (above) of ~450°C, the lineation is constrained to be latest Permian.

5.7 Follow US 44 west from STOP 1.
7.1 Turn left (S) on to Putnam Heights Road, proceed west.
7.8 Small road cut on right (N) side of Five Mile River Road. Park before the cut. This is private property! You must get permission form Mr. Dexter. Cutler (log house down driveway east of outcrop) before examining top of outcrop.

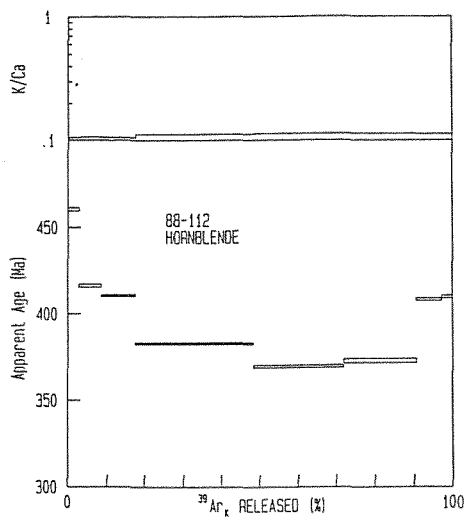
Stop 2. Avalon Zone, Hope Valley Subzone. (30 min.) These metamorphosed siliciclastic rocks were mapped as Plainfield Formation (pql, Cambrian?) by Dixon (1974) in Thompson quadrangle, Connecticut. The rocks are tan to purplish, slabby weathering, well foliated quartz, plagioclase, muscovite, biotite schist and quartzite. Ten percent of the rock contains folia of almost pure muscovite-biotite schist layers. Isoclinal folds and boudinaged quartz veins present on the top of the outcrop probably reflect strain caused by movement of the foot wall rocks to the east under WSW verging thrusts. The grade of metamorphism in this outcrop cannot be defined narrowly, because the proper mineral assemblages are not present. The lack of aluminum silicates in this outcrop and regionally probably defines temperatures less than the kyanite or sillimanite isograd (≥500-550°C). The lack of garnet or staurolite could be due to a relatively magnesium-rich or aluminum-poor bulk composition. In any case, the grade is probably lower than Stop 1, just 2 km to the east.

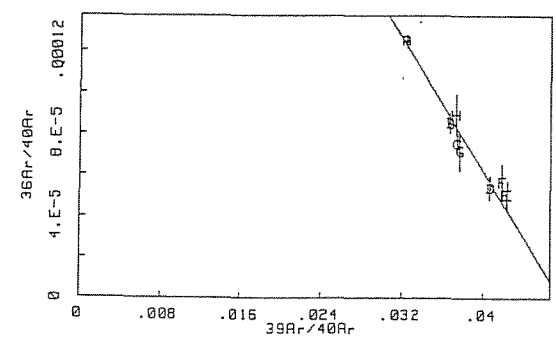


Muscovite and biotite in the quartzite occur as 50 x 20 μm flakes intergrown at the scale of 25-50 μm. Muscovite in the schist layers occurs as flakes up to 100 x 100 μm, with only a few biotite inclusions. Biotite flakes are up to 100 x 500 μm, with rare opaque inclusions. Muscovite in this sample yields a plateau age of 248 ± 1 Ma, and includes 60 % of the potassium derived ^{39}Ar. The age spectrum steps down from slightly older ages in the low temperature steps, probably reflecting excess ^{40}Ar. The biotite age is indistinguishable from the muscovite age, which can be interpreted as either a relatively high closure temperature of biotite, or the presence of excess ^{40}Ar in the structure. In view of the evidence for excess ^{40}Ar in the muscovite, the latter interpretation seems top be the more likely.

7.8 Continue west of Five Mile River Road.
8.9 Turn Right (N) on East Putnam Road.
9.4 Cross US 44, stay on east
9.7 Crossing the approximate trace of the Lake Char fault.
10.3 Bear right (N) at cemetery.
10.9 Hamlet of Quaddick. Turn left (NW) on to Quaddick Hill Road.
11.5 Outcrops of Quinebaug Formation on the south slope of Quaddick Mt.
13.4 Jct. of SR 193 and 200 in the village of Thompson, and the approximate trace of the Tatnic Fault. Continue NW on SR 200.
15.2 Intersection of Ct Rts. 200 and 12. Turn left (S) on Ct 12.
16.2 Bear left, leaving Ct 12 (to the right), and enter the ramp to I 395.
16.3 Park of the right shoulder of the ramp, before the RR overpass. Walk 100m up the ramp to stop 3.

Stop 3. Putnam-Nashoba Zone.(45 min). This outcrop is included in this trip because it shows several characteristic rock types present in the Putnam-Nashoba zone, here close to the base of the Tatnic Hill formation (Dixon, 1982) near the Tatnic fault. No samples were analyzed from this outcrop, but the amphibolites sampled from both the Putnam and the Nashoba belts from nearby outcrops are similar to amphibolites present here. The rocks of this zone are significant because they are clearly very high grade, beyond the second sillimanite isograd of the upper amphibolite facies: much higher in grade than rocks on either side of the Hope Valley fault at stops 1 and 2. Part of the argument for a net thrust motion on the Lake Char fault comes from the higher grade Putnam-Nashoba rocks occurring over lower grade Avalonian rocks across this fault. Much of the deformation in this outcrop occurred at very high temperature, and thus must be Devonian or older (Fig. 3).

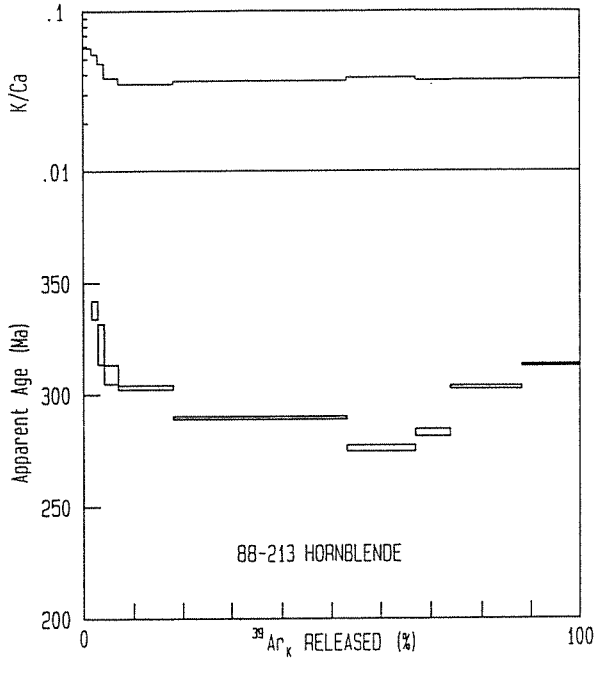


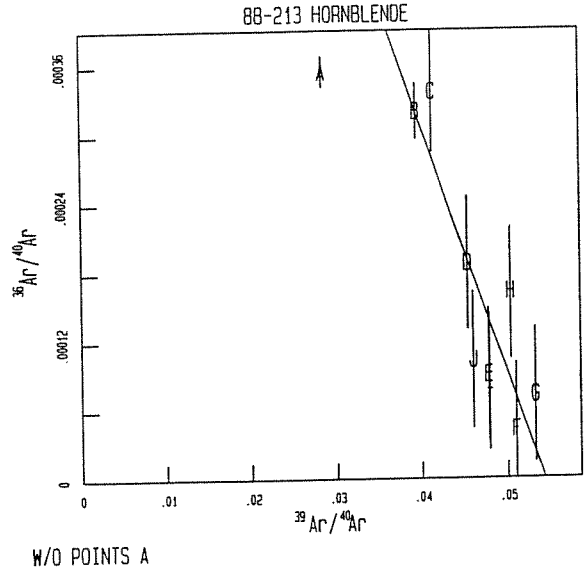


Two hornblendes from amphibolites in this area were dated by the $^{40}Ar/^{39}Ar$ method (Wintsch et al., 1992). Sample 88-112 (below) and sample 88-116 were collected from the Tatnic Hill and the Quinebaug formations, respectively, on opposite sides of the Tatnic fault, in the Thompson quadrangle (Dixon, 1974). These amphibolites were dark olive green to black, massive weathering, well foliated, hornblende, plagioclase, biotite, amphibolite, with local biotite, chlorite, sphene, and magnetite. Hornblende grains are typically equant, 1000 μm in diameter. A few very small inclusions are present in patches in most grains, but all grains have larger, inclusion free areas. Most hornblende grains show at least mild undulose extinction. Although these and three other hornblende samples from the Quinebaug Formation exhibit U-shaped age spectra that do not define age plateaus, they do form linear arrays on isotope correlation diagrams that yield "isochron" ages of ~330-350 Ma and very high initial (trapped) $^{40}Ar/^{36}Ar$ ratios (Table 1). For example, sample 88-112 (below) gave a minimum age of 370 Ma, but a correlation age of 336 $\pm$ 5 Ma, and a $^{40}Ar/^{36}Ar$ ratio of 2616 $\pm$ 135. Moreover, hornblende from sample 88-313a (Q, Fig. 2) yields a plateau age of 339 $\pm$ 2 Ma using a present-day atmospheric argon correction (Wintsch et al., 1992). The internal consistency of hornblende ages obtained from both plateaus and correlation diagrams, with sphene ages (Wintsch et al., 1992) indicates that hornblendes throughout the Putnam belt cooled through their argon closure temperature about 340 million years ago.

16.3 Return to vans. Proceed south on I 395.
27.7 Exit at interchange 91, following US Rt 6 east through South Killingly.
31.3 Cuts of Ponaganset Gneiss.
32.2 Drive just into Rhode Island, turn around, and proceed back west on I 395 access road.
33.3 Road cuts of Plainfield Quartzite
35.5 Road cuts on the right (N) side of the road just behind the Exit 90 road sign.

Stop 4. Avalon Zone, Hope Valley Subzone. (30 min.). This road cut is 210 m southwest of the northeastern-most cut on the north side of the highway joining I-395 with U.S. Rt. 6, 2.2 km southwest of South Killingly, Connecticut. It exposes rocks of the upper (quartzite) and upper middle members of the Plainfield Formation (Late Proterozoic) of Moore (1983) in the East Killingly quadrangle, Connecticut/Rhode Island. This locality has also been described by Goldstein and Owens (1985, Stop 8). Two important rock types are exposed here. Most abundant is a gray to tan, massive to locally slabby weathering, well foliated quartz, muscovite, biotite quartzite. Muscovite flakes range from 25 x 50 μm to 50 x 500 μm. Grains are inclusion free, but are locally intergrown with biotite. Biotite flakes range from 20 x 50 to 50 x 300 μm in size, with a few grains as long as 500 μm. Inclusions are rare. At the northeastern limit of the road cut, a massive, dark olive green weathering, well foliated hornblende, chalcopyrite, quartz, plagioclase, sphene, epidote amphibolite. Hornblende needles range in size from 30 x 30 to 150 x 350 μm. The grains are untwinned, and fewer than 10% of the grains contain small (<5 μm) inclusions.





Muscovite from the quartzite here yields a plateau age of 250 $\pm$ 1 Ma, and biotite gives a total gas age of 244 $\pm$ 1 Ma. These are both consistent with the regional Permian cooling throughout the Avalon zone. However, the acicular hornblende (E, Fig. 2; below) produces a highly discordant, U-shaped age spectrum, distinctly different from the plateau ages of most other Avalonian hornblendes. It has an age minimum of 276 Ma, which is conservatively interpreted to be the maximum age of closure (Wintsch et al., 1992). Reexamination of these data using isotope correlations shows a strong linear array containing 98.3 % of the gas, with a correlation age of 277 +/- 8 Ma, and a $^{40}Ar/^{36}Ar$ of 850 +/- 103 (Table 1; below). The agreement between these two ages is strong confirmation that 276 Ma is the time of closure of this hornblende. I speculate that the grade of metamorphism here is low enough that these rocks were not heated to temperatures much above closure. The sample contains excess ^{40}Ar very likely incorporated during growth, and the age may represent one of growth, and not of cooling. By this interpretation this locality, high in the structural level of the Avalon zone, would follow an isopleth on Fig. 3 of about 12 km, while rocks deeper (Stop 1) would have followed the 20 km isopleth. The further implication of this Permian age of amphibole, is that rocks of the Hope Valley subzone did not experience a metamorphism of greater than middle amphibolite grade prior to the Permian. Return to vans.

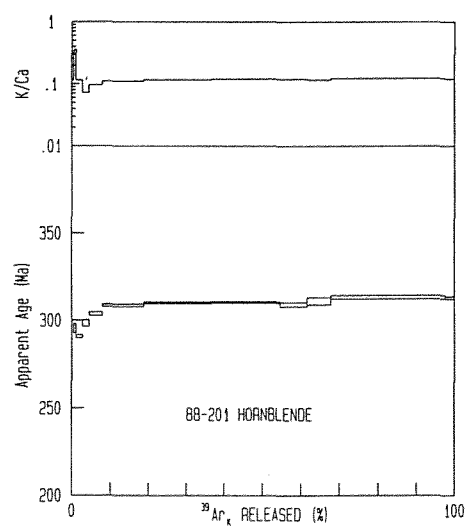
35.5 Continue west on I 395 connecter.
36.4 Leave highway at exit 90, turning right (N) on to Squaw Rock Road.

37.0 Crossing approximate trace of the Lake Char fault.
37.2 Turn right (N) on Green Hollow Road.
39.2 Turn left (W) on Franklin St.
39.4 Approximate trace of the Lake Char fault.
39.6 At flashing stop light take a ½ right to US Rt. 6, moving to the left lane around the round about.
39.8 Bear right (W) on US Rt 6.
40.6 Passing rare road cuts of lower member, Quinebaug Formation on the left, opposite McDonald's, being actively destroyed in June, 1992.
44.2 Laurel Road and approximate trace of the Tatnic fault.
46.8 Passing road cuts of Canterbury Gneiss (intrusive into Central Maine zone rocks).
48.9 Road cuts on the right (N) side of US Rt 6. Park on shoulder of highway.

Stop 5. Central Maine Zone. (30 min.). This road cut along U. S. Rt 6, 100 ft east of its intersection of St. Rt. 97 exposes relatively fresh well layered biotite schist and calc-silicate-granofels, and was included as a field trip stop by Dixon et al., (1968). The age of the hornblende in this rock contrasts with ages of hornblendes in rocks from adjacent lithotectonic zones, and indicates an allochthonous relationship between these rocks and the underlying Putnam-Nashoba rocks.

This rock is dark greenish gray and purplish gray, medium grained, layered (cm scale) hornblende granofels and biotite schist. The granofels layers 1/2-2 cm thick, contain 100-300 u grains of essential scapolite, hornblende, plagioclase and quartz, all in random orientation, with accessory zoisite, clinozoisite, calcite, sphene and zircon. Porphyroblasts of hornblende and scapolite are up to 1 mm long. Subhedral hornblende grains are between 100-1000u long, and 50-300u wide. They very commonly contain rounded inclusions of scapolite, quartz, and plagioclase, but show straight extinction, and uniform pale green birefringence. The biotite schist layers 0.5-1.5 cm thick contain essential quartz, plagioclase, biotite and scapolite, with accessory zoisite, hornblende, sphene, and zircon. All grains are anhedral except biotite, which is subhedral and evenly disseminated; it does not define biotite-rich folia. Several isoclinal Z-folds are present, which may reflect eastward vergence of thrust nappes aver these rocks.

The age spectrum from hornblende in this outcrop plateaus at 310 ± 2 Ma. With the possible exception of a small tail-down shape in the low temperature steps, suggesting a small ^{40}Ar loss in the early Permian, there is no evidence for high grade Alleghanian metamorphism. The Pennsylvanian age of this sample is distinctly different from hornblende ages in both the Avalon and Putnam-Nashoba zones. In fact the *younger* age of this hornblende requires that it cooled about 30 m.y. after the hornblende in the structurally lower Putnam-Nashoba zone rocks. Normal cooling in a stack of rocks in thermal equilibrium would require that structurally higher rocks cool before those below. Thus the younger age of these hornblendes dictates these 'Hebron' rocks to have cooled to below about 500°C elsewhere (to the west?), and have been thrusted over Putnam-Nashoba rocks since amphibolite facies conditions, or in post Pennsylvanian times.

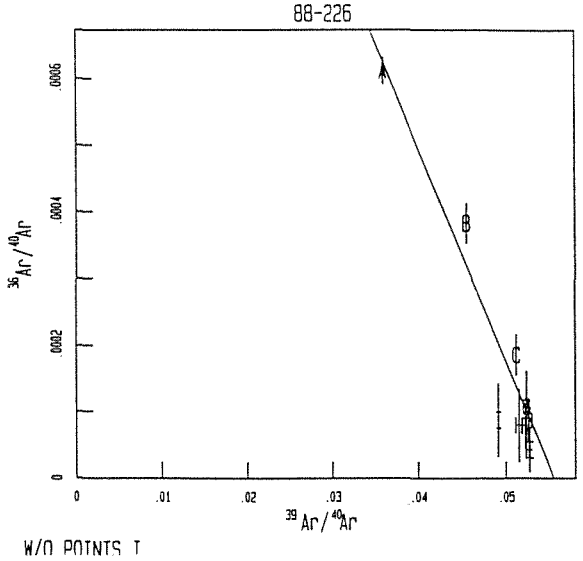


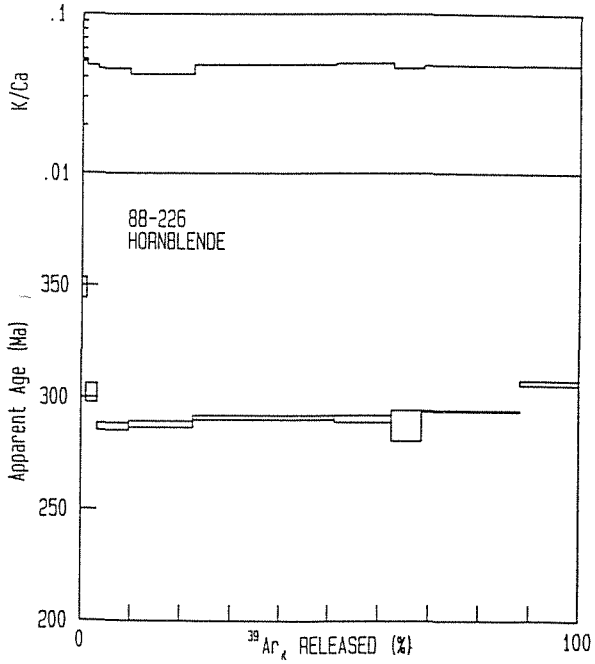
48.9 Return to vans, continue west on US Rt. 6
53.4 Crossing approximate trace of Clinton-Newbury fault zone.
54.8 Road cuts on left (S) are basal Putnam-Nashoba zone (Wintsch and Fout, 1982, Stop 3).

55.3 Crossing the Willimantic fault (Putnam-Nashoba - Avalon zone boundary).
56.9 Bear right on to US. Rt. 6 (limited access).
59.9 Exit at ramp to Ct. Rt. 32.
60.2 Park at bottom of ramp, well on to shoulder.

Stop 6. Avalon Zone, Willimantic Window. (45 min.). This road cut along CT, Rt 32 at its junction with U.S. Rt 6 (formerly I-84) is one of the best studied road cuts in the area. It was originally mapped as part of the Quinebaug Formation (Putnam-Nashoba zone) by Snyder (1964) in the Willimantic quadrangle. These rocks are now reinterpreted as the Late Proterozoic metavolanic Mansfield Hollow Lithofacies of the Hadlyme Formation (Wintsch et al., 1990), and are correlated to part of the Waterford complex in the New London area. This outcrop was first described by Wintsch and Fout (1982, Stop 1), and isotopic work has also been done by Getty and Gromet (1992). The outcrop contains several rock types, and many interesting structures. The most abundant rock type is a pale gray, massive weathering, well foliated and layered plagioclase, quartz, biotite K-feldspar, sphere, (muscovite or hornblende) granofels. K-feldspar grains range from 100-400 μm in diameter, are inclusion-free, weakly twinned, and are not perthitic at the optical scale. Biotite flakes range from 100 to 700 μm long, are nearly inclusions free, and are rarely intergrown with trace amounts of retrograde muscovite and chlorite. A second conspicuous rock type is a dark gray to black, massive, unlayered, well foliated hornblende, plagioclase, biotite, magnetite, sphene amphibolite. Hornblende grains range from 200 to 1000 μm long, and 100-500 μm wide. Inclusions of biotite are rare. Equant plagioclase grains 300-500 μm diameter are optically zoned and twinned.

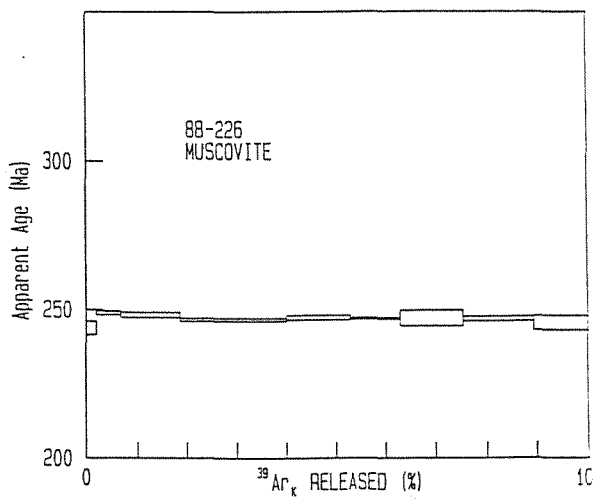
The outcrop is cut by several 0.5-1.0 m thick zoned granitic pegmatites, and by felsic sill-like layers parallel to the layering in the gneiss. The pegmatites are cored by quartz, have K-feldspar intermediate zones, and muscovite bearing margins. The sill-like structures are dominated by and are thickest where microcline-rich porphyroblasts bow out the surrounding foliation. They are joined together in necklace fashion by thinner layers of quartz and plagioclase.

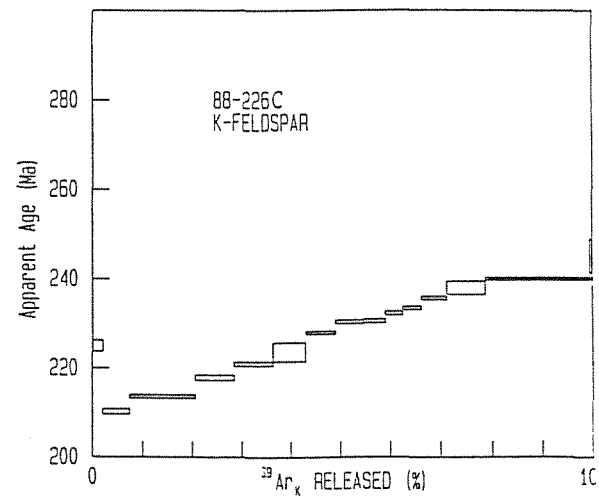




Hornblende from the amphibolite (L, Fig. 2) produced a discordant, U-shaped age spectrum with an age minimum of 287 Ma, but the same data plotted on an isotope correlation diagram produces an "isochron" age of 281 Ma (Wintsch et al., 1991). Muscovite from a cross-cutting pegmatite, and biotite, and K-feldspar from the plagioclase gneiss yield cooling ages of 247, 243, and 228 Ma respectively. The cooling curve produced by these data (Fig. 3) is very similar to that for the Hope Valley zone, except for a 15 m.y. older hornblende. Incompletely reset Late Proterozoic sphene (Getty and Gromet, 1992) in these rocks shows that the rocks did

not reach upper amphibolite facies temperatures ~550°C (sphene closure) in the Alleghanian. The temperature-time curve for these rocks must thus flatten with increasing age. This curve (Fig. 3) is quite similar to the 12 km isopleth of Fig. 4, and to the inferred history of the rocks at Stop 4. Given that the rocks at these stops are 5-10 km structurally above the rocks yielding the ages used to model the thermal history of Fig. 4, they were all apparently in thermal equilibrium during metamorphism and cooling.



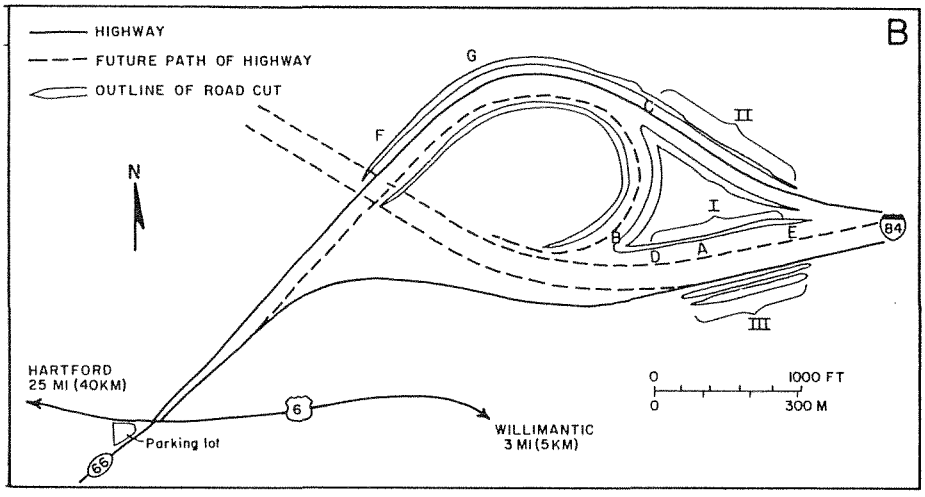


Knowledge that these rocks never exceeded 600°C is critical to interpreting the pegmatites and feldspar-rich 'sills,' in that they could not have been produced by the crystallization of silicate liquids. Partial melting is not a possibility, because even the most volatile-rich liquids do not form below 600°C. Thus all of these structures must have crystallized from an H_2O-rich liquid. The rocks also show moderate ductile deformation, but the necks between amphibolite boudins are rather straight, and are filled primarily with quartz. This is consistent with lower to middle amphibolite facies deformation, with fracturing of the relatively strong amphibolite during ductile deformation of the relatively weak quartz-feldspar host gneiss. A locally strong lineation defined by biotite streaks and quartz and feldspar rods apparently formed at about 400°C (Wintsch and Fout, 1982), and by use of the T-t curve of Fig. 4, they formed during the early Permian. Return to vans.

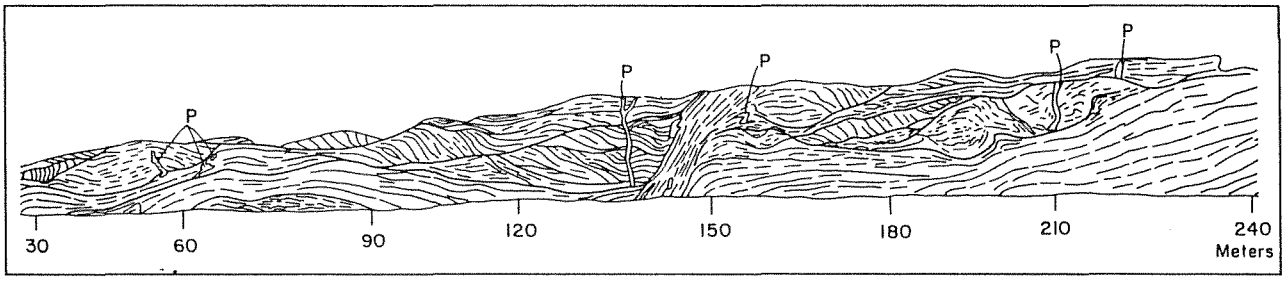
60.2 Reenter US Rt. 6, west bound.
61.0 Crossing trace of Avalon - Putnam-Nashoba zone boundary.
61.6 Drive through cuts in the Willimantic fault zone.
62.6 Park in the Park and ride, walk back up the exit ramps to Stop 7.

Stop 7. Putnam-Nashoba Zone, Willimantic Window. (60 mins.). These road cuts along the interchange of CT Rt 6 (formerly I-84) 300 m northeast of its intersection with CT Rt 66, 3 km west of Willimantic, Connecticut expose some of the most spectacular structures in eastern Connecticut. They were mapped as the Tatnic Hill Formation by Snyder (1967) and included in the garnet-biotite schist unit of Dixon (1964) by Wintsch (1979). The rocks have been previously described by Wintsch and Fout (1982, Stop 5; Wintsch, 1987, Stop 1). The most stunning features exposed in the road cuts near Willimantic are the large and very large tectonic blocks separated by anastomosing shear zones with a normal sense of motion (Getty and Gromet, 1992b; Fig. below). The dome shape of these blocks leads to quaquaversal foliation patterns in natural exposures in the Willimantic and Columbia quadrangles (Snyder, 1964; 1967) that is diagnostic of the basal Tatnic Hill Formation. Because they are usually 100ft (30m) or more long in the E-W direction, they are best viewed from a distance of 100 ft (30m) or more. Face I (Fig. below) is particularly well suited for viewing from a distance because the highway is not finished; but for those especially interested in these structures, a walk through all the cuts is imperative. The lack of continuity of the blocks or the shear zones on either side of the road along the north side of the interchange (at II, Fig. 1) indicates that these structures are not longer than 100 ft (30m) in the N-S direction, and thus the blocks must be lens-shaped. Some of these blocks may have

developed as large drag folds (e.g., 100m, Fig. below) with axes striking N30 E, evolving into these discrete blocks as strain was concentrated on the long limbs (Wintsch, 1979, Fig. 5; see also Soula and Bessiere, 1980). However, many blocks do not appear to be rotated, and some degree of mega-boudinage (e.g. 200 m, Fig. below) was probably also involved. These blocks and most small-scale structures were probably produced during southeast motion of the Putnam-Nashoba terrane over the Avalon terrane (Wintsch, 1979), and at least the latter stages of this deformation occurred in the late Paleozoic.

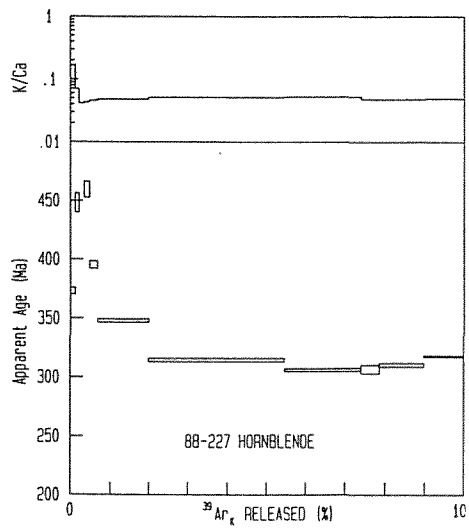


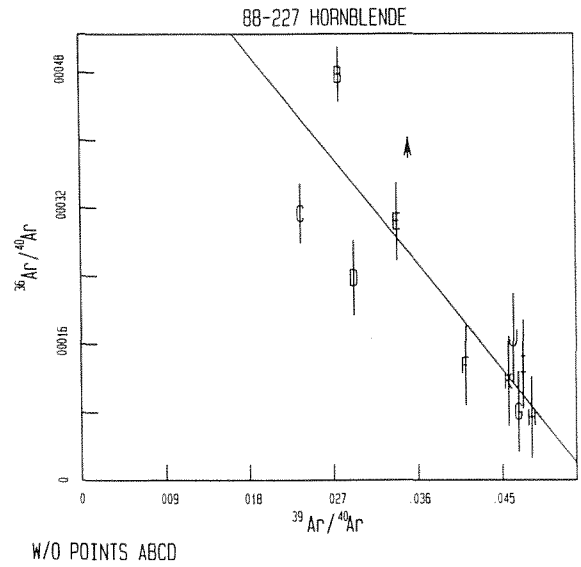
Augen gneiss, blastomylonitic gneiss, mylonitic schist, and mylonite are present in order of decreasing abundance in these rocks. The highest grade assemblages are best preserved in the augen gneisses inside and away from tectonic blocks (e.g., east end, 150 to 250 m, Fig.3). These quartz-plagioclase-biotite-K-feldspar-garnet-sillimanite bearing gneisses locally contain K-feldspar and plagioclase augen up to 5.5 in (14cm) in diameter. These augen probably grew as porphyroblasts, and did not crystallize from a melt. Evidence for this comes from rotated biotite inclusion trails in some crystals. The incorporation of these rotated inclusions demands growth of a rigid crystal in a solid matrix. Moreover, the bulk composition of the augen gneisses is strongly syenitic or monzonitic, and does not reflect the minimum melt composition of these pelitic gneisses, which should be close to a granitic eutectic. Sillimanite is ubiquitous in these gneisses, and commonly occurs as randomly oriented needles in fibrolite mats. However, some sillimanite does occur with a strong preferred orientation trending approximately N60°W, parallel to locally developed biotite streaks and quartz-feldspar rods, parallel to the overall southeast transport direction (Wintsch, 1979).



The shear zones surrounding and cutting these blocks of high-grade gneiss contain blastomylonitic schist and gneiss with lower grade kyanite-bearing assemblages (e.g., localities A, B, C below, if not collected out). Still lower grade, slabby, strongly lineated and layered mylonitic schists are well exposed on the natural cliff face III, south of the road. This schist projects under all rocks exposed in the road cuts, and its strong N-S trending lineation parallel to tight isoclinal folds suggests a similar change in fault motion direction. Another less slabby, even textured, and finer grained mylonitic schist forms a gently folded 10 to 12 ft (3 to 4 m) thick layer at D (Fig. 1) between 60 and 120 m; (Fig. 3). This later foliation is itself foliated into small isoclinal folds, but they are difficult to find because the rock almost totally lacks compositional layering. True mylonites are rare, but

late, fine-grained, middle to lower greenschist facies mylonite up to 1 in (2 cm) thick cuts the other gneisses in the steeply west-dipping shear zone which cuts the entire exposure at A (Fig. 1; 140 m, Fig.3). A brittle fracture zone occurs at E (Fig. 1) where the K-feldspar-chlorite-epidote bearing assemblage reflects very shallow alteration of this zone. Together these shear zones and their associated mineral assemblages demonstrate repeated deformation in this fault zone during shallower and lower metamorphic grade conditions.





Several other rock types are present in these cuts. Interlayered in the pelitic blastomylonitic schists and gneisses are at least 28 thin (12 in; 30cm) amphibolite layers, all boudinaged (between A and D, Fig. 1). Successive boudinage of a larger (3 ft; 1 m) amphibolite boudin at its tapering neck can be seen at 215 m (Fig. 3). Several 12- to 20-in-thick (30 to 50 cm) layers of diopside-bearing marble are present at F (Fig. 1). The margins of one of these contained hornblende porphyroblasts up to 4 in (10 cm) in diameter (now apparently collected out). A 6 ft (2 m) diameter pod of ultramafic rock, now chlorite-talc schist is present at G (Fig. 1).

Only hornblende from a massive weathering, well foliated hornblende, plagioclase, biotite, chlorite amphibolite, from the (unfinished) westbound entrance ramp was dated. It came Hornblende grains range from 0.5 to 3 mm in diameter, some with a few plagioclase inclusions. Most grains contain rectangular exsolution lamellae from 10-100 μm long. A few larger grains are composed of smaller rounded, mutually embayed, randomly oriented grains, suggestive of recrystallization and grain growth. This together with the common exsolution lamellae suggest either a complicated cooling history, or a lower amphibolite facies reheating. Other evidence for a secondary event includes the partial replacement of biotite grains 1/2 to 3 mm in diameter and by equally coarse grained chlorite, and the strong development of deformation-twinning in plagioclase grains, 200-500 μm in diameter. This hornblende (88-227; U; Fig. 2) produced a U-shaped age spectrum (Wintsch et al., 1991) with an age minimum of 306 Ma. However, an isochron containing over 98% of the $^{39}Ar_K$ in the sample defines an age intercept of ~273 Ma (Table 3), and is probably a better reflection of the time of cooling. This hornblende age is not consistent with monazite and sphene ages of 401 and 340 Ma respectively (Getty, 1990), but is consistent with Alleghanian cooling in the Avalon zone. The monazite and sphene apparent ages are easily interpreted as a cooling age from pre-Silurian metamorphism, and define a curve almost indistinguishable from that obtained from the rocks in the east (Fig. 3). Moecher and Wintsch (1991) interpret the hornblende age of ~270 Ma as reflecting Alleghanian reheating. This is consistent with the occurrence of overprinting metamorphic assemblages containing andalusite and kyanite in the ductile, Willimantic fault zone within 100 m above the Avalon zone contact (Wintsch, 1980). Indeed, complex exsolution lamellae (not present in hornblendes in the east) may reflect such heating. This Permian resetting may have been generated either by conduction across the fault or by frictional heating (Barr and Dahlen, 1989). In view of all the other data that support isotopic resetting in this fault zone (Getty and Gromet, 1989), this hornblende apparent age is probably best interpreted as locally thermally reset to a temperature > 500°C, but less than ~600°C, because sphene is not reset.

End of trip. Return to vans, return to Ct Rt 2 (Stop 6), follow Ct-MA Rt 32 north to U. S. Rt 20, Rt 20 west to Mass. Rt 181, Rt 181 north through Belchertown to Mass. Rt 9; follow 9 west to Amherst.

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THE FIRST MESOZOIC MAGMA OF THE HARTFORD BASIN: EXAMINATION OF THE FEEDER DIKE, A LACCOLITH, AND THE TALCOTT LAVA FLOW.

by
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INTRODUCTION

The igneous activity associated with the crustal extension that led to the formation of the Mesozoic Hartford basin occurred in three separate episodes. Each produced chemically distinct basaltic flows (Puffer et al. 1981) whose stratigraphic position amongst the sedimentary rocks of the basin, which exhibit Milankovitch cycles, indicate periods of quiescence between eruptions of several hundred thousand years (Olsen, 1988). The first basalt was the Talcott, followed by the Holyoke, and finally the Hampden (Fig. 1). Each basalt can be matched through a distinctive geochemical signature with one of the three regional diabase dikes that traverse southern New England (Fig. 1), the Talcott correlating with the Higganum dike, the Holyoke with the Buttress dike, and the Hampden with the Bridgeport dike (Philpotts and Martello, 1986). A fourth short dike--the Fairhaven, which occurs only in the Hartford basin, actually connects with the Talcott basalt; this dike is interpreted to be the upper part of the Higganum dike that was downfaulted into the Hartford basin by the eastern border fault (Fig. 2).

The object of this excursion is to examine the full range of magmatic features associated with the first period of igneous activity. Faulting associated with the formation of the Hartford basin has exposed the feeder dike over a range of depths. At stop one, the dike cuts Paleozoic metamorphic rocks that are estimated to have been ~10 km beneath the surface at the time the dike was intruded (Fig. 2). At stop two, Mesozoic arkose is cut by the dike at the point where it erupted onto the surface to form the Talcott basalt. Although the dike has identical compositions at these two localities, the difference in crystallization pressure caused orthopyroxene at depth to be replaced by olivine at the surface. The feeder dike is many tens of meters wide, and magma probably passed through it turbulently, as suggested by abundant partial melting of wallrocks at all depths of exposure. Assimilation of a selective fraction of this partial melt resulted in minor contamination of the basaltic magma. Not all of the magma that rose through the feeder dike was extruded. A large volume was intruded as sills and laccoliths near the base of the sediments in the Hartford basin. At stop three, we will examine the western end of the Sleeping Giant laccolith. Finally, at stop four in the Farmington River gorge at Tariffville, a complete section through the Talcott basalt is exposed, passing from the underlying sediments into the pillowed lower part of the flow, through the upper part with large half-moon vesicles filled with zeolites and datolite, to the flow-top breccia and overlying sediments which contain salt casts.

GENERAL PETROGRAPHY AND GEOCHEMISTRY

The first magma associated with the Hartford basin produced three petrographically distinct rock types as a result of crystallization at different pressures. At the deepest exposure, which is provided by the Higganum dike to the east of the Hartford basin (Fig. 1), the magma was chilled to a porphyritic diabase containing rounded glomeroporphyritic aggregates of augite and euhedral phenocrysts of plagioclase and orthopyroxene. Where the Fairhaven dike connects with the Talcott basalt, the chilled diabase contains similar augite aggregates and plagioclase phenocrysts, but orthopyroxene phenocrysts have been strongly resorbed and the rock contains small, euhedral olivine phenocrysts (Philpotts and Martello, 1986). The Talcott basalt contains rounded glomeroporphyritic aggregates of augite, euhedral phenocrysts of plagioclase and olivine, and rare orthopyroxene phenocrysts, which are mantled by olivine reaction rims (Philpotts and Reichenbach, 1985). In rising from the depth of the Higganum exposure (~10 km) to the surface of the Earth, orthopyroxene was replaced by olivine as a primary mineral on the liquidus of the first magma in the Hartford basin.

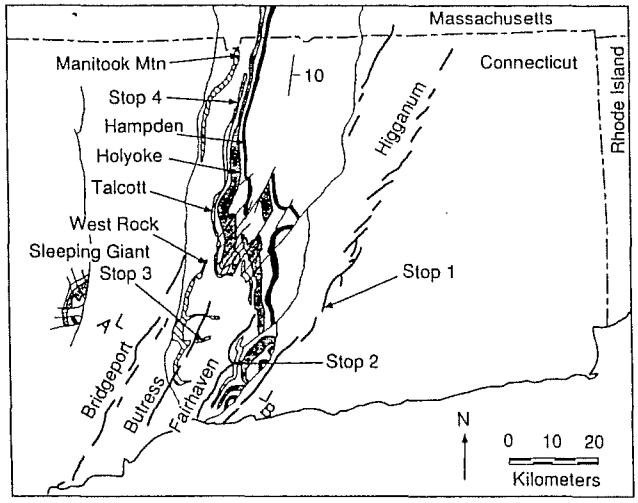


Figure 1. Igneous rocks of the Hartford basin.

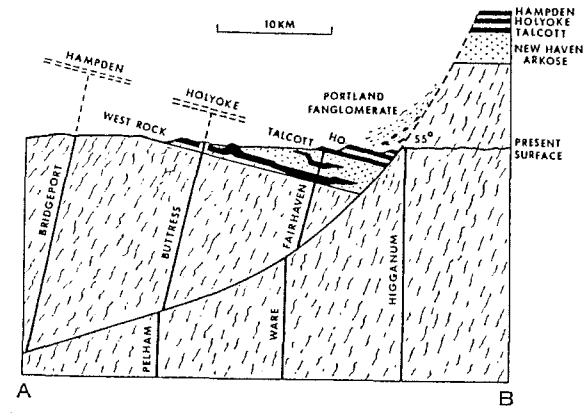


Figure 2. Cross section through southern part of Hartford basin (A-B in Fig. 1) showing relation of dikes to flows (from Philpotts and Martello, 1986).

The pyroxene phenocrysts in the Higganum dike are strongly zoned from Mg-rich cores to more Fe-rich rims (Fig. 3). In the chilled margins, the euhedral orthopyroxene phenocrysts are unusual in that they exhibit monoclinic morphology (Philpotts and Gray, 1974). Away from the margins, slower cooling allowed reaction rims of pigeonite to develop on the orthopyroxene phenocrysts, and at still greater distances from the contact, pigeonite changed to inverted pigeonite. Glomeroporphyritic aggregates of augite are of two types; one is intergrown with plagioclase and has cores containing about 1500 ppm Cr, and the other is not intergrown with plagioclase and has cores containing approximately 4500 ppm Cr. The Cr-rich cores are also slightly more magnesian than the less Cr-rich ones (Fig. 3). When plotted on the pyroxene quadrilateral of Lindsley (1983), the Cr-rich cores of augite and cores of orthopyroxene phenocrysts lie on the pyroxene solvus between 1250 and 1275°C at pressures up to 10 kb, whereas pyroxene rims and groundmass plot between 1200 and 1000°C. The high temperatures indicated for the most magnesian pyroxenes suggest that these crystals may have come from the source region of the magma and that adiabatic cooling during ascent of the magma lowered the temperature to about 1200°C at the depth of crystallization of the Higganum dike.

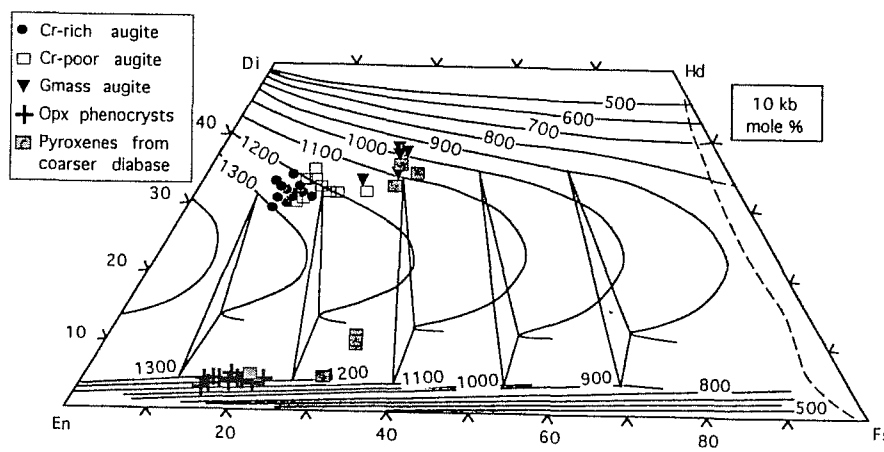


Figure 3. Plot of pyroxene compositions from the Higganum dike shown with the 10-kb pyroxene solvus of Lindsley (1983). The core compositions may have crystallized at this pressure, but the rims of crystals would have formed at lower pressures.

Although plagioclase phenocrysts in the Higganum dike are invariably euhedral, most contain a distinct rounded core surrounded by a euhedral oscillatory-zoned margin. The cores are typically An_{60} and contain approximately 1 mol % Or. The anorthite content abruptly increases to An_{74} across the core/margin boundary and then decreases in an oscillatory manner to An_{60} at the rim. The abrupt increase in anorthite is accompanied by a decrease in Or to 0.6 mol %. Within the oscillatory-zoned margin, spikes of high K_2O content are found, which we interpret to reflect wallrock contaminants entering the magma.

The resorbed cores of the plagioclase phenocrysts and the abrupt increase in An content across core/margin boundaries probably result from magma mixing, the higher anorthite content reflecting the fresh influx of more primitive magma into the dike. The Higganum dike is a multiple intrusion, and at most localities an inner dike is chilled against an outer one. In addition, less distinct features, such as slight concentrations of phenocrysts, gabbroic pegmatitic patches, and geochemical variations may indicate other pulses of magma that were separated by shorter time periods than those that produced the inner and outer dikes.

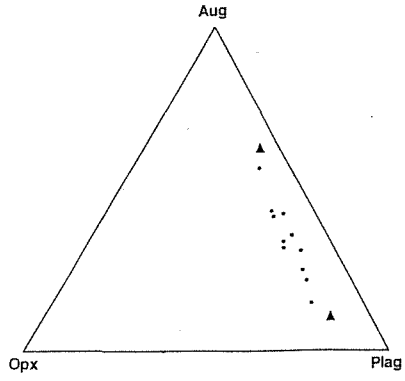


Figure 4. Variation in phenocryst abundances in chilled margins of the Higganum dike (after Koza, 1976).

Variations in proportions of phenocrystic minerals in the chilled margins are remarkably consistent over the 250-km length of the Higganum dike (Koza, 1976). Figure 4 shows that plagioclase and orthopyroxene phenocrysts in almost all samples are present in a proportion of 6/1. The content of augite phenocrysts, however, varies considerably from zero to 21 %. Koza (1976) suggested that this variation may indicate that augite is xenocrystic. We, however, believe that the augite is an integral part of this magma, and that its abundance is affected by the assimilation of crustal contaminants that convert augite to orthopyroxene and plagioclase (see discussion below).

Rocks formed from the first magma in the Hartford basin belong to Weigand and Ragland's (1970) eastern North American high-TiO_2, quartz normative type. Because of its relatively constant composition and wide geographic distribution, Puffer (personal communication) has suggested that this magma may be a primary one. Averages of analyses of the Talcott basalt, Fairhaven dike and Higganum dike are remarkably similar (Table 1). The inner and outer parts of the Higganum dike are also similar, with the inner dike being slightly more magnesian (Asher, 1988). When plotted in the pseudoternary diagrams Olivine-Plagioclase-Diopside (Fig. 5) of Walker et al. (1979) and Olivine-Augite-Quartz (Fig. 6) of Grove et al. (1983), analyses of the Higganum dike are tightly clustered along the olivine-augite-plagioclase cotectic. This is surprising in that olivine has never been found in the Higganum dike, despite its occurrence in the higher-level Fairhaven dike and Talcott basalt. Note that the compositional variation amongst the most mafic samples of the Higganum dike cannot be explained by variable amounts of the phenocrystic phases that it presently contains, that is, orthopyroxene, augite, and plagioclase. The conclusion that olivine must have been a primary phase on the liquidus at depths greater than the present exposures of the dike seems inescapable. Because plagioclase phenocrysts have resorbed cores and the augite phenocrysts are rounded, olivine in the dike at depth may have been completely resorbed when the magma underwent decompression during ascent. These relations require that olivine be replaced by orthopyroxene as a primary mineral on the liquidus at the intermediate depths at which the Higganum dike crystallized. It then became a primary liquidus mineral again at low pressures, forming euhedral crystals in the Fairhaven dike and Talcott basalt and mantling partially resorbed orthopyroxene crystals in the Talcott basalt.

Table 1. Analyses of rocks associated with the first magma of the Hartford basin. All totals recalculated to 100%. FeO/Fe_2O_3 taken to be 8.53 (Philpotts and Reichenbach, 1985).

	1	2	3	4	5	6	7	8	9	10	11	12
SiO_2	52.29	52.70	51.09	50.93	75.34	56.12	74.54	78.88	60.58	51.31	52.64	52.68
TiO_2	1.08	1.14	1.13	1.21	0.20	0.75	0.37	0.00	0.33	1.14	1.11	1.38
Al_2O_3	14.44	14.35	13.69	13.74	14.12	17.43	12.81	10.86	15.57	13.75	15.17	15.62
Fe_2O_3	1.05	1.09	1.17	1.21	0.23	0.83	0.33	0.07	0.65	1.17	1.40	1.93
FeO	8.94	9.27	9.97	10.29	1.95	7.06	2.77	0.62	5.57	10.01	8.59	8.87
MnO	0.16	0.18	0.19	0.19	0.03	0.10	0.08	0.00	0.15	0.19	0.17	0.16
MgO	8.07	8.01	8.20	7.58	1.75	4.86	0.97	0.00	6.91	8.24	7.42	5.79
CaO	11.37	10.79	11.68	12.08	1.51	9.61	5.75	0.95	6.82	11.74	11.09	10.71
Na_2O	2.08	3.09	2.24	2.15	3.02	2.33	2.21	0.88	2.83	2.25	2.00	2.29
K_2O	0.51	0.38	0.66	0.64	1.85	1.10	0.24	7.74	0.60	0.20	0.39	0.56

1. Talcott basalt average from Puffer et al. (1981)
2. Fairhaven diabase dike (Stop 2).
3. Average of 20 samples of Higganum diabase from the inner dike at Ponset (Stop 1).
4. Average of 11 samples of Higganum diabase from the outer dike at Ponset (Stop 1).
5. Macroscopic granophyric vein cutting chilled margin of Higganum diabase at Ponset (Stop 1).
6. Contaminated diabase from Higganum dike.
7. Low-K granophyre wisp stretched by flow of magma in margin of Higganum diabase dike.
8. High-K granophyre filling pressure shadow on end of phenocryst in margin of Higganum diabase dike.
9. Most contaminated coarse-grained diabase in contact with andesine-quartz-hornblende gneiss from Ponset.
10 Least contaminated diabase in Higganum dike at Willington Hill, CT.
11. West Rock diabase, on West Rock Ridge (from Schnabel, 1969).
12. Diabase from head of Sleeping Giant laccolith -- Stop 3 (from Schnabel, 1969).

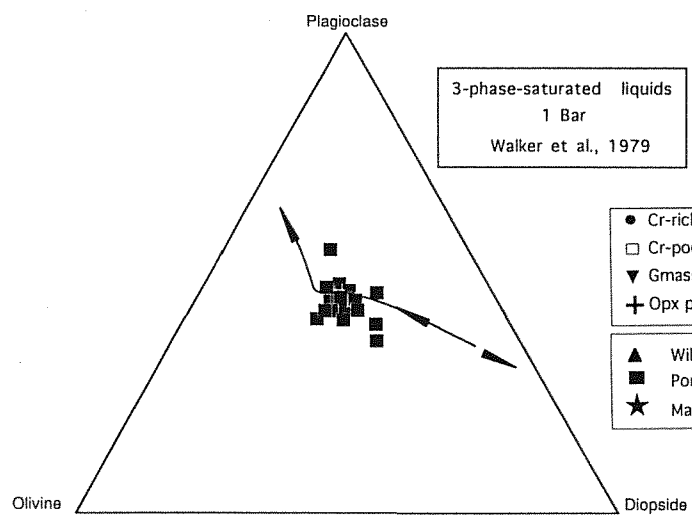


Figure 5. Analyses of samples from the Higganum dike plotted in terms of Olivine-Plagioclase-Diopside following the method of Walker et al. (1979).

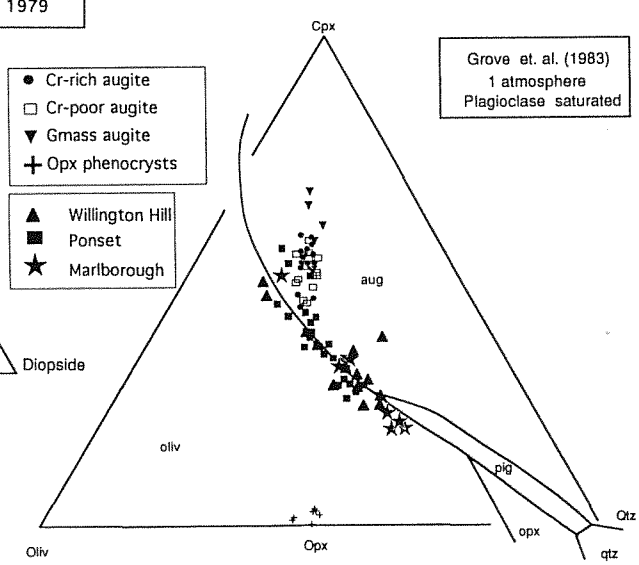


Figure 6. Analyses of samples from the Higganum dike plotted in terms of Olivine-Augite-Quartz following the method of Grove et al. (1983).

CRUSTAL CONTAMINATION OF THE MAGMA

The diabase in the Higganum dike shows clear evidence of crustal contamination where it comes in contact with xenoliths. The dike is typically about 50 m wide and consists of an inner dike chilled against and outer one, with the widths of the two parts being about the same. The contacts of the outer dike with country rock are typically marked by long sheet-like xenoliths or screens of wallrock ranging in thickness from centimeters to several meters (Fig.16); these appear to have spalled off the walls during the flow of magma and have undergone partial melting and some mixing with the diabase. The inner dike was intruded near the center of the first one and thus is sheathed by diabase; xenoliths of country rock have not been found in the inner dike, at least not at the present level of exposure. This part of the dike may, however, have been in contact with crustal rocks at deeper levels.

The evidence for partial melting of country rocks can be found at most contacts, especially where the wallrocks contain quartz and alkali feldspar. In thin section, narrow zones of granophyre are seen to separate the quartz and feldspar grains. The amount of granophyre is noticeably greater where biotite has decomposed and presumably provided water as a flux for melting. Some granophyre leaves grain boundaries and forms narrow veinlets that cut the surrounding grains (Fig. 7). These small veinlets may also cut the chilled margin of the diabase, where they commonly coalesce to form macroscopic centimeter-wide granophyric veins, which may extend several meters into the dike (Table 1-5). These rheomorphic veins formed after the mafic magma stopped flowing, and were probably injected by the pressure differential resulting from the volume expansion on melting of the country rock and the volume decrease on crystallization of the diabase. Examination of thin sections of the contact of the dike, however, reveals that some granophyre entered the dike at an earlier stage while the mafic magma was still flowing (Fig. 8). This granophyre forms wisps that were dragged in the direction of flow of magma in the dike. Where this occurred, the diabase was contaminated with a granitic fraction derived from the country rocks (Table 1-6).

Figure 7. Granophyre formed by partial fusion between orthoclase and quartz in wallrock of diabase. Small veins of granophyre extend out from boundaries into surrounding grains. Width of field 7 mm.

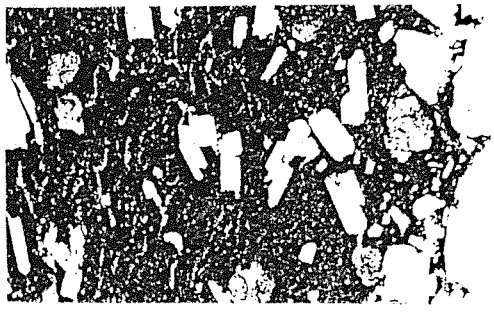


Figure 8. Wisps of granophyre entering the dike were streaked out in the direction of magma flow and then folded when the magma reversed its flow. Width of field 4 mm.

Figure 9. Fingerprint or fritted texture formed by partial melting of plagioclase. Width of field 2.5 mm.

Figure 10. Fine-grained zone of oligoclase developed between orthoclase and diabase. Believed to have formed by diffusional exchange between molten orthoclase and mafic magma. Width of field 4.5 mm.

Although most melting in the wallrocks occurred along quartz/feldspar grain boundaries to produce granophyre, feldspar crystals immediately in contact with mafic magma also melted. Plagioclase developed the characteristic fingerprint or fritted texture (Fig. 9). Orthoclase, on the other hand, reacted in a more complex manner. Most orthoclase is separated from diabase by a bulbous protuberance into the diabase of extremely fine-grained oligoclase (Fig. 10). The oligoclase is believed to have crystallized from a melt, because phenocrysts from the diabase are impressed into it, and the protuberances are commonly deformed by the flow of the mafic magma. Even though the protuberances are believed to have consisted of melt, the boundary between this melt and the mafic magma remained distinct, presumably because of slow rates of diffusion of silica and alumina across the boundary. Alkalies and calcium, however, were exchanged rapidly, potassium leaving the felsic melt and entering the diabase and sodium and calcium travelling in the opposite direction. The resulting reaction rims on orthoclase in contact with diabase are almost identical to those obtained in experiments by Watson (1982), who found that elements partition themselves between molten orthoclase and basalt in a manner similar to that between immiscible liquids.

The same type of element partitioning between melted orthoclase and mafic magma can be found between granophyric melts and mafic magma. Where the granophyres came in contact with molten diabase potassium was transferred to the mafic melt and sodium and calcium to the granophyric melt. The potassium, on entering the mafic magma was homogenized by diffusion and flow. Thus granophyric wisps that entered the diabase when it was still molten have low potassium contents (Table 1-7). In contrast, granophyres entering the dike after the diabase had solidified, are potassium rich (Table 1-8), contain sanidine, and have abundant biotite developed at their contacts with enclosing diabase.

Contamination of the magma in the dike with a purely granophyric fraction appears to be restricted to a narrow zone near the margins of the dike. Such contamination increases the silica content of the diabase. Diabase with elevated silica contents has been found only within a few centimeters of the contact. Contamination of the diabase with potassium that was exchanged between felsic and mafic melts at an early stage is, however, seen as a viable mechanism for contamination of the mafic magma in the main part of the dike.

Some idea of how much potassium may have been assimilated in this manner can be obtained from the potassium content of the zoned plagioclase crystals. The Fairhaven dike is unusual in that it still contains undevitrified Mesozoic glass. From the K_2O content of this glass and that of the included plagioclase phenocrysts we determine the plagiocalse/liquid partition coefficient for K_2O in this diabase to be 0.19, which is close to the average value of 0.17 given by Henderson (1982). The K_2O content of the core of many plagioclase phenocrysts in the Higganum dike is 0.16%, which indicates that the magma from which these crystals formed would have contained about 0.8% K_2O. This is significantly higher than the 0.6 % which is typical of both the inner and outer parts of this dike (Table 1-3 and 4)). This suggests that the core plagioclase crystallized from a magma that had been significantly contaminated with potassium. Immediately outside the core, however, the K_2O content of the plagioclase drops to 0.1%, which indicates that the magma would have contained only 0.5% K_2O at this time. If this reflects magma mixing, then the new batch of magma must have contained less than 0.5% K_2O. We have found some rapidly chilled diabase in the Higganum dike that contains as little as 0.2% K_2O. Analysis 10 in Table 1, we believe, represents the least contaminated magma.

The large amount of melting in the wallrocks of the Higganum dike is due to the great width of the dike and the nature of the contacts. Both the inner and outer parts of the dike are sufficiently wide that magmatic flow would probably have been turbulent (Huppert and Sparks, 1985). This would have caused considerable heat to be transferred to the wallrocks. In addition, the contact is usually not simple but is marked by numerous screens of wallrock into which heat could flow from both sides (Fig. 16). This further increased the chance for melting of the country rocks.

A numerical model of the temperatures generated in the wallrock screens at stop 1 on the excursion (Fig. 16) are shown in Figure 11. The calculations were done in two ways. In the first, diabase was emplaced instantaneously between the screens and then allowed to cool. In the second, the screens spalled off the wall of the dike sequentially, breaking off at the appropriate distance when the temperature at that point rose to 500°C. In both models, the initial temperature of the country rock is taken to be 300°C, the heat capacity of the magma to be 800 J kg^{-1} $^{o}C^{-1}$, the thermal diffusivity of magma and wallrocks to be 10^{-6} m^2 s^{-1}, and the crystallization of the magma releases a total of 400 kJ kg^{-1} linearly between 1200 and 1000°C. The heat effect of partial fusion of the screens is ignored because the amount of melt formed is relatively small.

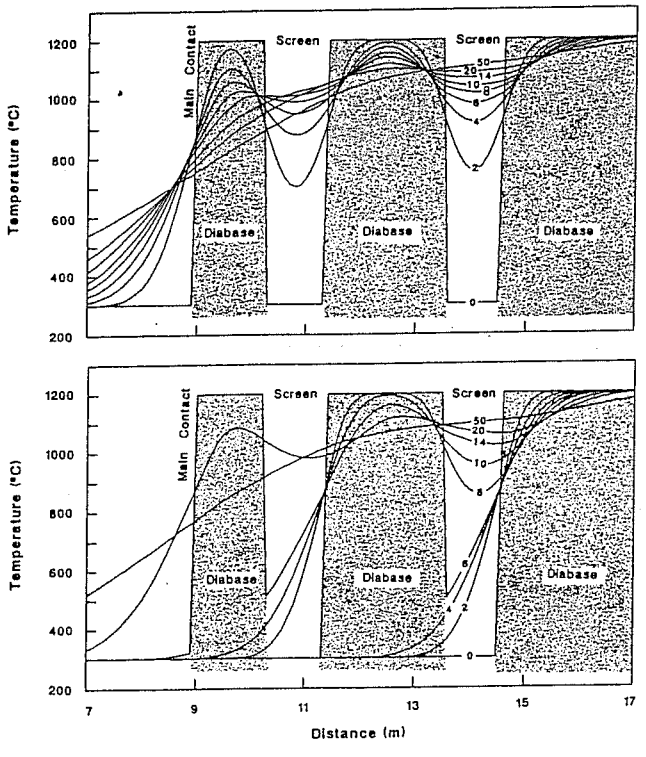


Figure 11. Calculated temperature distributions in screens at the margin of the Higganum dike, with time in days shown on the lines. In the upper diagram diabase was intruded simultaneously between the screens, whereas in the lower one it was intruded sequentially from right to left.

In both models, the temperatures in the screens rise rapidly to over 1000°C and remain there for 50 days before beginning to cool. This temperature is sufficient to bring about water-saturated melting in the screens. The actual temperatures in the screens would probably have been somewhat higher than calculated in the models and elevated temperatures would have been maintained longer, because of feeder flow in the dike. Abundant evidence of melt back of initial chilled margins clearly indicates that such flow occurred. The chilled margins against some xenoliths have been completely removed and in its place coarse-grained diabase has crystallized. These contacts exhibit the largest degree of partial melting in the country rocks and the greatest contamination of diabase (Table 1-9). In this contaminated diabase, phenocrystic and groundmass augite has been completely replaced and in its place Fe-rich orthopyroxene is present (Asher, 1991).

MAGMATIC FLOW DIRECTION IN THE FEEDER DIKE

In recent years there have been numerous claims that magmatic flow in regional diabase dikes may have been mainly horizontal rather than vertical, with the magma spreading laterally from central conduits. The problem with such claims is that there is usually very little direct field evidence for flow directions. The Higganum dike, however, provides many features, which are visible in thin sections that indicate the direction of flow, and even backflow.

The euhedral plate-like phenocrysts of plagioclase and the elongated prisms of orthopyroxene were oriented in an imbricate manner against the wallrock by the flow of magma (Fig. 12). In addition, the shear associated with the laminar flow of magma past phenocrysts near the margins of the dike just prior to complete solidification resulted in low pressure regions developing on opposing ends of phenocrysts. Late stage granophyric liquids entered these regions and now form asymmetric tails on the phenocrysts (Fig. 13), which indicate the sense of shear in the magma in the same way that pressure shadows on rotated porphyroblasts do in metamorphic rocks. Partial melts developed in the wallrocks were forcefully injected into the diabase, presumably by the volume expansion accompanying melting. When this occurred while the mafic magma was still flowing, the granophyric liquid was stretched into thin filaments in the direction of flow of the magma. In some cases these filaments were then folded back on themselves by the magma flow back down the dike (Fig. 8). Another flow feature seen in large cut surfaces are narrow zones of parallel-oriented phenocrysts that veer off from the contact into the dike. These resemble small thrust faults, and were probably formed by the magma being sheared along the contacts just prior to complete solidification.

Figure 12. Imbrication of plagioclase and orthopyroxene phenocrysts against the wall of the dike, indicating magma flow from right to left. Width of field 2 mm.

Figure 13. Granophyric segregations developed in low-pressure regions on opposing ends of phenocrysts in diabase indicate magma at top flowing from right to left. Width of field is 2 mm.

All of these features are seen clearly only in sections cut parallel to the direction of flow and perpendicular to the contact. Such directions can be found by trial and error, but it is quicker first to prepare a section parallel to the contact within the first millimeter of diabase, and look for granophyric wisps on opposing ends of phenocrysts. This defines the line of flow but not its sense. A section is then cut parallel to this and perpendicular to the contact to give the sense of shear.

The Higganum dike forms an enechelon set (Fig. 1), so it is unlikely that flow could have been lateral for any great distance, at least at the present level of exposure. We have made measurements of flow direction only at stop 1 of the excursion, where flow was upward and toward the south (Fig. 16). Because this locality is at the southern end of one of the enechelon segments, the flow direction may indicate that magma spread from the center of each enechelon segment. Without more measurements, however, this is pure speculation.

INTRUSIVE SHEETS

A number of mainly concordant sheet-like bodies occur in the Hartford basin (Fig. 1). All that have been analyzed have compositions that would relate them to the first magma of the basin (Talcott) rather than to the Holyoke or Hampden magmas. The largest of these bodies is the West Rock sill (Table 1-11), which was intruded near the base of the Mesozoic sedimentary sequence and is exposed along the western margin of the basin. It occupies a similar stratigraphic position to the Palisades sill in the Newark basin. At its southern end, West Rock sill is almost connected through the East Rock body and several smaller sheets and dikes with the Fairhaven dike, from which it is believed to have been fed. The Barndoor intrusion is another sheet-like body along the western side of the Hartford basin in northern Connecticut

Two laccolithic bodies rise from these basal sheets, the Sleeping Giant body coming from West Rock sill and Manitook Mountain forming at the northern end of the Barndoor intrusion. The sedimentary rocks in contact with these bodies show partial melting, and rheomorphic granophyric veins cut the marginal diabase. Although olivine phenocrysts can be found in chilled margins, the mineralogy of the coarser diabase in the interiors of the bodies consists predominantly of plagioclase, augite, pigeonite, and magnetite. The rock is rather homogeneous (Table 1-12), except for thin pegmatitic diabase sheets and veins that occur approximately two thirds of the way up through these bodies. These granophyres commonly contain long, curved pyroxene crystals that have a core of pigeonite and rim of augite.

EXTRUSIVE ROCKS

The first magma of the Hartford basin erupted to form the Talcott basalt, which is quite variable in its mode of occurrence. It does not form thick massive flows like the Holyoke and Hampden basalts do. Instead, it forms thinner sheets, many of which divide into individual lobes and even pillows, indicating that at least some magma erupted into a lake. In the southern part of the basin much of the Talcott is an agglomerate. Gas cavities are common in the Talcott rocks. Pipe vesicles are common near the base of massive flows and in pillows. Those in pillows are of particular interest because they grew inward from the chilled margin and thus are arranged approximately perpendicular to the cooling surface in a radial fashion. They show that gravity was not important in determining their orientation, thus casting doubt on the commonly-held belief that they are formed by gas rising into a flow (Philpotts and Lewis, 1987). Vesicle cylinders, which typically have dimensions similar to those of a broom stick handle and are composed of highly vesicular basalt, are seen near the base of massive parts of flows. They are usually oriented parallel to what initially would have been vertical, and are therefore probably formed by diapirs of vesicular basalt rising buoyantly into the flows, but only after flow had ceased (otherwise they would be tilted). The upper part of massive Talcott flows are characterized by large half-moon vesicles, which are up to 30 cm in diameter (Gray and Simmons, 1985). These large vesicles have a domed roof and a flat floor which may even be slightly convex upward. Their shape resembles that of a bullet or crescent moon (Fig. 14). They may have risen from the upper end of vesicle cylinders. Indeed many have a base of highly vesicular basalt similar to that seen in the vesicle cylinders. Because of their size, these vesicles would have risen with velocities up to 1 m/s, and Gray and Simmons (1985) interpret their streamlined shape as resulting from these high velocities. These vesicles are also commonly asymmetric, presumably as a result of flow (Fig. 14).

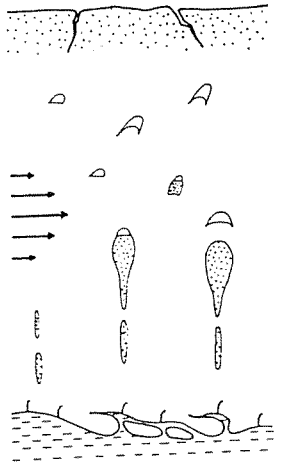


Figure 14. Schematic representation of the formation of half-moon vesicles and the interpretation of the flow direction based on their asymmetry (after Gray and Simmons, 1985).

The pillows, gas bubbles, agglomerate, and small size of individual flows render the Talcott basalt highly susceptible to alteration. Consequently, the original composition of this basalt is more difficult to determine than that of the other two basalts in the Hartford basin, which form thick massive flows. Some idea of the possible severity of the chemical alteration is given by the series of analyses in Table 2, which are of samples taken across a pillow of Talcott basalt from Tariffville (Stop 4). The most severe alteration is of the glassy rim, which is converted to palagonite. Its low silica and high MgO and FeO contents could be mistakenly interpreted as indicating an ultramafic composition for the Talcott. Even inside the glassy selvage, the crystalline basalt shows loss of silica and enrichment in sodium for several centimeters in from the rim. The average analysis of the Talcott basalt given in Table 1-1 from Puffer et al. (1981) was based on carefully selected samples and is believed to be relatively free from the effects of alteration.

Table 2. Analyses across a pillow of Talcott basalt from Tariffville, Connecticut. Distances are measured from the glassy selvage (devitrified) to the core of the pillow (13).

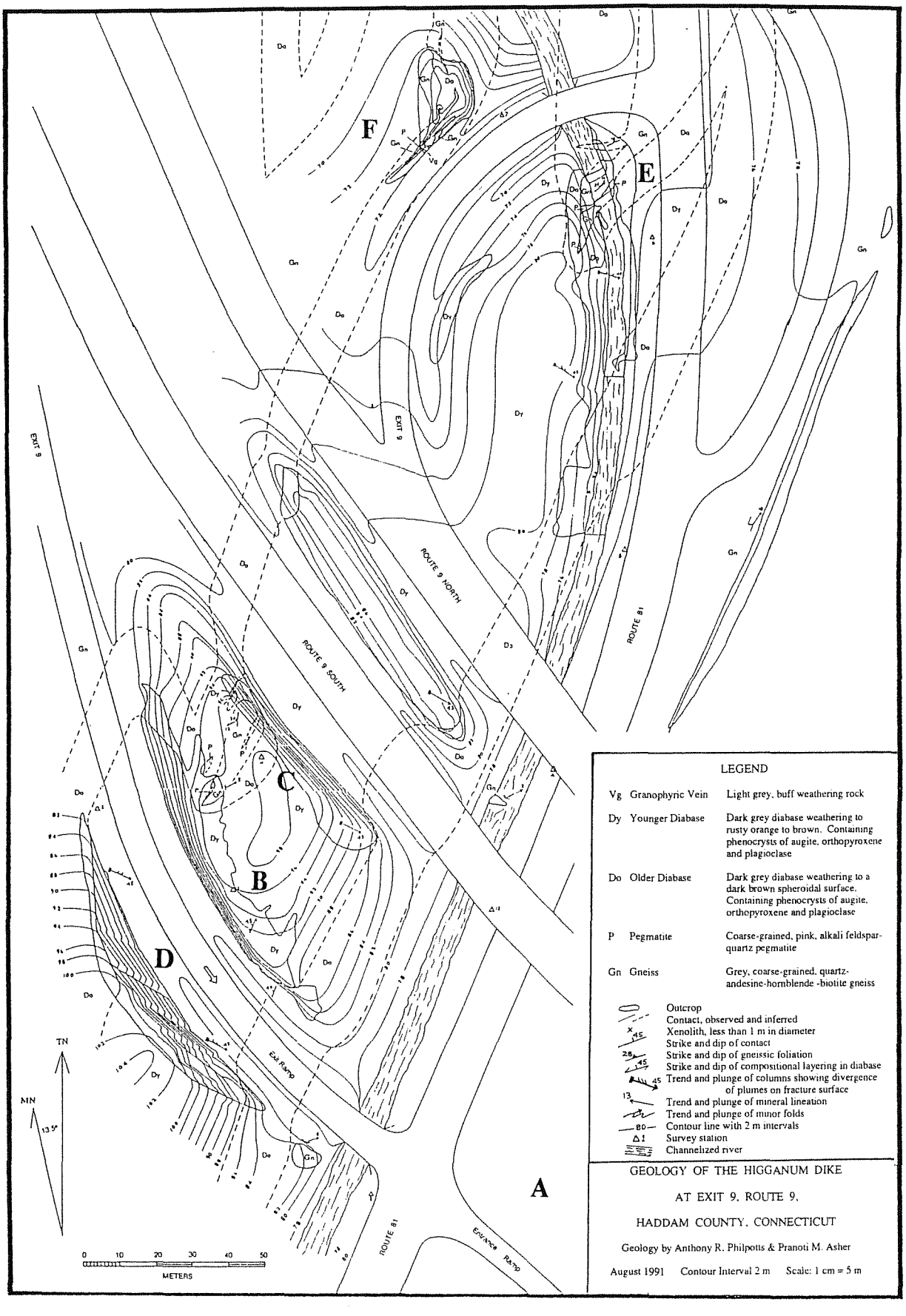
Distance from rim (cm)	0	2	4	7	10	13
SiO_2	39.67	45.36	48.21	49.17	51.71	51.16
TiO_2	1.50	1.41	1.40	1.27	1.31	1.29
Al_2O_3	14.44	15.32	15.41	14.26	15.05	13.47
Fe_2O_3	8.37	5.14	4.51	4.14	5.26	4.49
FeO	11.58	7.11	6.24	5.73	7.28	6.21
MgO	20.16	8.20	7.30	10.39	4.14	8.97
CaO	2.21	12.96	12.57	11.23	11.59	11.51
Na_2O	1.40	4.06	3.91	3.23	2.97	2.02
K_2O	0.67	0.42	0.46	0.58	0.68	0.87

Although the devitrified glassy selvages of pillows clearly indicate the Talcott basalt contained olivine phenocrysts, fresh olivine has never been found. It is always replaced by calcite, serpentine, clay, or even quartz. As mentioned above, this basalt does contain rare orthopyroxene phenocrysts that are mantled by olivine, which are interpreted to have come from depth in the feeder dike. Augite, which commonly forms rounded glomeroporphyritic aggregates, and plagioclase are the other phenocrystic phases. The groundmass is composed of plagioclase, augite, pigeonite, and a mesostasis consisting of magnetite dendrites and small iron-rich spheres in devitrified silica-rich glass, the latter two being the products of late-stage liquid immiscibility.

Melting experiments indicate that the eruption temperatures of the three basalts in the Hartford basin dropped progressively from a high in the Talcott of 1178°C to 1160°C in the Holyoke and 1148°C in the Hampden (Philpotts and Reichenbach, 1985). The falling temperature could be interpreted as resulting from the cooling and fractionation of a magma in a deep-seated reservoir. However, the length of time between successive eruptions, based on Milankovitch cycles in the intervening sedimentary rocks, is too short for the removal of the amount of heat necessary to cool and fractionate (heat of crystallization) a basaltic magma. The decrease in temperature between successive basalt units may instead be caused by the partial melting of an adiabatically rising source region beneath the Hartford basin (Philpotts, 1991).

ROAD LOG

The excursion starts at Exit 9 on Route 9 (Killingworth, Clinton, Route 81 Exit), 9 miles south of Middletown, CT. Park on grassy shoulder on east side of Route 81 immediately to the east of Route 9. This is the first stop, where we will spend about two hours.
Figure 15 (next page) is a detailed map of this interchange with bold letters indicating the specific localities described below.


LEGEND
Vg Granophyric Vein — Light grey, buff weathering rock
Dy Younger Diabase — Dark grey diabase weathering to rusty orange to brown. Containing phenocrysts of augite, orthopyroxene and plagioclase
Do Older Diabase — Dark grey diabase weathering to a dark brown spheroidal surface. Containing phenocrysts of augite, orthopyroxene and plagioclase
P Pegmatite — Coarse-grained, pink, alkali feldspar-quartz pegmatite
Gn Gneiss — Grey, coarse-grained, quartz-andesine-hornblende -biotite gneiss
Outcrop
Contact, observed and inferred
Xenolith, less than 1 m in diameter
Strike and dip of contact
Strike and dip of gneissic foliation
Strike and dip of compositional layering in diabase
Trend and plunge of columns showing divergence of plumes on fracture surface
Trend and plunge of mineral lineation
Trend and plunge of minor folds
Contour line with 2 m intervals
Survey station
Channelized river
GEOLOGY OF THE HIGGANUM DIKE
AT EXIT 9, ROUTE 9,
HADDAM COUNTY, CONNECTICUT
Geology by Anthony R. Philpotts & Pranoti M. Asher
August 1991 Contour Interval 2 m Scale: 1 cm = 5 m
A
B
C
D
E
F
EXIT 9
ROUTE 9 NORTH
ROUTE 9 SOUTH
ROUTE 81
EXIT RAMP
ENTRANCE RAMP
TN
MN
13.5°
0 10 20 30 40 50
METERS

WARNING! The diabase and basalt we will see on this excursion are notorious for producing high speed projectiles when hit with a hammer. One of us (A.R.P.) managed to remove one of his teeth with such a piece while sampling the diabase at the first stop. Please be careful and mindful of others while collecting samples. At the second stop, we will see what surely must rank as one of the world's most lush growths of poison ivy -- you have been warned.

STOP 1. HIGGANUM DIABASE DIKE (<2 hours)

We will begin by examining the country rocks into which the Higganum dike was intruded at this locality. These are well exposed in roadcuts on south-bound Route 9 immediately south of the interchange. They are best viewed looking east from the rock-divider separating Route 9 from the south-bound entrance ramp (Fig. 15 **A**). A short trail climbs this divider from the northeast corner of the intersection of Route 81 and the south-bound entrance ramp. Ordovician meta-volcanics of the Middletown formation, which consist of leucocratic quartz-andesine-hornblende gneisses and some dark amphibolite layers have been deformed into folds that are overturned toward the south (makes for a great photograph). These gneisses are cut by granitic pegmatites, which are common in the Middletown area.

On returning to Route 81, we have an excellent view of the Higganum dike in the south-bound exit ramp from Route 9. Almost the entire 70-m width of the dike is visible. Prominent columnar joints indicate that the dike dips 45° toward the northwest. Such a shallow dip is unusual for the dike. Along most of its length the dip is near vertical. The shallow dip of the dike at this locality is a primary feature and is not due to later rotation, as indicated by the asymmetric development of features that depend on gravity for their formation (vesicles along only one contact, for example).

Proceed across Route 81 and climb onto the top of the roadcut through the dike on the northeast side of the south-bound exit ramp (Fig 15 **B**). Be careful of traffic on the ramp. From this point the entire width of the dike is clearly seen on the west side of the exit ramp. The most striking feature is the columnar jointing, which forms columns over a meter in diameter that plunge to the southeast at 45°. These columns are developed only in an inner dike, which weathers to a distinctive rusty red. This inner dike is chilled against an outer dike, which weathers to a browner color and typically forms crumbly spheroidal bodies between intersecting joint sets. Joints parallel to the contact are more prominent than columnar ones in this outer dike. As shown on the map (Fig. 15), the contacts between the inner and outer dikes can be traced through all of the exposures of the dike at the Route 81/9 interchange. With appropriate lighting, plumose structures up to 4 m long can be seen on the columnar joints, indicating that fractures propagated inward from both contacts (these are shown on the map, although you may require a magnifying glass to see them). Such long fracture-steps and the large diameter of the columns indicate that the dike, at this exposure, must have been at considerable depth when it cooled. The depth is estimated to have been approximately 10 km based on the geometry of the Higganum and Fairhaven dikes and the dip of the eastern border fault of the Hartford basin, as illustrated in Figure 2 (see Philpotts and Martello, 1986, for discussion).

At **B** you are standing near the center of the inner dike. From here to the lower contact with the older diabase, faint centimeter-scale rhythmic layering is produced by variations in the concentration of plagioclase and pyroxene. Best examples of this are seen, unfortunately, by peering over the edge of the roadcut in the first 6 m north from station 1 (be careful). In walking northwestward from B we approach the top of the inner dike and the chilled margin with the older diabase, which has a distinctly more crumbly weathered surface. In the upper part of the inner dike numerous coarser-grained pegmatitic diabase segregations are seen to have accumulated in what must have been irregularities in the downward solidifying roof of the inner dike.

As soon as the contact with the older diabase is crossed a large xenolith of gneiss and pegmatite is encountered. No xenoliths are found in the inner dike. The contacts between the diabase and pegmatite are extremely irregular, with small protuberances and veins of felsic rock cutting the chilled margin of the diabase. Apart from the phenocrysts, the diabase at this contact is aphanitic. Nonetheless, the pegmatite has undergone partial melting and backveins the diabase.

Proceed across the hilltop to point **C** for an excellent view of numerous roadcuts through the dike all the way across to the upper contact at point F in the north-bound exit ramp from Route 9. The outcrop in the median divider of Route 9 has no xenoliths in it whatsoever. In contrast, at point C large xenoliths of gneiss and pegmatite are encountered in the older diabase by walking to the northwest. The contact between diabase and pegmatite seen at this point by climbing down the face of the roadcut toward the highway is quite different from that seen at the previous exposure in that the diabase is extremely coarse grained. In addition it is highly contaminated and has a pink color due to the presence of about 50% granophyre, which was derived from the partial fusion of the pegmatite. Any chilled margin that developed on the lower side of this xenolith has been completely melted back. The contact of the outer dike with the country rocks is not quite exposed to the northwest of point C, but excellent samples of the chilled porphyritic diabases against gneissic xenoliths can be obtained here.

Return to point **B** and note that no xenoliths occur in the dike on the southwest side of the exit ramp. The presence of xenoliths between points B and C can be seen on the map (Fig. 15) to correspond to the place where the dike takes a small jog. The large xenoliths at points E and F are located at a still larger jog in the trend of the dike. These probably correspond to points where enechelon segments came together during dilation of the dike.

Descend the path at the northwest end of the roadcut and carefully cross the exit ramp to the southwest side (**D**). While walking back to Route 81, look for the contacts between the inner and outer dikes, calcite-filled vesicles near the upper contact of the internal dike, rhythmic cross-bedded layering in the center of the dike, and plumes on the fracture surfaces of the columnar joints indicating the direction of propagation of fractures. Just before reaching Route 81 is a small outcrop of gneiss near the concealed contact with the diabase.

Walk northward along the northwest side of Route 81. The river along the side of the road cuts obliquely across the contact of the dike. Between the south- and north-bound lanes of Route 9 gneiss is exposed in the river bed, but on the east side of the north-bound lane diabase appears. The actual contact with the gneiss is not exposed, but the diabase at this point is aphanitic and contains several small xenoliths of gneiss. On continuing down the road, the contact of the inner diabase with the outer one cuts obliquely across the river. The rusty color and prominent columnar joints of the inner diabase are clearly evident. This inner dike is actually a branch off the main dike along the top of the bank on the west side of the river.

Continue to point **E**. On the west bank of the river is a very large xenolith of gneiss surrounded by the older diabase. The xenolith also contains two pink granitic pegmatites. Along the upper side of the xenolith, diabase shows no decrease in grain size toward the contact whatsoever, whereas along the lower side of the xenolith, which is exposed on the east bank of the river, diabase is noticeably finer-grained against the xenolith. The lack of a chilled margin on the upper side of the xenolith indicates that meltback of the chilled margin must have occurred. The gneiss and pegmatite along the upper side of the xenolith show extensive melting. The andesine in the gneiss has a fingerprint texture (Fig. 9) and quartz-andesine granophyre is present on grain boundaries in the gneiss. Extensive melting occurred in the pegmatite, with the development of granophyre along grain boundaries (Fig. 7). The coarse-grained diabase in contact with the gneiss was contaminated with partial melts from the gneiss. Because these melts were quartz-andesine granophyres they raised the silica content of the diabase but not its potassium content (Table 1-9). The augite in the contaminated diabase was completely replaced by Fe-rich orthopyroxene. On cut surfaces of the contact, faint finger-like protrusions of felsic concentrations can be seen extending out from the contact into the diabase. The volume expansion on melting in the gneiss may have forced these fingers into the mafic magma. Buoyancy of the lighter felsic melt could also have been a factor.

Cross onto the north side of the north-bound exit ramp (Fig. 15 **F**) where a long roadcut exposes a number of sheet-like xenoliths of gneiss near the western contact of the dike (Fig. 16). The actual contact with the country rock is exposed on the back side of this outcrop. This outcrop provides an excellent exposure of the typical appearance of the contact of the Higganum dike, which is commonly marked by thin screens of country rock.

Three main screens are seen extending along most of the length (30 m) of the outcrop, yet the screens are only centimeters thick in places. The orientation of the gneissic foliation in each screen is similar to that in the country rock, indicating that the screens may still be attached to the wallrock in the third dimension. The diabase between the screens is relatively fine-grained and becomes aphanitic against the two screens closest to the contact. Diabase can also be found quenched against diabase, indicating that there must have been several pulses of magma.

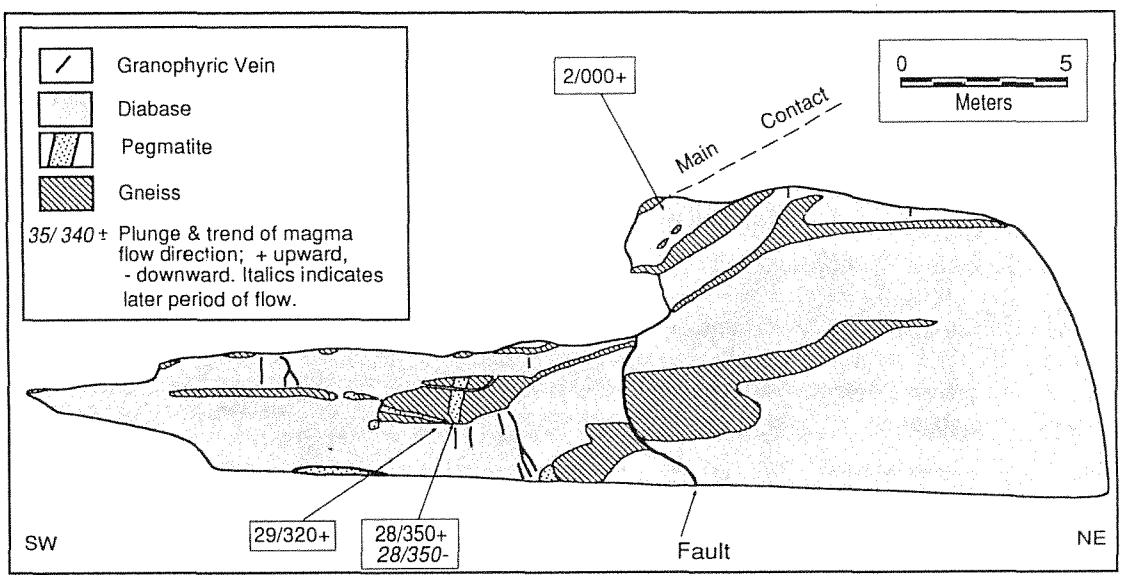


Figure 16. Schematic vertical section of roadcut on north-bound exit ramp from Route 9 (Locality F, Fig. 15)

The diabase between the screens is cut by a number of centimeter-wide granophyric veins, which are oriented approximately perpendicular to the contacts with the screens. The granophyre is pale green but weathers to a cream color. It is fine-grained, and is composed of alkali feldspar and quartz with minor ankerite, muscovite, ilmenite, chlorite, sericite, and pyrite. The alkali feldspar is partially kaolinized, except in 2-mm thick zones along the margins of the veins and around coarse patches in the central part of the veins where perfectly clear crystals of almost pure albite ($An_{0.2}Ab_{99.2}Or_{0.6}$) and quartz surround patches of ankerite and chlorite. The carbonate weathers out to leave pits on the surface. The granophyric veins thin as they approach the screens and appear to terminate within a centimeter of the contact in the chilled diabase. In thin section, however, the macroscopic granophyric veins can be seen to branch into a number of thin veinlets that traverse the chilled margin of the diabase and connect with granophyre developed on quartz/feldspar grain boundaries in the screens. As indicated in the thermal models of this outcrop (Fig. 11) the gneiss in the screens would have been undergoing melting while the diabase was solidified. The associated volume changes would have caused the granophyre to be forcefully injected into the diabase. This injection, however, must have occurred after the diabase had ceased moving, because the granophyric veins show no signs of deformation.

Earlier formed granophyre was injected into the dike while the magma was still moving, but it was streaked out into thin filaments in the direction of flow. These filaments are visible only in thin section but have been found in the fine-grained diabase against most of the screens seen in this outcrop. Because this earlier granophyre coexisted with a mafic liquid it lost potassium to the mafic liquid and gained sodium (Table 1-7).

The flow direction of magma in the dike, as indicated by stretched granophyric wisps (Fig. 8), imbrications of phenocrysts (Fig. 12), and asymmetric granophyric segregations on the ends of phenocrysts (Fig. 13), is shown in Figure 16. Flow was mainly up the dip of the dike, with a lateral component toward the south. As mentioned above, this flow direction may indicate that magma was spreading toward the southern end of the particular enechelon segment in which it was rising. The one measured direction of backflow was found to be the opposite of the direction of primary flow.

Return to vehicles.

Mileage (in miles). Time to next stop 45 minutes.

0.0 Road log begins at the intersection of the north-bound exit ramp from Route 9 and Route 81. Proceed north on Route 81 to Higganum.

1.7 Reach T-junction at stop light. Turn left on Route 154 and then immediately turn left again by bronze war memorial plaque onto Candlewood Hill Road.

3.0 CAUTION -- an unexpected stop sign!

5.0 At stop sign, turn right on Foothills Road.

7.9 Take sharp left at base of hill onto Johnson Road. At this point you are following the eastern border fault of the Hartford basin.

9.75 At stop sign, bear left on Maiden Lane. Continue on this road, passing several stop signs, into Durham.

11.0 Junction with Route 17. Turn left onto Route 17 south.

12.1 Hill to left of road exposes fanglomerates that contain boulders of Talcott basalt, indicating that the first lava in the Hartford basin extended east of the present border fault.

17.9 Citgo Gas station on right -- if needed.

21.8 Small shopping plaza on right. Get prepared to turn left onto Warner Road.

22.5 Turn left onto Warner Road. Intersection and name are not obvious from any distance.

23.5 At stop sign, proceed across intersection taking slight jog to left.

23.8 Stop 2 is in small quarry on east side of road opposite house (# 214, Warner Road). Because of lack of parking space continue south on Warner Road.

24.0 At T-junction, turn right and park along side of Barberry Road opposite the stables. Return up Warner Road to small quarry on foot.

STOP 2. FAIRHAVEN DIKE, AND TALCOTT AGGLOMERATE AND PILLOW BASALT (1 hour)

**********BEWARE OF THE POISON IVY******

At this stop, the Fairhaven dike is seen where it reached the Mesozoic land surface and erupted to form an agglomeratic phase of the Talcott basalt, some of which flowed back into the center of the dike. Immediately above the agglomerate, the Talcott becomes pillowed. The Fairhaven dike is exposed in the cliff at the base of the steep hill on the east side of Warner Road. Immediately above the first cliff the dike passes up into the agglomerate and then into the pillowed basalt. The following describes the rocks encountered while climbing the hill. This hill is steep and in places has very active talus -- be careful.

The small cliff facing Warner Road is essentially the western contact of the Fairhaven dike, which dips vertically and strikes parallel to Warner Road. At the north end of the cliff, arkosic sediment can still be seen attached to the wall of the dike. Wisps of this sediment in the margins of the dike have been partially melted. Large blocks in the central part of the cliff have fallen out along columnar joints exposing several meters of the width of the dike. These columnar joints are remarkable in that a late injection of magma was intruded along them and froze to a glass, much of which has not yet been devitrified. The glass at one point forms a skin that is draped across the columns. **Please do not hammer this surface -- samples of fresh glass can be found below it and in the loose blocks at the base of the cliff.** Analyses of the glass are given in Philpotts and Martello (1986). This glass also contains completely unaltered euhedral olivine phenocrysts. This is the only locality the authors are aware of in which fresh olivine has been preserved in rocks formed during the first period of igneous activity in the Hartford basin. Rounded and resorbed phenocrysts of orthopyroxene are also present as are euhedral plagiocalse phenocrysts and rounded glomeroporphyritic aggregates of augite.

On climbing around the southern end of the cliff face, we pass into the dike. It has many internal contacts and was clearly the conduit for repeated surges of magma. One of these surges cooled and formed a striking set of small columnar joints, which are approximately horizontal and have diameters in the centimeter range. These are so much smaller than those seen at Stop 1 in the Higganum dike, and they were clearly formed at a much shallower depth. This depth is estimated to be only a few meters beneath the surface, because the columns terminate toward the center of the dike against a breccia, which we interpret to be a flowback breccia. Immediately above this columnar-jointed zone, the dike passes up into a thick agglomerate, which is cut by several small meter-wide vesicular basaltic dikes. Eventually at the top of the hill the agglomerate changes into pillowed basalt, which forms a long outcrop extending southward along the east side of Warner Road. Imbrication of the pillows suggests flow toward the south. Return southward along this outcrop to the vehicles.

Mileage Time to next stop 20 minutes.

0.0	Set odometer to 0. Continue on Barberry Road.
0.2	Bear left on Thompson Street.
0.4	Immediately turn right onto Arrowdale Road.
0.9	Turn right at T-junction onto North Hill Road.
1.1	Take first left onto Beach Lane.
1.5	Turn right at T-junction onto a street with no name.
1.6	Turn left onto Montowese Avenue.
1.75	Cross Route 17 (Middletown Av.) and continue on Montowese Av.. This is a busy intersection -- be extremely careful.
2.7	Turn right onto entrance ramp to I-91 north. Once on I-91, the profile of the Sleeping Giant (head to the left) is visible on the horizon. The head of the giant is separated from the body by a deep valley eroded along a fault.
4.1	Take Exit 10 off I-91 for Mount Carmel, Route 40, Hampden, Cheshire. Stay on this ramp, which becomes Route 40 north.
6.5	Roadcuts through New Haven arkose with green caliche horizons.
7.3	At end of expressway turn right onto Route 10 north.
8.65	At traffic light, turn right onto Mt. Carmel Road. You will see a sign for Sleeping Giant State Park.
9.0	Turn left into Sleeping Giant State Park (opposite Quinnepiac College), and follow circular road to large parking area near toilets.

STOP 3. SLEEPING GIANT LACCOLITH (Lunch stop + 1 hour)

The purpose of this stop, apart from eating a picnic lunch in pleasant surroundings, is to examine the western end of the Sleeping Giant laccolith. This body was mapped by Frittz (1963), who showed that it had a general laccolithic form, although few contacts with the intruded New Haven arkose are exposed. Large columnar joints indicate that the body has a general domical form. We will visit the western end of the intrusion (the head of the Giant), where a large quarry provides an excellent exposure through a considerable vertical section of the body and its contact with the country rocks.

Follow the blue dot trail from the northern side of the parking area around the head of the Giant to where a quarry provides a frontal lobotomy into the laccolith. Just prior to arriving at the quarry the trail follows a high ridge of diabase, which occupies what is probably the feeder to the laccolith, the magma having risen from the underlying West Rock sill, which forms the major hill to the west. From this ridge one has a good view of the quarry. It is bounded on its western side by a low wall in which the New Haven arkose is in contact with the diabase of the laccolith. Toward the east, the quarry walls rise steeply, following what is essentially the roof of the intrusion. No sedimentary rocks remain on the roof, but xenoliths and rheomorphic granophyric veins in fine-grained diabase are exposed along the blue dot trail where it climbs the forehead of the giant. One large granophyric vein, in fact, can be seen extending the length of the ridge on which we are standing.

At the end of the diabase ridge, the blue dot trail meets the red diamond trail. Follow the red trail into the quarry and proceed to the talus slope on its northwestern side, at the top of which the contact of the diabase with the New Haven arkose is exposed. Columnar joints in the diabase are perpendicular to the contact, which is conformable with the bedding in the sediments. The arkose here has been baked to a hornfels in which there is minor partial melting. Fractures in the hornfels cut across pebbles rather than going around them as they typically do in the unmetamorphosed arkose. Cross to the eastern side of the quarry where large blocks have fallen from the main quarry face. The diabase is remarkably constant in composition and texture throughout most of the thickness of the body. It is composed essentially of plagioclase, augite, and pigeonite, but olivine crystals are present in chilled margin samples. Analyses show that its composition (Table 1-12) corresponds to that of the first magma in the Hartford basin. The only significantly different mafic rock in the quarry is a pegmatitic phase of diabase, which forms thin sheets approximately two thirds of the way up through the body. These sheets commonly contain centimeter-long curved pyroxene crystals, which have cores of pigeonite and rims of augite. This pegmatitic diabase also contains a slightly higher percentage of interstitial granophyre than does the normal diabase. Examples of this pegmatitic diabase can be found in blocks that have fallen to the quarry floor. Granophyric veins from the upper contact are also found in these blocks.

Return to the vehicles by the red diamond trail.

Mileage Time to next stop 1 hour.

0	Set odometer to 0 as you exit the Park. Follow Mount Carmel Road back to Route 10.
0.3	Turn right onto Route 10 north.
4.6	McDonald's to the right and Dunkin Donuts to the left, if you must.
10.1	Take exit from Route 10 onto I-691 west (toward Waterbury). Roadcut on south side of highway is the northern end of the West Rock sill, which fed the Sleeping Giant laccolith.
11.2	Take exit 2 to I-84 east (toward Hartford) passing New Haven arkose on right-hand side of road.
21.0	Large quarries on north and south sides of I-84 are in Holyoke basalt underlain by sandstones of the Shuttle Meadow member of the Meriden formation.
21.7	Opposite Exit 36, note contact between upper and lower Holyoke basalt. The upper flow is characterized by splintery joints and is a rusty yellow color compared with the darker brown of the lower unit.
25.3	Leave I-84 from the left lane at Exit 39 (Route. 4, Farmington).
26.3	Continue across traffic light on Route 4. On the left side of the road is Holyoke basalt, which is faulted in contact with pillowed Talcott basalt, which is exposed on the right side of the road. Beneath the Talcott, red New Haven arkose and shales are exposed and one white bed of sandstone that shows green-weathering copper mineralization.
27.2	Junction with Route 10. Turn right onto Route 10 north.
32.9	Traffic light at junction with Route 44. Turn left on Route 44.
33.65	Turn right onto Route 202, which is also Route 10 north.
34.6	View to the right of Talcott Mountain and the Hublein Tower.
35.7	King Phillip's Cave, visible at north end of escarpment to the right, may have eroded out a filled lava tube in the Talcott flow.
40.4	Turn right onto Route 315 toward Tariffville and bear left after crossing the bridge over the Farmington River.
42.0	Turn right at stop sign on Route 315.
42.5	Turn left at junction with Route 189, then turn immediately to the right, and park on Tunxis Road or in the parking lot of the old factory.

STOP 4. SECTION THROUGH THE TALCOTT BASALT ALONG THE FARMINGTON RIVER GORGE AT TARIFFVILLE (1 hour)

The purpose of this stop is to examine a complete section through the Talcott basalt, which is approximately 30 m thick in the Tariffville gorge. This locality has been described in detail by Gray (1987) in the Geological Society of America Centenial Field Guide, from which the following notes have been taken. The locality is of importance because here the Farmington River has cut a deep gorge through all three lavas of the Hartford basin. Exposures of their upper contacts allowed Rice (1886) and Davis (1898) to prove that these trap sheets were of extrusive rather than intrusive origin.

We will examine the lower part of the Talcott basalt in the road cut along the north side of Route 189, proceeding east from the vehicles, and then descend to the Farmington River to see the upper part of the basalt. The long roadcut begins with an exposure of several meters of red New Haven arkose and shale beneath the Talcott basalt. The sediments were not metamorphosed by the basalt, which is pillowed over the first 1.5 meters. Some sediment was forced up between the pillows. Above the pillows, the basalt is massive and has columnar joints. Near the base of the massive part, some vesicle cylinders are present. In the upper half of the massive flow vesicles are present, many of which have a characteristic half-moon shape, which is slightly deformed by flow to the east. The vesicles are filled with calcite, quartz, prehnite, datolite, and gash-like crystal cavities indicate the former presence of anhydrite. Half way along the roadcut a 0.5-m-wide vein of breccia is present, in which fragments of basalt are surrounded by datolite and arkosic sediment, which indicates that the vein must have connected with the sediment above or below the flow. The sediment in the vein lacks quartz, which must have been consumed during the formation of the datolite. Extremely small euhedral grossular garnets are present in the datolite. The source of all the boron necessary to form datolite is uncertain. At about 15 m the basalt becomes highly vesicular and is covered by a flowtop breccia. This terminates the exposure in the outcrop on the north side of Route 189.

Continue along the side of the road to the end of the guard wires and then descend to the Farmington river, where another 15 m of thin, interfingering flow lobes form an upper part to the Talcott basalt. Erosion by the river has clearly exposed the lobes. Pipe vesicles are extremely common in the rims of the lobes. They are oriented perpendicular to cooling surfaces rather than only extending vertically upward, suggesting that buoyancy is not critical in their formation (Philpotts and Lewis, 1987). The top of the flow is marked by an agglomerate mixed with red sediments.

The Shuttle Meadow sediments overlying the Talcott consist of interlayered red sandstones and shales. These continue upward for 17 m before being overlain by the Holyoke basalt. Many contain calcite-filled evaporite crystal casts up to 5 cm in size. The form of many of these suggests that the original mineral was glauberite (Gray, 1987). The upper part of the sedimentary sequence consists of finely laminated soda-rich mudstones, which Gray (1987) interprets may have formed from soda rich pore fluid reacting with clay-rich sediment during thermal metamorphism caused by the the overlying Holyoke basalt.

Climb back up to Route 189 and return to vehicles.

To get to I-91, follow Route 189 south for 2.2 miles; then take Route 187 south for another 5.5 miles, until reaching a sign indicating I-91 to the left. Turn left and after 1.6 miles you will be at I-91. From here you can plot your own course. Amherst is to the north.

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GROUND WATER AND SOIL VAPOR INVESTIGATION AND REMEDIATION TECHNIQUES

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BACKGROUND

Petroleum "discharges" or "releases" associated with underground storage tanks (UST) are one of the most common causes of environmental contamination in New England. In 1988, the EPA estimated that up to 25% of all UST were leaking. The authors' experience with thousands of tanks suggests that the actual number of leaking tanks is far fewer; however, many structurally sound underground storage systems have significant environmental contamination associated with them due to careless historical product handling procedures (e.g. tank overfills, spills during tank filling, etc.). In RCRA Subtitle I, enacted in 1988, the EPA required that underground petroleum storage systems be upgraded to meet specific design criteria, required that tank testing be performed on a regular basis (with the testing schedule a function of tank age and construction), and required that tanks that do not meet the design criteria be "closed" or taken out of service by 1998. All tanks undergoing closure require an evaluation of environmental impact. (Inclusion of these regulations in RCRA represented the first significant regulation of petroleum as a hazardous substance under CERCLA/RCRA. Petroleum had previously been specifically exempt from these regulations.) Many states had parallel programs in place when Subtitle I was enacted, and other states have since enacted their own versions of underground storage tank programs. All meet the basic criteria of Subtitle I but many state programs have their own idiosyncrasies.

The huge number of underground storage tanks in the ground, coupled with the existence of regulations requiring simultaneous tank upgrades and environmental evaluation, have resulted in the discovery of a large number of sites contaminated with petroleum. By far the majority of "disposal [contaminated] sites" listed by the Massachusetts DEP are petroleum-contaminated sites.

Petroleum is a complex mixture of hundreds of compounds. Benzene, toluene, ethyl benzene and xylenes (BTEX), which are aromatic constituents of gasoline, and, to a lesser extent, fuel oils, are generally used as indicator compounds to assess petroleum contamination. Benzene is a carcinogen and therefore has a stringent drinking water standard of 5 parts per billion (ppb). The presence of benzene in ground water in concentrations significantly in excess of 5 ppb is often the critical factor which forces site remediation. The risk presented by the release to human or environmental receptors may also be factored into the decision to remediate.

REMEDIATION

When petroleum is released to the subsurface, the majority of the petroleum sorbs to sediment grains. Some dissolves in ground water, some volatilizes to occupy the pore spaces in the vadose zone, and, if enough petroleum is released, some pools on the surface of the water table to form a floating layer of "free product". Transport of contaminants, required to affect sensitive receptors, generally occurs through flow in the dissolved phase, and, to a lesser extent, the free product phase. Hydraulic gradients always exist to drive dissolved-phase flow. Advective vapor-phase migration can occur if pressure gradients are present; In many instances, lateral vapor-phase migration is due only to diffusion and is negligible. Vapors will rise from contaminated ground water and therefore vadose-zone contamination will spread as dissolved-phase contamination spreads. [Note that local pressure gradients are often induced around building foundations due to air flux created by heating systems.]

Gasoline compounds are generally susceptible to aerobic degradation in the subsurface. Anaerobic degradation is probably orders of magnitude slower than aerobic degradation. Petroleum-utilizing bacteria are ubiquitous in the environment but their activity is typically oxygen- or nutrient-limited. Degradation is oxygen-limited in the "heart" of a plume and occurs very slowly. Degradation is significant at the leading edge of a plume, where mixing with more oxygenated waters occurs. For a plume to spread, the flux of contaminants must be greater than the rate of degradation. Many plumes associated with minor releases reach an approximate steady-state configuration. Because most of the contaminants are sorbed to sediments and dissolve slowly, the plume is continuously replenished and does not disappear, but the plume does not spread.

The immediate goals of site remediation are generally source reduction (i.e. remediation at the heart of the plume) and elimination of impact or threat to sensitive receptors. Because free product layers are sources of dissolved contamination and can produce dangerous vapor levels, free product removal is also an initial goal. Ideally, remedial design is undertaken after investigations are performed to provide an understanding of the distribution of contaminants and of the local ground water flow system. These investigations typically include monitor well installation, ground water and sediment sampling, determination of flow paths and velocity, and performance of ground water pumping tests and pilot studies of other technologies such as soil vapor extraction. In practice, many remediation systems are installed on an emergency basis before this information is available.

There has been much discussion recently regarding the limitations of ground water pump-and-treat technology for cleaning up sites. This is primarily because most contaminants are sorbed to soils and slowly leach into the ground water. It has become somewhat fashionable to dismiss pump-and-treat as ineffective. However, ground water pumping is the most effective way to prevent contamination migration by creating hydraulic control over the contaminated area. Prevention of migration must be incorporated into most remedial plans. Additional remedial activities are often warranted to provide more effective means of contaminant removal.

TRIP OBJECTIVE

Environmental consulting firms and regulatory bodies are the largest employers of geologists in New England. The environmental remediation market is projected to be a growth industry throughout the 1990's, and many of today's geology students will probably seek employment in the environmental field. The purpose of this trip is to introduce students to diverse types of remedial investigations and ground water and soil remediation systems. Four sites that are undergoing site remediation will be visited. Regulatory, scientific, engineering, political and economic constraints at each site will be discussed, and the decision-making process which led to the implementation of the system at each site will be presented.

SITE VISITS

Several features will be highlighted at each site. These features are summarized at the beginning of each site description. A narrative discussion of the site is also included in each description. Due to requests for privacy from the site owners, stop locations, a road log and figures with site-specific details are not provided herein.

Stop 1. Soil Vapor Recovery, Ground Water Recovery Trenches, and Activated Carbon Treatment

At this stop, investigative techniques for designing soil vapor extraction systems will be demonstrated. A soil vapor extraction and treatment system, and a separate ground water recovery and treatment system employing recovery trenches and activated carbon, are operational at the site.

This site has the following features:

* Complex geology with thick vadose zone
* Saturated sediments are poorly-permeable till
* Site access constraints
* Downgradient private water supply wells
* Soil vapor recovery system, soil vapor treatment, nested monitoring
* Ground water recovery trenches and pneumatic pumping system

The source of release at this site not known but is suspected to have been an underground storage system removed in the early 1980's. Apparently some primitive remediation was performed at that time, but almost no records documenting the work exist. Free product was observed seeping out of the base of a slope into a wetland in the late 1980's, prompting performance of a hydrogeologic investigation to delineate contamination at the site. The site is constructed on up to 30 feet of boulder-strewn fill. The native, saturated sediments are poorly-permeable till. Because of the great vadose thickness and the anticipated low yield and high cost of ground water recovery wells, soil vapor extraction (SVE) was proposed. A SVE design study, including installation of nested monitoring points, was performed. Results were used to design a SVE system. A schematic of the system, which is operational and has removed over 1,000 gallons of vapor-phase gasoline, is included on Figure 1.

Private wells are located downgradient of the facility. In order to prevent migration of contaminants towards the wells, two ground water recovery trenches were installed in a wetland downgradient of the site. Recovery trenches create a hydraulic barrier across their entire length, assuring that migration of dissolved-phase contaminants near the water table surface is eliminated.

Site 2. Air Stripping, Catalytic Oxidation and Redundant Systems

This stop will demonstrate a complex ground water recovery and treatment system and will also highlight some of the regulatory and political constraints that must be factored into remedial design. Air stripping, which is one of the most common and cost-effective ground water treatment systems, will be demonstrated and explained.

This site has the following features:

* nearby proposed municipal wellfield
* stringent local regulations
* high-flow setting
* on-site septic system; barrier wall installed
* ground water and soil vapor recovery
* redundant treatment systems with complex engineering controls

The site is located approximately 1,200 feet from a proposed municipal supply well. The site is situated in sand and gravel deposits, and yields of individual recovery wells range up to 20 gallons per minute. The site is too small for discharge of treated water to the ground, so ground water must be discharged to surface water via the storm drainage system after it is treated. The municipality in which the site is located has very stringent local ordinances governing discharges to surface water, which required that the remediation system be designed with several redundant treatment mechanisms and many fail-safe mechanisms to prevent discharge of any contaminants in the treated water. Because of the relatively high ground water recovery rates, air stripping was selected to treat the water. The stripper off-gases and soil vapors are treated by a catalytic oxidation unit, equipped with dual heat exchangers. One exchanger recovers heat to lower operation costs. The other pre-heats water entering the stripping tower to provide more effective stripping (Henry's Law constants increase with temperature, which means that organic vapors will partition better into the vapor phase at higher temperature.) The system is graphically

Stop #1

Schematic Layout

Figure 1

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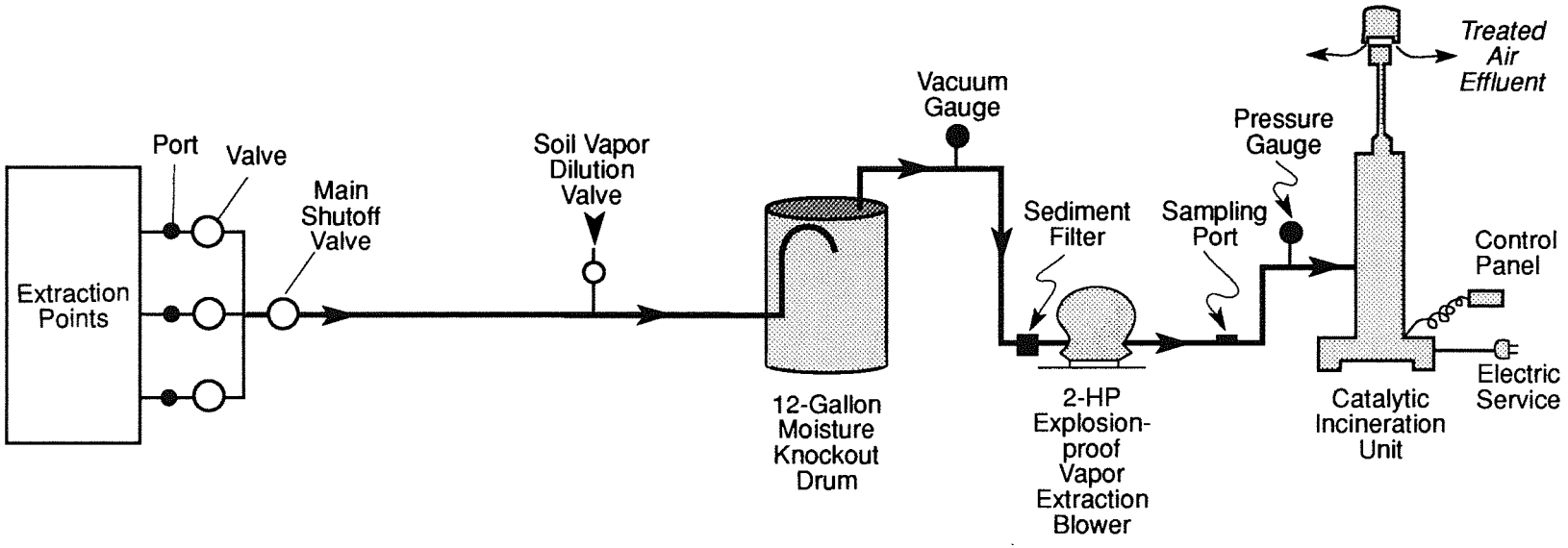


Schematic Layout
Not to Scale

➤ Direction of Air Flow

Date: 07/11/91
Revised: 01/03/92

Compiled by: CM
Drafted by: MED/KT

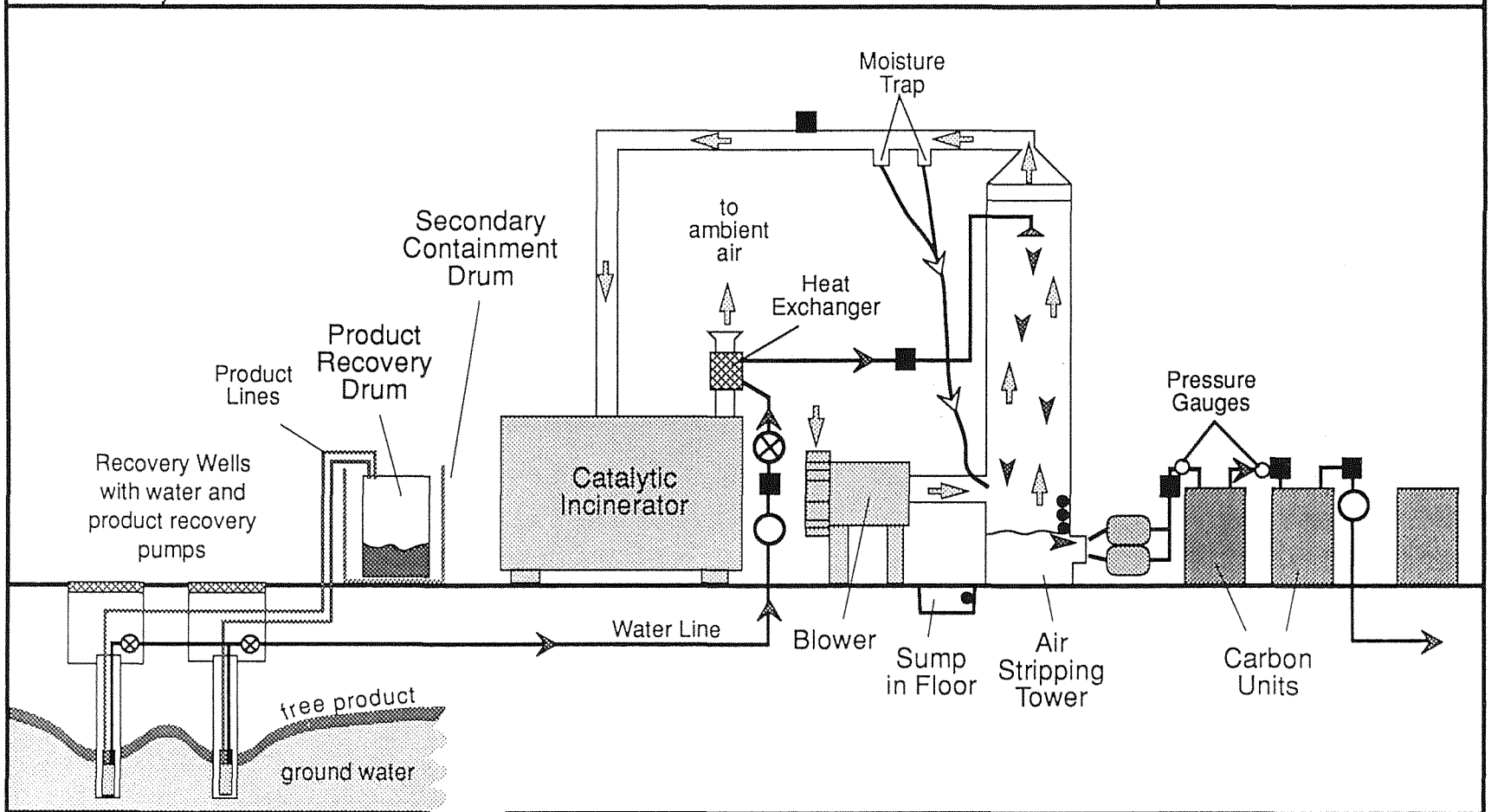
Stop #2

Figure 2

Treatment System Schematic

Hydro-Environmental Technologies, Inc.

54 Nonset Path
Acton, MA



Not to Scale

■	Sampling Port	●	Water Level Switch
⊗	Flow Meter	⮝	Water Flow Direction
○	Pressure Gauge	⇦	Air Flow, Direction

Date: 1/2/92
Revised: 4/8/92
Compiled by: DT
Drafted by: DT

summarized in Figure 2.

The entire remediation system was installed simultaneously with construction of a new station, which created interesting engineering opportunities and problems.

Stop 3. Low-Flow Multi-Well System and Ground Water Modelling

This stop will demonstrate an alternative remedial solution to ground water recovery in a low-flow environment. The use of aquifer simulation for design of remedial systems and for evaluation of impacts on sensitive receptors will be discussed.

The site has the following features:

* Thin, sandy, saturated zone over thick glacial lake bottom sediments.
* Confined aquifer, tapped by municipal wellfield beneath lake bottom sediments
* Relatively low dissolved concentrations but concern due to possible impact on wellfield
* Numerical aquifer simulation to design recovery system
* Low flow ground water recovery system including five recovery wells and granulated activated carbon treatment. Iron pretreatment system present.

The site is located in a former wetland which has been partially excavated and filled. The site geology presently consists of about five feet of sandy fill (three feet of which are saturated) overlying glacial lake bottom silts and fine sands. Peat and organic muck are present in patches. Borings near the site terminated at depths of 60 feet did not penetrate the bottom of the silt. A cross-section developed from borings performed at the site is included as Figure 3. Municipal wells less than 1,000 feet from the site tap a gravel layer located beneath the silt. Initially, a recovery trench was proposed to limit migration of contaminated ground water and to enhance recovery yield. However, regulators required the trench to be over-designed, with seemingly little anticipated technical benefit but great additional cost. The regulatory demand increased the cost of the recovery trench to the point where it was no longer feasible. Therefore, an aquifer simulation was performed using data developed during extensive subsurface investigations. The simulation, performed using the PLASM code, predicted that a wellfield of 5 recovery wells would provide hydraulic control over the plume. This system was proposed, approved and installed and is presently operative.

Monitoring wells have not been installed into the deep aquifer at the site, due to fear of cross-contamination. Deep and shallow well couplets installed downgradient of the plume indicate that there is a downward hydraulic gradient; however, groundwater flow velocities are probably low enough that no impact on the wellfield will occur. The possible impact on the confined aquifer will be modelled as part of a Risk Assessment for the site.

Stop 4. Flow in Bedrock and Protection of Water Supply

At this stop, the water table is located in bedrock in some areas and in the surficial deposits in others. Investigative techniques for complex sites such as this will be presented. Permitting issues associated with overlapping regulatory jurisdictions will be discussed.

* Complex hydrogeology, with the water table located in bedrock in some areas.
* Municipal reservoir located downgradient
* Fracture trace mapping to estimate preferred contaminant transport paths
* High dissolved iron concentrations and iron treatment system

Stop #3

Figure 3

Geologic Cross-Section E-E'

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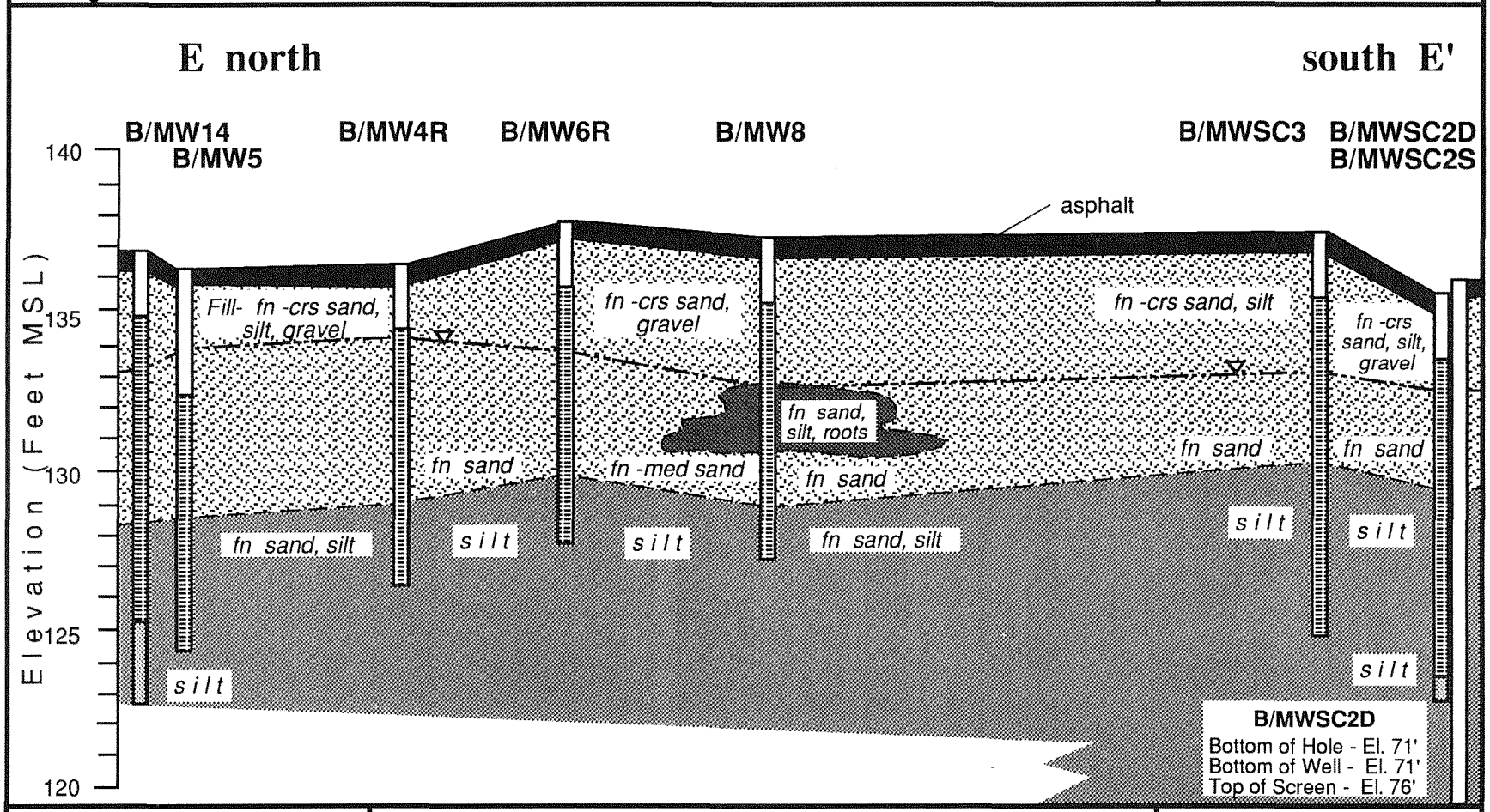

Horizontal Scale
0 — 50
Feet

Notes: 1) Soil descriptions generalized see boring logs for specific descriptions
2) Sedimentary unit boundaries are approximate
3) Piezometric data - 3/26/91

Well Casing
Well Screen
Boring

Date: 7/1/91
Revised:

Compiled by: AS
Drafted by: DT

Stop #4

Figure 4

Recovery System Schematic

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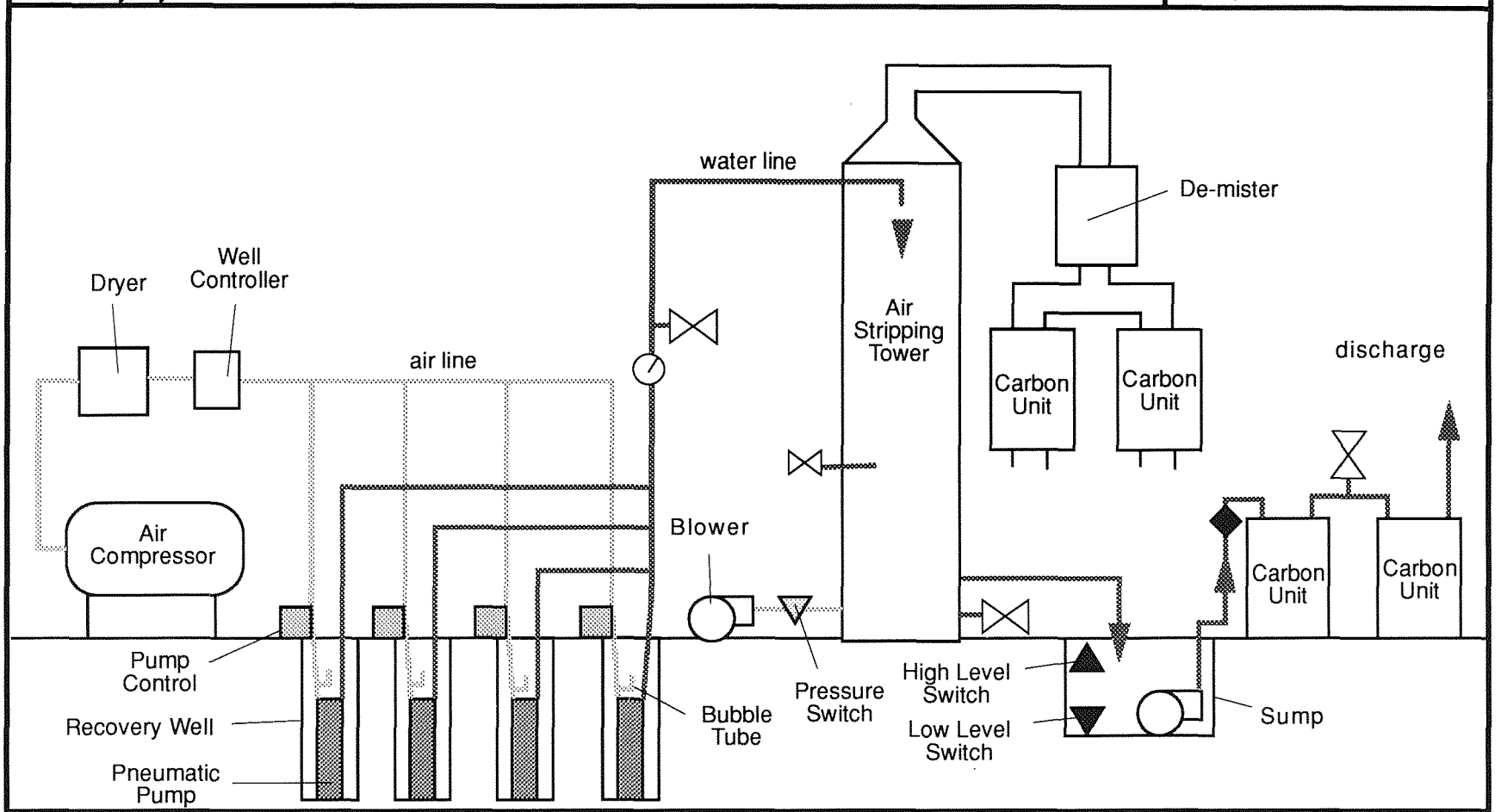

Not to Scale

- Check Valve
- Sampling Port
- Pressure Switch
- Pump Controller
- Water Level Switch
- Water Line
- Air Line
- Blower/Pump
- Flow Totalizer
- Flow Direction

Date: 2/27/92
Revised: 7/9/92 MS
Compiled by: MS
Drafted by: DT

Free product was detected on the water table during a site assessment conducted prior to a tank replacement. A remedial effort was initiated on an emergency basis but was not brought on line for over a year due to delays in obtaining permits from local regulators. The delay was due to refusal by a local water department to grant permission for discharge of treated water, primarily because the water department insisted on using the permit as leverage for negotiations on an unrelated issue. Both the client and state regulators were anxious to begin remediation to prevent possible impacts on water supply, but were unable to do so. The emergency remedial system is now operational (Figure 4). Additional remedial measures are being planned as of this writing.

STRATIGRAPHY AND STRUCTURAL GEOLOGY IN THE ACADIAN GRANULITE FACIES

by
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INTRODUCTION

The Merrimack belt in central Massachusetts (fig. 1) contains strongly deformed, metamorphosed sedimentary rocks. Similar rocks are preserved among the abundant plutons of New Hampshire and extend into central Maine where the metamorphic intensity is less and Silurian fossils are present. The stratigraphic interpretation presented on this trip depends on correlations extended from the more complete, better preserved, and fossil-controlled stratigraphic sections of Maine. On this trip we will see rocks in the area between Brimfield and Sturbridge, Massachusetts and Union, Connecticut (fig. 2). This area includes the highest grade Acadian metamorphic rocks in the Appalachians, reaching into the granulite facies. Even though the rocks have been affected by partial melting and high-grade metamorphism, relict primary stratigraphic features can be discerned.

Because the stratigraphic interpretation is so vitally important, one objective of the trip is to show participants a variety of common and distinctive rock types in hopes that they may be familiar to workers from surrounding areas. The second objective is to demonstrate the critical relationships on the ground that have been used to reconstruct the sequence of structural, plutonic, and metamorphic events in this area, independent of regional stratigraphic considerations.

PREVIOUS WORK

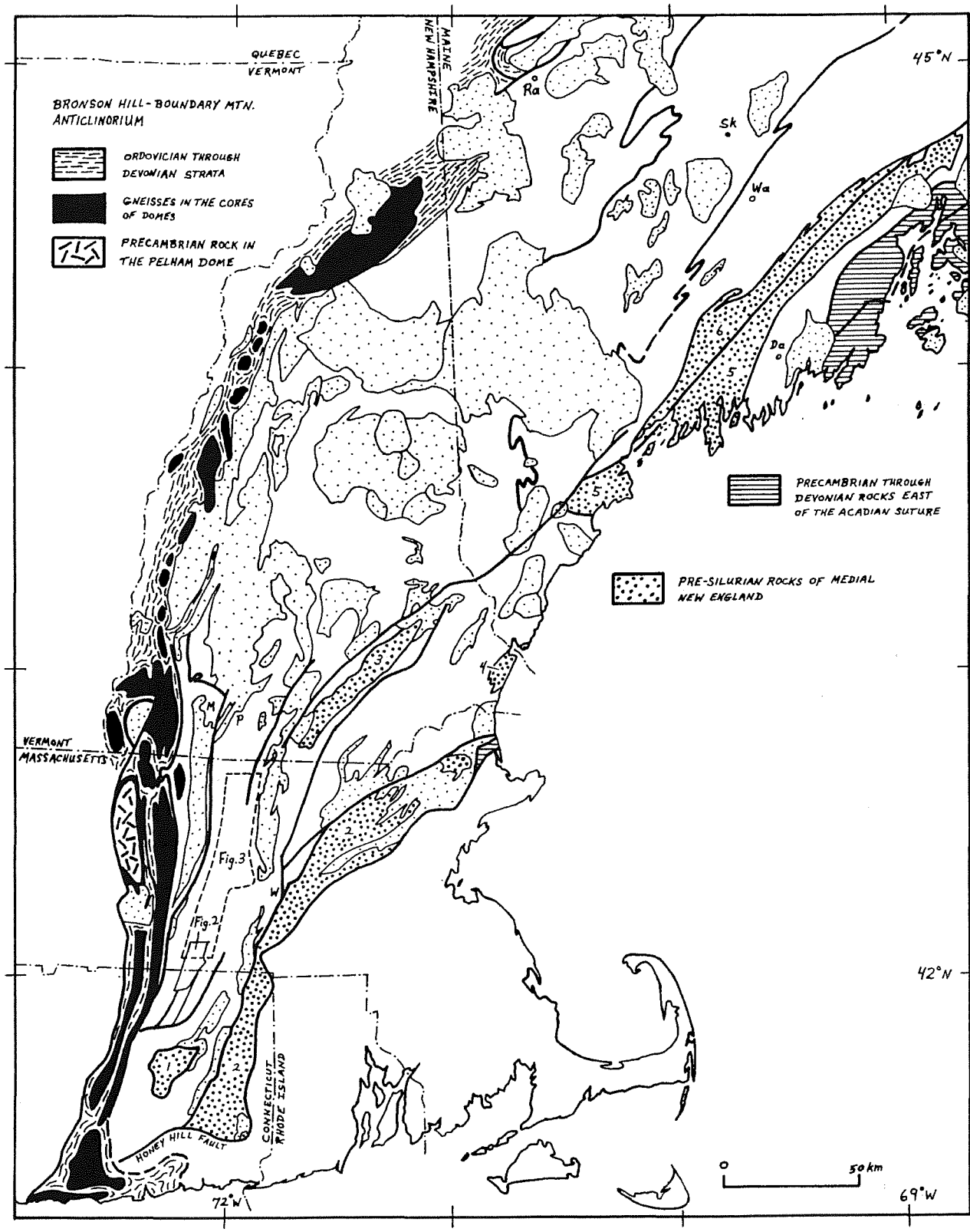
Emerson (1917) made the distinction in central Massachusetts between Brimfield schist, dominated by sulfidic, graphitic schists, and Paxton quartz schist, a flaggy, biotitic quartz schist with calc-silicate lenses. In the late 1960's and early 70's, a U.S.G.S detailed mapping program in the area around the Massachusetts-Connecticut state line produced 1:24,000 quadrangle maps for the Warren (Pomeroy, 1977), Wales (Seiders, 1976), Westford (Peper and Pease, 1975), and Eastford (Pease, 1972) quadrangles (fig. 2). The increased detail of this work allowed stratigraphic subdivision into new units, as summarized by Peper and others (1975), and Peper and Pease (1976). They assigned rocks of Emerson's Brimfield Schist and western parts of his Paxton quartz schist to the Hamilton Reservoir Formation, and eastern parts of the Paxton quartz schist were divided into the Bigelow Brook and Southbridge Formations. All three west-dipping Formations were interpreted to comprise a non-repeated, westward-facing stratigraphic sequence from the Southbridge up through the Bigelow Brook and the Hamilton Reservoir.

Peper and Pease (1976) recognized the overall similarity of the Hamilton Reservoir Formation to rocks in the Peterborough quadrangle, southern New Hampshire as mapped by Greene (1970). Since that time, there has been a major revision of the stratigraphy in the Peterborough area (Duke, 1984; Duke et al., 1988), based on extensions from western Maine through central New Hampshire (Hatch et al., 1983). This suggested that the Hamilton Reservoir Formation, by chain reaction, might also warrant a new stratigraphic interpretation. My early suggestions (Berry, 1985; Robinson et al., 1989) have been more fully developed (Berry, 1989) and are presented on this trip. In re-mapping this area, the distribution of rocks as shown on the U.S.G.S. quadrangle maps has not changed significantly, except for the additional mapping of thin white schist units (fig. 2). Interpreted stratigraphic and structural relationships, however, have changed dramatically.

STRATIGRAPHY

Leadmine Pond Gneiss

The Leadmine Pond Gneiss is a new, informal name applied to a unit characterized by medium-gray to white plagioclase-quartz-biotite gneiss with subordinate interlayered biotite-richer gneiss and amphibolite. Local varieties may also include combinations of orthopyroxene, clinopyroxene, hornblende, garnet, K-feldspar, and magnetite. The gneisses occur in a variety of textural types from thinly layered to massive, with layered types more common. In addition, the Leadmine Pond includes subordinate thin units of schist, quartzite, granofels, sillimanite-bearing gneisses, and calc-silicate rocks, interpreted to be metamorphosed


QUEBEC
VERMONT
MAINE
NEW HAMPSHIRE
45°N
BRONSON HILL-BOUNDARY MTN. ANTICLINORIUM
ORDOVICIAN THROUGH DEVONIAN STRATA
GNEISSES IN THE CORES OF DOMES
PRECAMBRIAN ROCK IN THE PELHAM DOME
Ra
Sk
Wa
Da
PRECAMBRIAN THROUGH DEVONIAN ROCKS EAST OF THE ACADIAN SUTURE
PRE-SILURIAN ROCKS OF MEDIAL NEW ENGLAND
VERMONT
MASSACHUSETTS
M
P
W
Fig.3
Fig.2
42°N
CONNECTICUT
RHODE ISLAND
HONEY HILL FAULT
72°W
0
50 km
69°W

sedimentary rocks. The Leadmine Pond Gneiss is named for Leadmine Pond in Sturbridge, Massachusetts, where the largest section through the unit is exposed (fig. 2). It is also exposed in several thin, structurally isolated belts to the west.

Rocks assigned here to the Leadmine Pond Gneiss were previously interpreted to be gneiss members interstratified with schists of the Hamilton Reservoir Formation (Seiders, 1976; Peper et al., 1975). A major difference between the two interpretations is that the Leadmine Pond Gneiss is thought to be older (Precambrian?-Ordovician?) basement upon which the (Silurian?) Rangeley Formation was deposited, whereas the alternating schist and gneiss members of the Hamilton Reservoir Formation are thought to be interstratified parts of a single unit that becomes progressively younger toward the west. Geochemical studies of the gneisses are underway that might discriminate between these two interpretations, although the effects of granulite facies metamorphism and partial melting must be understood before the igneous characteristics of the gneisses can be isolated. Any geochronological studies intended to assign igneous crystallization ages are likewise expected to be significantly complicated by inherited and metamorphic components.

Rangeley Formation

The Rangeley Formation in the Brimfield-Sturbridge area is interpreted to rest unconformably on the Leadmine Pond Gneiss. It is characterized by gray-, red-, brown-, or rusty-weathering schists interlayered with quartz-rich, feldspathic or calc-silicate granofels. The granofels layers are commonly 1/2 to 15 cm in thickness, but their abundance relative to schist is highly variable over short distances. In detail, the internal stratigraphy of the Rangeley is complex. Thin units of gray-weathering schist (Srg), sulfidic white schist (Srw), and calc-silicate granofels (Src) are mapped separately, some of which are shown in figure 2. In the western part of the field trip area, the upper part of the Formation is exposed, including the upper contact against the Smalls Falls Formation. East of the westernmost exposure of Leadmine Pond Gneiss, the Rangeley is exposed in a series of thrust slices that contain the base and lower parts of the Formation. A sequence of thin units of white schist, calc-silicate granofels, and gray schist are present in several of the slices within 300 meters of the base of the unit, suggesting a persistent set of thin units may be present.

Smalls Falls and Madrid Formations

The Smalls Falls and Madrid Formations succeed the Rangeley conformably. These two units are only preserved in the north-central part of figure 2, and are cut out southward by a fault. For lithologic descriptions, see Stop 3.

Paxton Formation

The field trip ends to the east in the Bigelow Brook member of the Paxton Formation. A lithologic distinction has been made farther to the south and east between the Bigelow Brook and Southbridge Formations (Pease, 1972), but this distinction has not been mapped northward from the state line very far into Massachusetts. Therefore, much of central Massachusetts remains mapped as the Paxton Formation, a

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Figure 1 (facing page). Regional geologic setting. Area of field trip, shown in more detail in fig. 2, is indicated on Massachusetts-Connecticut line. Field trip lies within an area of Silurian-Devonian strata (not patterned) that extends through Merrimack County, New Hampshire into central Maine. This area is bounded by older rocks to the west (Bronson Hill and Boundary Mountains anticlinoria) and by the Honey Hill and other faults of various age (heavy lines) to the east. Fine stipple: selected plutons. Localities: Ra=Rangeley, Sk=Skowhegan, Wa=Waterville, and Da=Damariscotta, Me.; P=Peterborough and M=Mt. Monadnock, N.H.; W=Worcester, Mass. Areas of pre-Silurian rocks in medial New England: 1=Willimantic dome, 2=Putnam-Nashoba zone, 3=Massabesic Gneiss, 4=Rye Fm., 5=Saco-Harpswell sequence, 6=Falmouth-Brunswick sequence. Original compilation from: Osberg et al. (1985), Osberg (1988), Moench and Pankiwskyj (1988), Hussey (1985; 1988), Berry and Osberg (1989), Billings (1956), Thompson et al. (1968), Lyons (1979), Bothner et al. (1984), Thompson (1985), Hall and Robinson (1982), Zen et al. (1983), Robinson et al. (1991), Peper and Pease (1976), Rodgers (1985), Wintsch (1985), Getty and Gromet (1992).

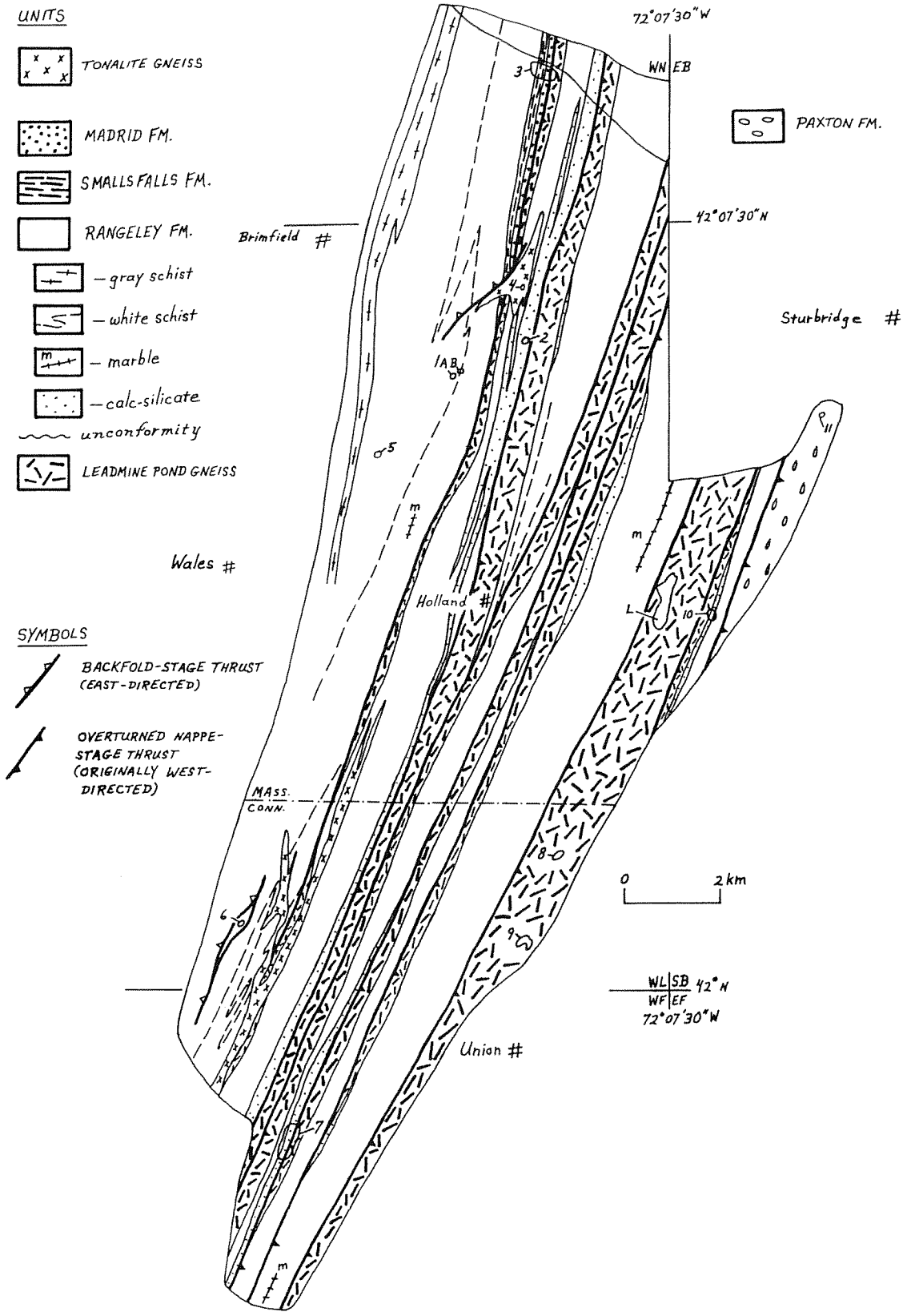


Figure 2. Generalized geologic map of the Brimfield-Sturbridge area. Numbers indicate field trip stops. See figure 1 for location. L=Leadmine Pond. Quadrangle names: WN=Warren, EB=East Brookfield, WL=Wales, SB=Southbridge, WF=Westford, EF=Eastford. (Adapted from Berry, 1989.)

heterogeneous unit that probably spans the entire Silurian Perod (Zen et al., 1983; Robinson and Goldsmith, 1991). The rock type characteristic of the Paxton Formation is gray-weathering, slabby, quartz-plagioclase-biotite granofels that is purplish-gray on a fresh surface (Stop 11). Thin layers and pods of green calc-silicate granofels are commonly present and locally abundant. Many lithologic units within the Paxton have been mapped, including rusty schist, rusty quartzite, and gray schist (figs. 2, 3), but shallow dips, poor outcrop, complicating late faults, and a probably complex internal stratigraphy have left stratigraphic relationships among many of these units unresolved.

Stratigraphic relationships between the Rangeley and Paxton Formations are also uncertain. Where the Rangeley is predominantly schist and the Paxton is predominantly granofels, the distinction is fairly straightforward. But in places where the Rangeley is schist with interlayered granofels and the Paxton is granofels with interlayered schist, the distinctions are more subtle, suggesting that the two Formations may at least partly interfinger. The Paxton has not been mapped since the Rangeley explosion of the past ten years, and the new stratigraphic perspective might help to solve some of these problems. Similar problems confronted workers in central Maine where the Huchins Corner (formerly Vassalboro) Formation was originally equated with the Madrid (formerly Fall Brook) Formation (Osberg, 1968), but now the Hutchins Corner is placed at the bottom of the section, and the Madrid is at the top (Osberg, 1988).

STRUCTURAL GEOLOGY

General Features

The geometric pattern of units in figure 2 and the surrounding region appears to be fairly simple. All layering and foliation strike north-northeast and dip to the west at 35 to 60° for hundreds of square kilometers. Local repetition of strata (Stop 7), and low-angle truncation of units (fig. 2), however, indicate the presence of isoclinal folds and map-scale faults. The significance of isoclinal folding is difficult to assess because fold hinges are approximately horizontal, so that few fold noses are exposed on the ground. Furthermore, some areas of the map lack distinctive marker horizons that would clearly discriminate between structural repetition and stratigraphic complexity. Relative motion of the faults is likewise difficult to prove in all cases, because not all faults display fault-related fabrics, although some do. Differences in local interpretation have produced significantly different tectonic models for this region (Rodgers, 1981a; 1981b; Robinson and Tucker, 1981), hinging mainly on the inferred sense of motion on the major map-scale faults. Recent work indicates that not all faults are the same age, and that both west-directed and east-directed faults are present (fig. 2).

While not all features can be assigned to a definite place in the deformational sequence, cross-cutting relationships and timing with respect to metamorphism allow certain features to be assigned to either the nappe stage, backfold stage, or dome stage of Acadian deformation (Robinson and Hall, 1980).

Nappe Stage

During the nappe stage of Acadian deformation, fold nappes and thrust nappes were transported westward over the Bronson Hill anticlinorium in Massachusetts and New Hampshire (Thompson et al., 1968; Robinson and Hall, 1980; Thompson, 1985; Robinson et al., 1991). Map-scale isoclinal folds and faults in the Brimfield area that pre-date the metamorphic peak are assigned to this stage. The nappe-stage folds are recognized by stratigraphic repetition and are truncated by nappe-stage faults (Stop 7). The nappe-stage faults (stops 3 and 7) are intruded, in turn, by tonalites that were recrystallized during the peak metamorphism (Stop 4). Therefore, the nappe-stage folds and faults pre-date the peak of metamorphism.

After the nappe-stage faults formed, the rocks were recrystallized at least during the peak metamorphism and perhaps during a pre-peak-metamorphic, andalusite-forming regional metamorphism and were affected by at least two phases of penetrative, ductile deformation. It is not surprising that even where the nappe-stage faults are exposed in outcrop no fault-related fabrics of this age have been recognized. The sense of displacement on the nappe-stage faults is based entirely on the inferred stratigraphic offset which depends on the stratigraphic interpretation. Each nappe-stage fault brings the Leadmine Pond Gneiss to the east against the Rangeley, Smalls Falls, or Madrid to the west. This gives an east-side-up sense of offset (Stop 3), consistent with the west-directed thrust nappes of the Bronson Hill in New Hampshire and northern

Massachusetts. The fact that these faults now dip to the west is attributed to the later, backfold stage of deformation.

Backfold Stage

The nappe-stage Brennan Hill and Chesham Pond thrusts in the western part of the Mondadnock quadrangle, New Hampshire, are west-directed, east-dipping thrust faults (Thompson, 1985). They can be traced southward along the east side of the Bronson Hill anticlinorium into southern Massachusetts where they are west-dipping, overturned to the east by a major backfold (Robinson et al., 1991; Robinson and Elbert, this vol., trip A-3). Westward dips continue uninterrupted from the east edge of the Bronson Hill anticlinorium to the Brimfield-Sturbridge area where the nappe-stage thrusts are apparently also overturned. The present, west-dipping orientation of the nappe-stage folds and thrusts in the Brimfield area is therefore attributed to eastward overturning during the backfold stage.

Minor structural features that are assigned to the backfold stage of Acadian deformation include mylonite zones from a centimeter to 100 meters in thickness (Stop 6), ductile shear zones, asymmetric boudinage and low-angle truncations of foliation, and a prominent west-plunging mineral lineation. These features are interpreted to have formed slightly after formation of the peak metamorphic minerals because peak metamorphic minerals are deformed, and yet the same minerals are recrystallized along shear planes and in the lineation direction. Where diagnostic, kinematic indicators consistently show west-side-up motion, although most rocks in the region contain a flattening fabric without obvious asymmetry. Apparent low-angle cross-bedding reported by previous workers (Peper et al., 1975) might be interpreted alternatively as asymmetric composite metamorphic foliation.

The backfold-stage deformational features are widespread and appear to be concentrated in certain lithologic types, particularly the tonalites and layered gneisses, rather than along mapped faults in the area of this field trip. Elsewhere, major faults both to the west (Peterson, this volume, trip A-1) and to the southeast (Peper et al., 1975) do contain associated minor features consistent with the east-directed backfold stage.

Dome Stage

Late, asymmetric minor folds are common in central Massachusetts (Robinson, 1979). Their hinge lines trend north-south and plunge within 15° of horizontal. A common, north-south mineral lineation is assigned to the same deformational phase. This lineation has been mapped to the northwest where it is kinematically related to the Bronson Hill gneiss domes. The remarkable similarity in texture between the east-west lineation and the north-south lineation give a cloth-like fabric to some samples, and suggest that the backfold- and dome-stage lineations formed under similar, high-grade metamorphic conditions.

Post-Acadian Features

While rocks in the field trip area do not appear to carry penetrative post-Acadian deformational fabrics, significant high-temperature, ductile deformation of Late Paleozoic age occurred to the south in the Willimantic dome (Getty and Gromet, 1992), to the east in the Worcester area (Goldstein, this vol., trip B-2), to the northwest in the Pelham dome (Robinson and six others, this vol., trip B-3), and to the northeast (at least thermal effects) in the Massabesic Gneiss (Eusden and Barreiro, 1988). The Oakham and Gardner anticlines are late foliation arches of unknown age (Robinson and Tucker, this vol., trip C-3), that have some structural similarity to the Willimantic dome. In some ways the most remarkable feature of the the Brimfield-Sturbridge area is that it was insulated from high-grade Late Paleozoic deformation (Gromet, 1989).

METAMORPHISM AND PLUTONISM

The entire field trip is in Zone VI, the highest-grade regional metamorphic zone in Massachusetts (Thomson et al., this vol., trip C-4). Only rocks at the easternmost stop (Stop 11) appear to be partially retrograded. U-Pb monazite analyses indicate that the age of peak metamorphism is Late Devonian (362-369 Ma), consistent with Acadian regional metamorphism and plutonism in central New Hampshire and central Maine (Thomson et al., this vol., trip C-4).

No crystallization ages have been determined for intrusive rocks within the region of figure 2. Samples collected by M.H. Pease from the Hedgehog Hill Gneiss to the south in the Stafford Springs quadrangle, Connecticut (Pease, 1975), yielded zircons that were analyzed by R.E. Zartman in 1979. The Hedgehog Hill Gneiss is an elongate body of foliated, metamorphosed tonalitic gneiss that extends northward from the Stafford Springs quadrangle into the southern part of figure 2 where it cuts a nappe-stage fault. Two zircon size-fractions were analyzed and both were discordant. Zartman (1988, written commun.) interpreted the discordia drawn through both analyses to give a meaningful upper intercept age of about 445 Ma, assuming the discordance was due to a post-crystallization Pb-loss event. Such an old crystallization age for a cross-cutting pluton suggests that the Hamilton Reservoir Formation should be revised from Silurian-Devonian (Peper et al., 1975) to pre-Late Ordovician (Pease and Barosh, 1981) or even pre-Ordovician (Barosh, 1982) in age. Alternatively, if the discordance of the Hedgehog Hill Gneiss zircons is due to a mixture of older xenocrysts with newly-formed igneous zircon, the intrusive age would be *younger* than the Pb-Pb age, allowing a Silurian age for the country rocks. Further, detailed study of the zircon population is required before a crystallization age can be assigned to the Hedgehog Hill Gneiss.

REGIONAL CORRELATION

Leadmine Pond Gneiss

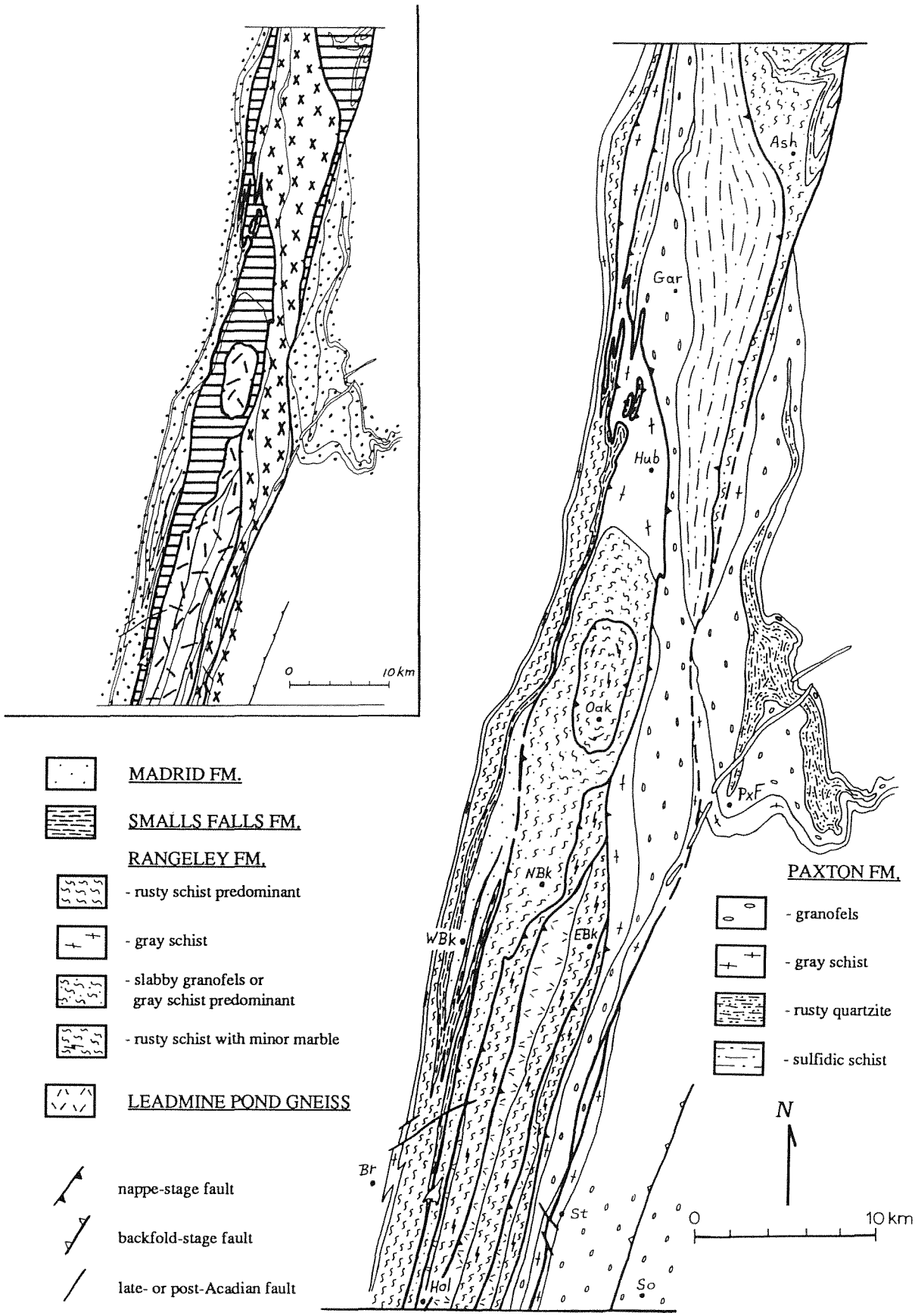
Part of the reason for considering the Leadmine Pond Gneiss to be pre-Silurian basement is that the Silurian sections of New Hampshire and Maine do not contain within them similar sorts of thick, interstratified gneisses. On the other hand, pre-Silurian units in the surrounding region do contain rocks and sequences of rocks similar to those of the Leadmine Pond Gneiss. Units that contain partly similar rocks include the Monson Gneiss and Ammonusuc Volcanics of the Bronson Hill anticlinorium in central Massachusetts; the Tatnic Hill Formation in eastern Connecticut and Nashoba Formation in eastern Massachusetts of the Putnam-Nashoba zone (see Goldsmith, 1991 for review); the Rye Formation in coastal New Hampshire; and part of the Cushing Formation in the Falmouth-Brunswick sequence, Maine (Hussey, 1985; 1988) (fig. 1). The broad similarities among these units suggests that a single continental basement, not to be confused with Cambrian-defined Avalon, extended beneath medial New England in Late Ordovician time (Osberg, 1978; Hall and Robinson, 1982; Berry and Osberg, 1989).

Silurian(?) Strata in Massachusetts

In the interpretation presented here for the Brimfield-Sturbridge area (fig. 2), the major structural features are imbricated nappe-stage thrust faults overturned to the east. This interpretation is incompatible with that of Zen et al. (1983) who showed the anticline at Oakham (fig. 3) extending southward, implying that rocks east of Holland face east and rocks west of Holland face west. It is also incompatible with the model of Rodgers (1981) in which the rocks face predominantly west. A new suggestion is offered in figures 3 and 4 which attempts to meld the interpretation of the Brimfield-Sturbridge area with previous mapping in the Oakham area to the north. This proposal is a conceptual model that is meant to inspire further field investigations.

In constructing figures 3 and 4, several observations and bits of information from different sources have been incorporated. These include:

- The Smalls Falls Formation is bounded to the east by the Madrid Formation in the West Brookfield area, the Leadmine Pond Gneiss in the Brimfield area, a gray schist unit in the Hubbardston area, and it ends northward at about the latitude of Gardner. Therefore, a fault is proposed that extends from the Brimfield area (Stop 3) northward across Massachusetts, approximately along the east side of the Smalls Falls.
- Field (1975) noted that the rocks east of the Smalls Falls at West Brookfield, mapped by him as Paxton Schist and shown here as Madrid Formation (fig. 3), have thinner layering and less calc-silicate rock than Paxton to the east. The fault proposed along the east side of the Madrid in the area north of W. Brookfield (fig. 3) is intended to separate these two rock types, although this subtle contact has not been mapped in the field.
- A unit of gray schist at Hubbardston is in contact to the south with a map unit dominated by slabby gray granulites with common calc-silicate granofels and interlayered rusty schist. This unit was


0
10 km
MADRID FM.
SMALLS FALLS FM.
RANGELEY FM.
- rusty schist predominant
- gray schist
- slabby granofels or gray schist predominant
- rusty schist with minor marble
LEADMINE POND GNEISS
nappe-stage fault
backfold-stage fault
late- or post-Acadian fault
Ash
Gar
Hub
Oak
PxF
NBk
WBk
EBk
Br
St
Hol
So
PAXTON FM.
- granofels
- gray schist
- rusty quartzite
- sulfidic schist
N
0
10 km

mapped by Tucker (1977) north of the Oakham anticline as a member of the Paxton Schist, and is assigned here to the Rangeley (fig. 3). Similar rocks are present in the Brimfield-Sturbridge area in the belt that includes Stop 2 (fig. 2). A similar set of rock-types is also present in the Ashburnham area, mapped by Peterson (1984) as members of the Littleton and Paxton Formations. This unit of the Rangeley includes thin layers of sulfidic white schist in the Brimfield (Berry, 1989), North Brookfield (Zen et al., 1983), and Ashburnham (Peterson, 1984) areas.

- The Rangeley rocks around Holland are dominated by rusty schists, but also include thin units of layered marble (fig. 2). Thin units of cocolitic limestone (graphitic diopside marble) were mapped by Emerson (1917) around East Brookfield. Rocks in the center of the Oakham anticline (fig. 3) are dominantly rusty schist. Although not exposed at the surface, limestone beds up to 15 feet thick are present in the rusty schist, reported by Fahlquist (1935) from the Quabbin aqueduct tunnel.
- A fault is postulated west of Gardner to account for apparent low-angle truncations of units along strike. In particular, the sulfidic schist west of Gardner ends just southwest of Gardner; the gray schist at Hubbardston ends toward the southeast; the belts of Leadmine Pond Gneiss in the Holland area do not extend as far north as Oakham; and the gray schist west of Sturbridge extends north to the latitude of Oakham and seems to disappear. Robinson and Goldsmith (1991) offer an alternative interpretation that does not require a fault here.
- The Bigelow Brook member of the Paxton near Sturbridge is similar to the Paxton Formation near Gardner in that they both contain mappable units of massive, brown-weathering sulfidic schist associated with the characteristic purplish-gray Paxton granofels.

The interpretation of figures 3 and 4 accounts for these stratigraphic features mainly by "telescoping" along nappe-stage faults. Additional backfold-stage faults and younger faults may well complicate the picture.

Figure 3 (facing page). Geologic map of central Massachusetts. See figure 1 for location. Inset: distribution of lithotectonic belts. Towns: Ash=Ashburnham, Gar=Gardner, Hub=Hubbardston, Oak=Oakham, PxF=Paxton Falls, NBk=North Brookfield, WBk=West Brookfield, EBk=East Brookfield, Br=Brimfield, St=Sturbridge, Hol=Holland, So=Southbridge. Reinterpreted from Zen et al. (1983) based on Emerson (1917), Field (1975), Tucker (1977), Peterson (1984), Berry (1989), and Robinson and Goldsmith (1991).

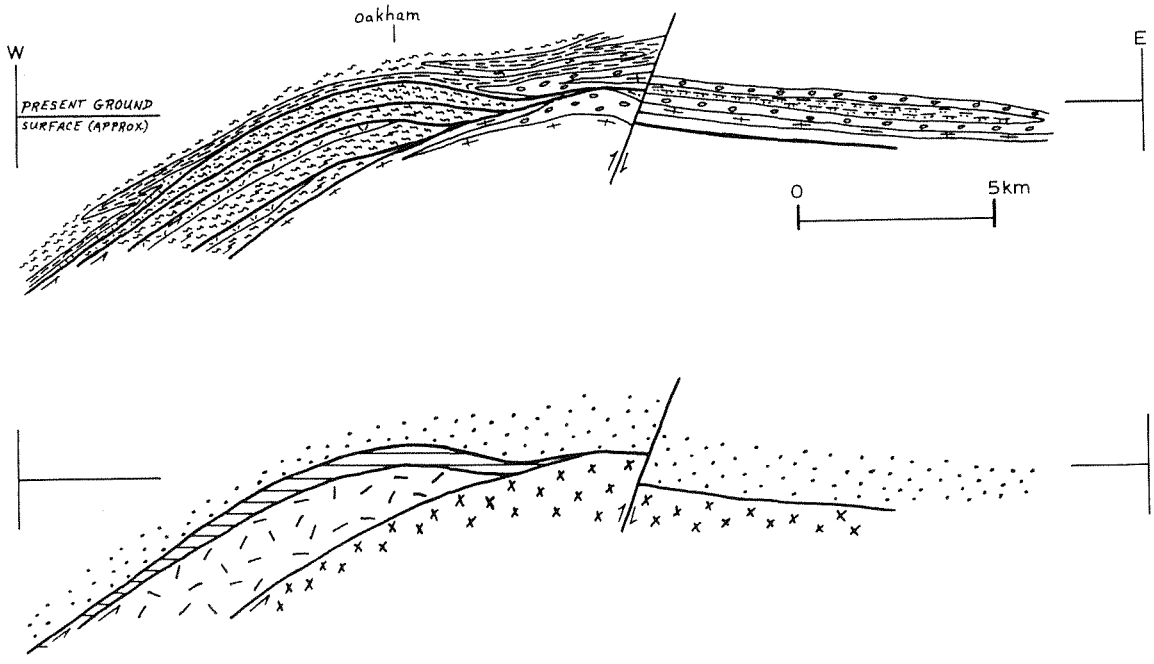


Figure 4 (above). Interpretive cross-sections through the area shown in figure 3. Line of section runs east-west through Oakham. Upper section patterned according to geologic units as in main part of figure 3. Lower section patterned according to lithotectonic units as in figure 3 inset.

Silurian Strata in Maine

In a general way, the Brimfield - West Brookfield strata are similar to parts of the section near Rangeley, Maine (fig. 1), with passable Smalls Falls and Madrid correlatives. The parts of the Rangeley in the North Brookfield - Ashburnham and Holland - East Brookfield areas contain more calc-silicates and minor limestones, more like Silurian rocks in the Skowhegan and Waterville areas. To the east, the Southbridge Formation is a layered granofels unit free of schist, much like the Bucksport Formation in the Damariscotta area (fig. 1). It is suggested that the distribution of Silurian facies that has been described in Maine (Moench and Pankiwskyj, 1988; Osberg, 1988) from proximal in the west (Rangeley) to distal in the east (Damariscotta) is also present in the Merrimack belt of Massachusetts. But the geologic elements that cover 120 km across strike in Maine, can be found within 20 km across strike in Massachusetts (fig. 1). In this model, the Southbridge, Bigelow Brook, and Hamilton Reservoir Formations might represent different Late Ordovician-Silurian facies that have been thrust together, rather than a continuously westward-younging stratigraphic section >21-km-thick (Peper and Pease, 1976).

ACKNOWLEDGMENTS

This field trip is based on the author's Ph.D. research at the University of Massachusetts (Berry, 1989) augmented by subsequent research. Guidance, discussions, assistance, and support from Peter Robinson during that time are gratefully acknowledged. Geologists familiar with Robinson's work will notice his influence in the content and style of this paper. Field work was supported by N.S.F. research grants EAR-81-1697, EAR-84-10370, and EAR-86-08762 (all to Peter Robinson).

ROAD LOG

Participants will leave at 8:00 Saturday morning from the parking lot north of the University of Massachusetts Football Stadium in Amherst according to meeting announcements. We can leave extra cars there, and travel together to the beginning of the trip in Brimfield, about an hour from Amherst.

Starting Place: Public parking lot 150 feet northeast of traffic light at the intersection of Rt. 20 and Rt. 19, Brimfield, Massachusetts at 9:00.

Mileage

0.0	Turn left from SW corner of parking lot.
0.05	Bear right around obelisk, then turn left onto Route 20 East.
0.4	Turn right onto Holland Road.
0.5	Bear left at fork, toward Holland.
1.2	Turn left onto Five Bridge Road.
1.4	Pass flood-control gate.
1.8	Park to the right.

STOP 1. VARIETIES OF THE RANGELEY FORMATION. (Note: Stop locations are indicated in figure 2.) In the Brimfield area a Range of lithologic types is assigned to the Rangeley. We will see two types here at Stop 1, a third type at Stop 2, and other variants during the day, some of which have been mapped separately.

1A. Schist and Granofels. Walk northwest from the road 300 feet along intended railroad bed to a cut on the right (northeast). The rock consists of rusty-weathering sillimanite-orthoclase-quartz-biotite-garnet-graphite-ilmenite-pyrhhotite ± cordierite schist interlayered with slabby, bluish-gray quartz-plagioclase-biotite ± garnet granofels in layers 5 to 15 cm thick. Schist is subordinate to granofels at this outcrop, but the proportion of schist to granofels is variable over very short distances and has not proven to be a useful criterion for mapping. The rock type here is common in the Rangeley Formation of the Brimfield area. This is probably what Emerson (1917) meant by Brimfield schist, since it is a "rusty, graphitic biotite schist" in Brimfield, but he did not give a specific type locality.

These rocks are lithologically similar and trace directly into rocks mapped as Rangeley in the eastern Monadnock (Thompson, 1985) and western Peterborough (Duke, 1984) quadrangles, New Hampshire (fig.

1). Compare this outcrop with Stop 6 of Thompson (1988) and with Stop 1 of Duke et al. (1988). It is by extension from those rocks in southern New Hampshire that the name Rangeley is used here.

Return to road. Turn left (east). Walk a short distance along Five Bridge Road. Turn left (north) onto old logging road. Walk north 600 feet along road to decrepit pavement outcrop.

1B. White Schist. This rock is a very sulfidic, sillimanite-rich schist. It contains mostly white minerals, namely quartz, sillimanite, orthoclase, cordierite, and pale to white, Mg-rich biotite; therefore the informal name "white schist" was used by Field (1975) for this rock type. The black flakes in the rock are graphite, and rutile rather than ilmenite is the primary titanium-bearing oxide phase. White schist can be distinguished in the field from ordinary schists such as the one at Stop 1A by the presence of tiny brown rutile needles and pale biotite (muscovite not being stable in schists at this metamorphic grade), a lack of garnet, and a bright orange weathering color with red splotches. This rock represents a metamorphosed, carbonaceous black shale. Sulfur isotopes in white schists from the region preserve a primary, biogenic signature indicating that the high sulfur content is inherited from the sedimentary protolith (Tracy and Rye, 1981; Tracy and Robinson, 1988).

This locality was mapped in reconnaissance for the state map as the southern extent of the sulfidic schist and quartzite unit of the Paxton Formation (unit Spsq of Zen et al., 1983; Robinson and Goldsmith, 1991). Subsequent mapping has shown, however, that the white schist here does not trace into the Spsq unit to the north. Furthermore, the Spsq unit rests against granulites of the Paxton Formation to the east while the white schist here is enclosed by schist units on both sides, now assigned to the Rangeley. Therefore, despite its similar lithology, the white schist here is interpreted as a lens in the Rangeley, and not correlative with the white schist of unit Spsq. Lenses of black, carbonaceous pelite are similarly present within the Silurian Sangerville Formation of west-central Maine (Ludman, 1976). We will see the white schist of Spsq (Smalls Falls Formation) and its stratigraphic setting at Stop 3.

Return to cars. Continue east on Five Bridge Road.

2.0 Outcrop to the left consists of brown-weathering, slabby granofels and schist of the Rangeley.

2.7 Ledge in the woods west of road (to the left) is reddish-brown-weathering Rangeley schist with interlayered granofels. Layering is thin, 2-4 cm thick, with intervals up to 30-cm thick of layered granofels without schist.

3.1 Park to right before small stream gulley. Outcrops are in the woods west of the road at base of slope. (Private property.)

STOP 2. RANGELEY CALC-SILICATE GRANOFELS UNIT WITH PEBBLY LAYER. The outcrops here consist of light gray to greenish-gray, thinly layered granofels. The layers are generally 1/2 to 4 cm thick and laterally continuous. Most layers consist of medium-grained quartz-feldspar-biotite granofels with a salt-and-pepper look. Somewhat thicker light green to greenish-gray layers of calc-silicate granofels are spaced about 20 to 40 cm apart, interspersed with the gray granofels. These green and white speckled, medium- to coarse-grained calc-silicate layers contain quartz, plagioclase, diopside, and garnet. Rare seams of calcite within the calc-silicate layers weather brown. Thin quartz veins and white pegmatite stringers parallel to layering are common. One layer of quartz-rich granofels 10 cm in thickness contains round lumps of quartz less than 1 cm long that look like pebbles. Besides quartz, the layer contains 5 to15% biotite and a few percent of calcic plagioclase. The pebbly aspect is best evaluated on the outcrop, because microscopic textures are wholly metamorphic. Calcareous sandstones and feldspathic wackes are likely protoliths for these rocks. Similar rocks are present in the Sangerville Formation near Skowhegan, Maine (Pankiwskyj et al., 1976; Ludman, 1977), among other places.

We are in one of the calc-silicate units of the Hamilton Reservoir Formation (hugc) mapped in the Wales quadrangle by Seiders (1976). This unit was assigned to the granofels member of the Paxton Formation (Sp) on the Bedrock Geologic Map of Massachusetts (Zen et al., 1983), and tentatively assigned to the Southbridge Formation (mapped as SOs?) along strike to the south in Connecticut (Rodgers, 1985). The interpretation presented on this trip, based on local mapping, is that this calc-silicate unit is within the Rangeley. Whether the strata here interfinger to the east with the Paxton Formation depends on regional relationships, particularly around the Oakham anticline, that are still problematic (fig. 3). In the Southbridge Formation at Southbridge (Peper et al., 1975), renamed the Southbridge member of the Paxton Formation (Zen et al., 1983; Robinson and Goldsmith, 1991), the quartz-feldspar-biotite layers have a

purple hue, the calc-silicate layers are commonly medium to dark green and smooth-looking, they are generally finer-grained than the rocks here at Stop 2, and not associated with schist. While the rocks here may correlate with some part of the Paxton Formation, the Southbridge member is probably not the right part. Compare the rocks here with those of the Paxton Formation at Stop 11.

Return to cars. Continue north on Five Bridge Road.
3.7 Stop sign. Proceed straight across Rt. 20, onto Little Alum Pond Road.
4.7 Little Alum Pond visible to left.
4.8 Outcrop on left before driveway, of light gray, interlayered quartz-feldspar-biotite granofels and calc-silicate granofels, similar to the rocks at Stop 2.
5.0 Outcrops of brown-weathering quartz-feldspar-sillimanite-garnet Rangeley "hardschist" to the left (west).
5.4 Ridge of light gray granulites and pegmatite along right side of road.
5.9 Turn left at sawmill onto dirt road.
6.3 Pull into field to right and park. (Private property.)

STOP 3. PRE-SILURIAN GNEISSES THRUST AGAINST RANGELEY, SMALLS FALLS, AND MADRID(?) FORMATIONS. The three important things to see at this stop are: 1. Pre-Silurian gneisses (locality A); 2. Silurian strata (localities B, C, and D); and 3. The absence of a prominent mylonite zone at the contact between them (in contrast with the mylonite we will see at Stop 6).

Walk across field and into woods, following logging road to the west then north. Locality A is where road again turns west (left) at stone wall, close to Massachusetts Turnpike fence.

3A. Layered Mafic and Felsic Gneisses of the Leadmine Pond Gneiss. Among the rocks interlayered here are black hornblende amphibolite; brown-weathering, equigranular plagioclase-quartz-hornblende ± biotite ± orthopyroxene gneiss; light gray plagioclase-quartz-biotite ± orthoclase ± garnet gneiss; and white plagioclase-orthoclase-quartz-biotite-garnet pegmatite. In the hazardous turnpike cut 200 feet north of here, a hornblende-orthopyroxene-olivine-spinel ultramafic pod is enclosed among the layered amphibolites. From their compositions, igneous protoliths are inferred for all these gneisses. Whether they were originally volcanic or intrusive is less certain, but the nature of the layering here suggests a volcanic origin.

Follow road up hill. We are intentionally walking past several outcrops that we will see on the way back. Walk to the crest of the hill and continue west, staying close to the fence. Outcrop B is just across stream gulley near a jog in the fence.

3B. Brown-Weathering Schist and Granofels of the Rangeley. The rock here is similar to that at Stop 1A, except that there is a greater proportion of schist here, with the granofels layers spaced 20 to 50 cm apart. The schist, as almost every schist in the region, is full of very thin quartz-feldspar pods and stringers concordant with foliation. This outcrop clearly shows the difference between these metamorphic features and the slabby, bluish-gray granofels layers that represent relict beds. At the top of the outcrop is a late-stage, asymmetric minor fold in a granofels layer that appears to be right-handed in plan view, but is actually plunging gently to the south, with east-side-up rotation sense. Shallow-plunging minor folds of this age are scattered through the region.

Walk back toward the east, then south along the ridge crest.

3C. White Schist and Quartzite of the Small Falls Formation. This unit includes very sulfidic, graphitic, sillimanite-orthoclase-quartz-biotite-cordierite white schist interlayered with yellow-weathering, white quartzite. The schist weathers deeply to a rusty, crusty, pitted surface. The quartzite is very clean, commonly with small amounts of sillimanite, graphite, and feldspar. Due to the resistant quartzites, this unit commonly forms topographic ridges, as it does here.

To the north, this unit is mapped more or less continuously to Templeton, Massachusetts as the sulfidic schist and quartzite member of the Paxton Formation (Spsq of Zen et al., 1983). Its correlation with the Smalls Falls Formation of western Maine, first suggested by Field (1975), is based primarily on the extreme sulfide content, and the consequently distinctive mineralogy of the schists (Robinson et al.,

1982b). In a great wave of correlation through the 1980's, most schists of high sulfide content through central New Hampshire (Hatch et al., 1983, e.g.), Massachusetts (Robinson, 1981), and into Connecticut (Berry, 1985) were swept into the Smalls Falls. Unfortunately, lithologically similar sulfidic schists also occur in lenses and thin layers at other stratigraphic horizons (fig. 2), and so some of the rocks that recently have been called Smalls Falls probably belong to lower parts of the section (Berry, 1989). I now believe that most (but not all) of the thin white schist units previously assigned to the Smalls Falls in this area (Berry, 1985; Robinson et al., 1989) represent deeper stratigraphic levels. The unit Spsq at this locality is still correlated with the Smalls Falls, however, because: 1) in addition to the distinctive sulfidic white schist it contains interlayered white quartzites, as does the lower member of the Smalls Falls in western Maine (Moench and Pankiwskyj, 1988); 2) it is fairly thick by Massachusetts standards (60-80 meters) and laterally persistent; and 3) it is between rocks similar to the Rangeley (Stop 3B) and Madrid (Stop 3D) Formations. The rock at Stop 1B does not pass these criteria.

Walk carefully down the steep, rubbly slope toward the east. Locality 3D consists of several small outcrops beginning partway down the hill where the slope is less steep and continuing intermittently down the hill.

3D. Light Gray, Thinly Layered Feldspathic Granofels of the Madrid(?) Formation. The principal rock type here is a light-gray-weathering, fine- to medium-grained, quartz-calcic plagioclase-biotite-garnet ± graphite granofels in layers 1/2 to 3 cm thick. Partings between layers are richer in biotite, causing the rock to split readily into tabular slabs. The fresh rock is dark bluish-gray and vitreous. Thin layers of fine-grained, green, diopside calc-silicate granofels are present, but not abundant. Within about 10 meters of the Smalls Falls contact (not exposed) the rocks are slightly rusty-weathering and contain thin interlayers of feldspathic sillimanite-biotite-garnet schist. Thin interlayers of sillimanite-biotite-garnet schist are also present near the bottom of the hill in the outcrop closest to the Leadmine Pond Gneiss where we began our traverse (Stop 3A). Fissile sillimanite-garnet-biotite schist is rare in this unit.

Correlation with the Madrid Formation of Maine was suggested by Field (1975) from lithologic similarity and the stratigraphic sequence. In western Maine, layering is thin near the lower contact of the Madrid Formation at the type locality (Osberg et al., 1968), whereas the upper part of the type locality and the "eastern facies" are thick-bedded (Moench and Pankiwskyj, 1988; Bradley and Hanson, 1989). Relationships of the thin-bedded rock here to the rest of the Paxton Formation to the east are uncertain (fig. 3).

If correlations with the Rangeley, Smalls Falls, and Madrid are correct at this stop, then there are two major implications: 1. There is a fault between the pre-Silurian Leadmine Pond Gneiss (locality A) and the Silurian Madrid Formation at the adjacent outcrop to the west (locality D); and 2. The Siluran sequence we have just walked across from localities B, C, and D is younging to the east. Since all units dip uniformly to the west, the sequence is overturned to the east, or stratigraphically upside-down (fig. 5).

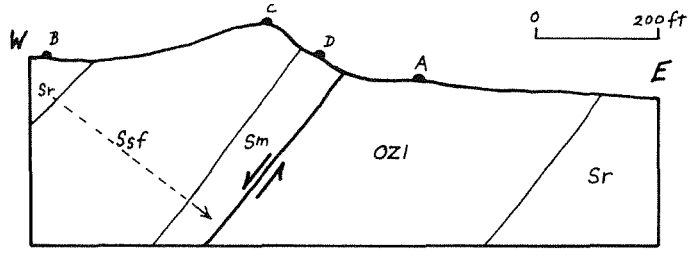


Figure 5. Geologic cross-section at Stop 3. Localities A-D are indicated. Units: OZl= Leadmine Pond Gneiss, Sr=Rangeley Fm., Ssf=Smalls Falls Fm., Sm=Madrid Fm. Dashed arrow shows younging direction inferred from stratigraphy.

Return to cars. Turn around and drive out the way we came.

6.7 Turn right onto paved road.

8.6 The rocks we are driving over here (and have been since the last stop) are Rangeley schist and calc-slicate granofels. Hillside outcrops behind the houses to the left (east) of us consist of plagioclase gneiss and amphibolite of the Leadmine Pond Gneiss. Therefore, there is a fault between us and the gneisses behind the houses. This is a nappe-stage thrust fault like the one we crossed at Stop 3 that brings older gneisses on the east up against Silurian rocks on the west.

8.9 Stop sign. Turn right on Rt. 20W.

9.1 Turn right onto old segment of Rt. 20.

9.2 Park cars.

STOP 4. TONALITE GNEISS. We are in a tonalitic pluton that cuts across mapped contacts, including the fault we saw at Stop 3. The rock is a massive, medium-gray plagioclase-quartz-biotite-hornblende gneiss with accessory apatite, magnetite, and sphene. Metamorphic pyroxene is locally abundant. Commonly, hornblende and magnetite grains form small clusters, elongate in the foliation. In addition to the common sort of white, medium-grained quartz-feldspar-biotite ± garnet dikes, the tonalite here is intruded by pegmatite dikes with red plagioclase. All the dikes were deformed together with the tonalite and display compatible foliations, lineations, folds, and boudinage.

One of the major mapping problems in this high-grade area is to distinguish metamorphosed Acadian intrusive rocks from stratified gneisses of the Leadmine Pond Gneiss. For this particular body, the mapped cross-cutting contacts demonstrate that it is intrusive. The intrusive rocks are generally massive, uniform tonalites in contrast to the Leadmine Pond Gneiss that is generally well layered, with a variety of mafic and felsic rock types and minor intercalated schist, quartzite, and calc-silicate rock. Ubiquitous apatite seems to be a common characteristic of the Acadian intrusive rocks.

The map pattern indicates that the tonalite is younger than the nappe-stage fault we crossed at Stop 3. The tonalite itself is internally deformed and was affected by the peak, granulite-facies metamorphism. Therefore, the nappe-stage fault at Stop 3 is older than the peak metamorphism.

Return to cars.
9.3 Stop sign. Turn right onto Rt. 20 W.
9.7 Roadcut on right at crest of hill contains the western contact of the tonalite body, with rusty schist and granulite to the west. Originally this was an intrusive contact, but due to subsequent deformation and high-grade metamorphism the foliation in the tonalite, the intrusive contact, and the foliation and layering in the schist are all parallel.
9.9 Roadcuts on curve, of strongly foliated, mylonitic gneiss with presumed tonalitic precursor.
10.0 Small roadcut to right (north) of interlayered fissile sulfidic schist and dark blue-gray quartz-plagioclase-biotite-garnet granofels.
10.2 Outcrops on right (north) of sulfidic schists and interlayered rusty-weathering quartzite. A thin, separately-mapped layer of white schist runs through here (fig. 2).
11.0 Roadcuts on both sides. If you look quickly, you might see very sulfidic, fissile, sillimanite-biotite-graphite schist, followed at the highest part of the outcrop by sulfidic calc-silicate granofels in blocky, thin layers and large (1 x 8 meter) pods, followed to the west by massive, light gray schist with big garnets. This gray schist can be traced northward for over 70 km into a pluton just short of the New Hampshire line (fig. 3; Zen et al., 1983). Small amounts of sulfidic white schist and black calc-silicate rocks are present intermittently along the contact.
11.4 Dangerous curve.
11.5 Turn left onto Holland Road.
11.6 Bear left toward Holland (as before).
13.3 Rattlesnake Mtn. visible ahead.
13.4 Holland town line.
13.6 Turn right onto North Wales Rd.
14.0 Park on right, safely past curve. Outcrops are to west of road. (Private property.)

STOP 5 (OPTIONAL). MEDIUM- TO COARSE-GRAINED PELITIC SCHIST OF THE RANGELEY. This is a brief stop to look at a pretty rock. We are somewhere in the midst of the Rangeley Formation, to the south and slightly west of Stop 1A. A few thin layers of dark bluish-gray granofels are present, but most of the rock consists of massive, medium- to coarse-grained sillimanite-biotite-garnet-cordierite-quartz-feldspar schist with quartz-feldspar veins and stringers concordant with the foliation. The lower part of the outcrop to the right (north) of the stream weathers brown, while most of the upper parts of the outcrop weather gray. Similar gray schists nearby have been mapped separately, although this particular one has not.

Return to cars. Continue south on North Wales Rd.
14.2 Continue straight. Now on Wales Rd.
14.5 Cross Wales town line. Road improves.
14.9 Cross Tenneco gas pipeline. Thick gravel deposit here continues northward through Brimfield center and beyond.

15.2 Stop sign. Turn left onto Rt. 19S.
15.9 Wales Market on right. Stay on Rt. 19S through Wales.
16.6 Turn left just before Lake George, onto Union Rd.
17.3 Bear left at fork.
18.4 Ridge of Rangeley red-weathering feldspathic schist on left.
18.6 Dangerous intersection. Turn right toward Stafford.
18.8 Rounded blocks of massive tonalite gneiss on slope to right.
19.2 Cross state line into Stafford, Connecticut.
19.8-20.0 Ledges to the right (west) of slabby Rangeley schist with calc-silicate granofels pods.
20.1 Outcrop in cellar hole next to boarded-up red building contains a layer of hornblende ± clinopyroxene amphibolite. A sample from here was studied by Schumacher (1986, her sample 76), that has the whole-rock chemistry of basalt. Rocks such as this with of a composition appropriate for metamorphosed volcanics are very rare in the Rangeley Formation of the Brimfield area. Scattered amphibolites are present in this belt of schist along strike to the north in central Massachusetts (Field, 1975; Tucker, 1977). Whether these rocks are metamorphosed intrusive or volcanic rocks is not known.
20.2 Slabby schist on left.
20.3 Turn left onto Burley Hill Rd. (Dead End).
20.4 Park valuable cars here. Rental cars and university vehicles may continue into the woods. Turn right onto woods road before farmhouse and proceed with care. The outcrop is within walking distance if time allows or if road conditions demand.
20.7 Park cars.

STOP 6. MYLONITE. The rock is a fine-grained, dark-gray to black mylonite with ribbons of quartz and feldspar, and porphyroclasts of plagioclase, biotite, hornblende, and sphene. The groundmass contains these minerals plus magnetite and apatite. Because of its mineralogy, the mylonite here is thought to have had a precursor of tonalite gneiss like the one we saw at Stop 4. The mylonite also contains embedded trains of quartz-feldspar rock, some of which have red plagioclase, believed to be dismembered equivalents of the dikes we saw in the tonalite gneiss at Stop 4. Rocks elsewhere in this mylonite zone are interpreted to be deformed schists of the Rangeley sort.

The mylonitic lineation plunges toward the west, approximately down the dip of the foliation. The lineation is best seen on flatter parts of the outcrop to the right. Work around the outcrop, circling toward the left, to an east-west striking vertical face that is approximately parallel to the lineation. There, asymmetric fabrics and rotated porphyroclasts consistently indicate a west-over-east shear sense. Although rotated porphyroclasts (delta grains) are by far the most common, there are also some rotated grains which have recrystallized asymmetric tails superimposed on them (complex delta-sigma grains of Passchier and Simpson, 1986). Farther around the outcrop, on the overhanging surface, dismembered bits of a red-plagioclase pegmatite outline a ghostly west-side-up asymmetric fold. This fold is tentatively thought to be younger than the mylonite formation.

Mylonite at this outcrop was shown on the map by Seiders (1976) who interpreted a west-side-up reverse fault through here. Finkelstein (1987) looked at some of the microstructures in this mylonite zone, which corroborate a west-side-up shear sense on the down-dip lineation. Berry (1989) mapped this zone of mylonitic rocks for 2 1/2 km along strike, with a maximum width of about 100 meters.

The relationships deduced here indicate that the age of mylonite formation post-dates the peak of metamorphism and tonalite emplacement. These mylonites and related kinematic indicators are therefore not related to the fault we saw at Stop 3, because that fault is older than the cross-cutting tonalite and the peak metamorphism (Stop 4). The early faults have been assigned to the west-directed nappe-stage of Acadian deformation (Robinson et al., 1991). The east-west trending, down-dip lineation and associated folding and mylonitic deformation, such as we see here, are found throughout central Massachusetts and have been assigned to the east-directed backfold stage of Acadian deformation (Robinson, 1979).

Return to cars. Turn around and drive back the way we came.
21.0 Turn left onto dirt road.
21.1 Turn right (north) onto New City Rd.
22.7 Enter Massachusetts.

22.8 Stop sign. Please be careful. The crossing road does not have a stop sign! Turn right.
23.5 Entering Union, Connecticut.
23.7 Massive tonalite gneiss outcrops in yard to the left (east).
24.7 Slabby, red-weathering Rangeley schist to the left.
25.6 Continue straight
26.1 Turn right at triangular intersection onto Staffordville Rd.
26.5 Rusty road cut on right contains felsic gneiss of the Leadmine Pond Gneiss. Parts of this outcrop have high proportions of orthoclase (~50%) and garnet (~25%) in addition to quartz, plagioclase, sillimanite, and biotite. It may represent either an unusual, perhaps altered protolith or else some type of restite. Adjacent outcrops to both sides include more normal kinds of mafic and felsic Leadmine Pond gneisses.
26.6 Bear left.
27.6 Turn left onto Webster Rd.
27.9 Pavement ends.
28.2 Roadcut to the left and stream exposures to the right of Leadmine Pond Gneiss. Some sillimanite-bearing gneisses and dark-gray, garnetiferous calc-silicate pods are present here.
28.3 Park on right. We will be partly on private property and partly in the Nipmuck State Forest.

STOP 7. SYNCLINE OF WEBSTER ROAD. This stop will entail a traverse through several units (fig. 6A). We will again cross an early, nappe-stage, pre-peak-metamorphic fault that brings the Leadmine Pond Gneiss to the east against parts of the Rangeley Formation to the west. The main features to see at this stop are: 1. Thin units of garnet quartzite and amphibolite in the Leadmine Pond Gneiss (localities B and C); 2. A sequence of brown-weathering schist, calc-silicate granofels and well-layered gray schist in the Rangeley (localities D, F, and G); and 3. Repetition of the calc-silicate unit about the gray schist, suggesting an isoclinal syncline that pre-dates the fault (locality H).

Leave road on obscure trail 30 feet east of telephone pole 1899. Follow trail to the south. Cross footbridge and walk up gentle incline to pavement outcrops in the trail (about 5 mins. from road).

7A. Felsic Gneiss of the Leadmine Pond Gneiss. The two outcrops in the trail show a type of strongly foliated, yet poorly layered felsic gneiss that is common in the Leadmine Pond. It is a light gray to white, medium-grained quartz-plagioclase-biotite gneiss with small, pinhead garnets. Thin, biotite-rich folia give the rock a striped appearance, but the folia do not form continuous layers. In some places, layers with slightly different shades of light gray can be distinguished. The rocks here are similar to parts of the Monson Gneiss of the Bronson Hill zone. Coarser-grained gneisses and pegmatites off the trail to the west (optional) contain coarse-grained orthopyroxene.

Continue south on trail to a fork. Take the trail less traveled by, and follow it down the slope to the right (southwest). Outcrops of interest are 10 meters off the trail to the left (south), just before the trail bottoms out in a swampy gulley.

7B. Garnet Quartzite of the Leadmine Pond Gneiss. The rock is a yellow- to red-weathering, watery-gray, garnet quartzite interlayered with milky-gray feldspathic, sulfidic garnet quartzite. The layers are 5 to 15 cm thick, and are believed to represent beds. The conspicuous 1-3 mm garnets make up commonly 10% and up to 50% of the rock. The protolith is uncertain, but might be a metamorphosed cherty sediment. The presence of interlayered sediments suggests that the Leadmine Pond Gneiss is a stratified unit, and that the gneisses may have been volcanic rather than intrusive rocks.

Continue on trail across the gulley (40 feet?) then walk north (to the right) about 150 feet to hillside outcrops west of the gulley.

7C. Amphibolite of the Leadmine Pond Gneiss. The outcrops along this small ridge are dominated by mafic gneisses including amphibolite and hornblende-biotite-plagioclase-quartz gneiss, with lesser amounts of quartz-feldspar-biotite gneiss.

Return to trail. Follow trail as it heads west over the small ridge, then turns south. We have left the Leadmine Pond Gneiss behind us, and are walking approximately on the trace of the fault that separates it from Rangeley rocks on the cliff to the west. Continue walking south along the trail

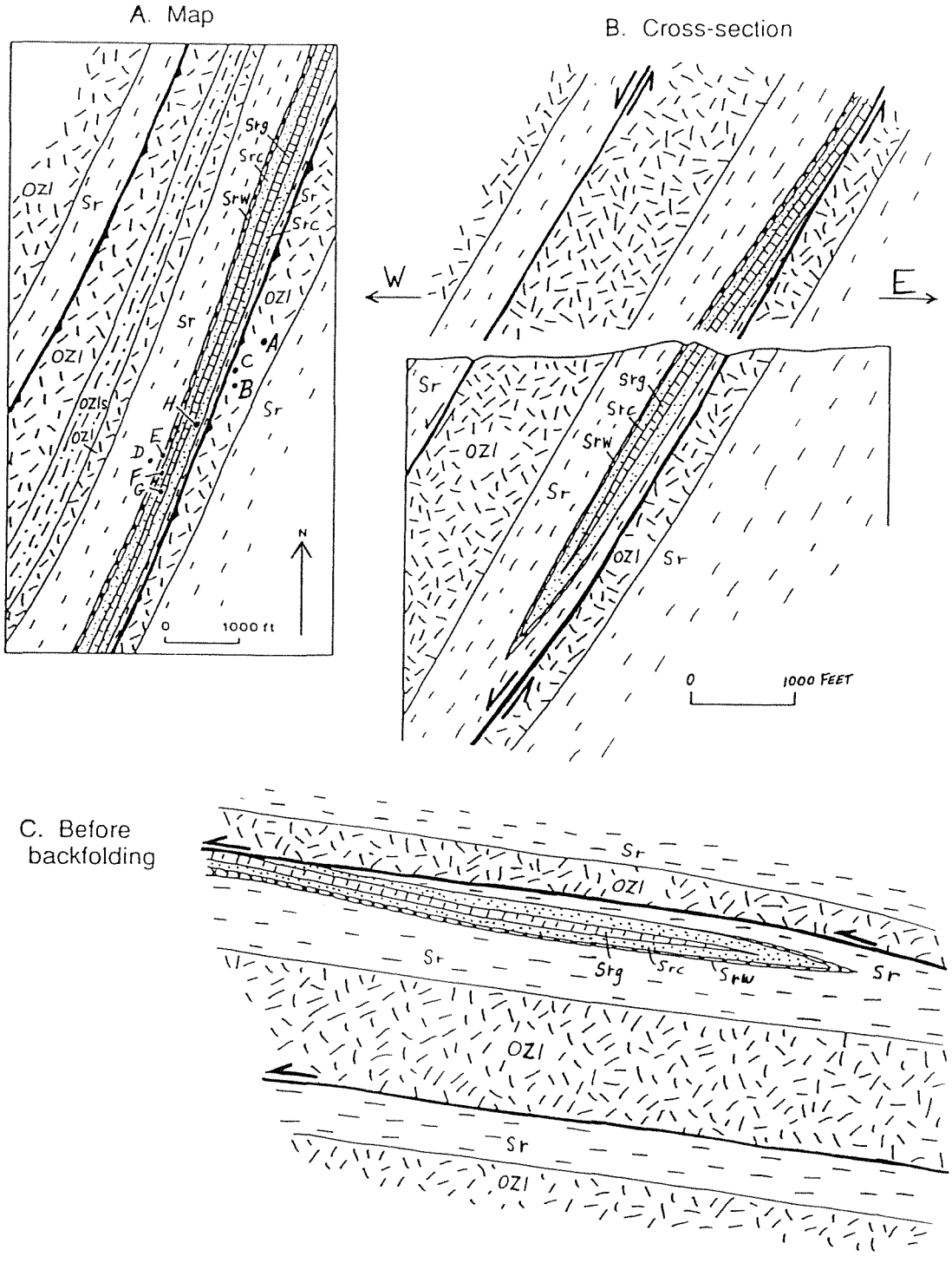


Figure 6. Geologic relationships at Stop 7. A. Geologic map. Localities A through H are indicated. Repetition of units about "Srg" reflects an isoclinal syncline. B. Cross-section showing present geometry, with isoclinal fold truncated by nappe-stage thrust fault. C. Inferred geometry at the time of thrusting, with Leadmine Pond Gneiss (OZl) thrust westward over younger, previously folded rocks. Leadmine Pond Gneiss units: OZl=undifferentiated gneiss, OZls=schist unit. Rangeley Fm. units: Sr=undifferentiated schist, Srw=white schist, Src=calc-silicate unit, Srg=gray schist.

in the shadow of the cliff for about 10 minutes. Then head west into area of abundant outcrop at the south end of the ridge. Please stay with the group. The intended outcrops will be difficult to find on your own. We will work through the outcrops from west to east in what is believed to be ascending stratigraphic order.

7D. Rangeley Schist and Granofels. The red- to brown-weathering, feldspathic sillimanite-biotite-garnet schist here is similar to rocks in other strike belts to the west (such as at stops 3B and 5), and is likewise assigned to the Rangeley.

Go east across logging road to small, barely-exposed outcrop under laurel bush.

7E. Rangeley White Schist. A small excavation project by Peter Robinson and me uncovered this occurrence of sulfidic white schist sandwiched between Rangeley schists to the west and a slabby calc-silicate granofels unit to the east. Although the contacts are not exposed, they are closely constrained by outcrop to within 10 meters, so the unit is certainly less than 20 meters thick, and probably less than 5.

Continue east a few feet to the next outcrop.

7F. Calc-silicate Granofels Unit. The rocks consist of medium- to coarse-grained, quartz-plagioclase-biotite granofels interlayered with green-speckled diopside-quartz-feldspar-biotite±garnet calc-silicate granofels. The green calc-silicate layers have pinch-and-swell structure, and isolated boudins are locally present. Small pegmatite pods are common. Sillimanite-bearing schist is virtually absent from this unit.

Follow abundant low outcrops of the calc-silicate unit to the southeast, diagonally across strike to a large outcrop sticking out of the south end of the hill.

7G. Gray Schist Unit. This unit contains coarse-grained, gray-weathering, pelitic schist interlayered with quartz-rich granofels. The schist, dominated by quartz, sillimanite, and biotite, contains abundant garnets ranging from 3 to 30 mm across. In some places the schist and light gray quartz-rich granofels alternate in beds 5 cm thick.

Walk down hill to the east to join the trail. Follow trail to the left (north). Partway back along the trail, go west to the base of the cliff.

7H. Calc-silicate Granofels Unit. We are again in the calc-silicate granofels unit, but now east of the gray schist. At locality 7F, this calc-silicate granofels unit was to the west of the gray schist. Both contacts of the gray schist unit, though not easily accessible, are exposed in the cliff above us. Repetition of the calc-silicate unit about the gray schist suggests that we have crossed the axial surface of an isoclinal fold (fig. 6B). The axial surface of this fold must trace on the ground through the gray schist unit. Whether this fold is a syncline or an anticline depends on the stratigraphic facing direction. If the sequence youngs toward the gray schist unit, as currently believed, this fold is an overturned syncline.

Along strike to the north of Stop 7, the calc-silicate granulite and gray schist units are gradually cut out by the fault to their east (fig. 2). Therefore, the syncline here is older than the fault. Both the syncline and the fault are believed to be nappe-stage features that were originally west-directed, and have been overturned by backfolding into their present orientation (fig. 6C).

Return to cars. Continue east on Webster Rd.

28.8 Series of low, flat outcrops to the right are made of brown- to rusty-weathering schist and granofels of the Rangeley Formation. In some places here and elsewhere on Bald Hill, the layering is a little thinner than average, about 1 to 3 cm thick. So far, rocks with this bedding style have not been mapped separately in this area.

29.0 Pavement begins.

29.3 Stop sign. Turn right and go about 30 feet to second stop sign.

29.31 Turn left onto Rt. 190E.

29.7 Continue straight toward I-84E.

29.8 Turn left onto I-84E.

31.1 Roadcuts on both sides, of rusty-weathering, crumbly, graphitic schist and interlayered quartzites of the Leadmine Pond Gneiss. The road exploits this unit for the next 1.8 miles.
32.1 The western contact of the rusty schist unit is exposed partway up the roadcut to the left (west), where light gray quartz-plagioclase gneisses with minor amphibolite layers rest above the rusty schist. Layering dips away from us (toward the west) at about 50 degrees.
32.6 In the large roadcut to the left the gneiss unit contains a prominent, 3-foot-thick black amphibolite layer, about 10 feet above the schist contact. The amphibolite layer is parallel to the schist-gneiss lithologic contact, indicating that the schist, gneiss, and amphibolite are concordant.
32.9 Road curves to the right. Dip-slope outcrops to the right contain rusty quartzite layers in schist.
33.5 Take Exit 74 toward Rt. 171, Union, Connecticut.
33.7 Stop sign. Turn left, crossing over interstate.
33.8 Turn left.
33.9 Go straight.
34.0 Turn left onto Moore Rd.
34.1 Park at end of road, by barrier. Walk into abandoned rest area.

STOP 8 (OPTIONAL). LEADMINE POND GNEISS. This is a brief stop to see two of the rock types in the Leadmine Pond Gneiss in the same strike belt as the type area at Leadmine Pond, Sturbridge (fig. 2).

8A. Layered Mafic and Felsic Gneiss. Under the old Ashford Motel billboard is a good exposure of thinly layered amphibolite and plagioclase gneiss.

Walk to the southwest about 500 feet to hillside outcrop near bottom of east-facing slope.

8B. Rusty Schist and Quartzite of the Leadmine Pond Gneiss. The schist is a rusty-weathering, feldspathic sillimanite-biotite-graphite schist that weathers rustier and is less aluminous than most Rangeley varieties, although similar schists can be found in the Rangeley. The main reasons that this schist is assigned to the Leadmine Pond Gneiss are a) because brittle white quartzite beds are interlayered with the schist, and b) this rusty schist unit is flanked to both sides by plagioclase gneiss and amphibolite.

Return to cars. Turn around.
34.2 Stop sign. Turn right.
34.4 Stay straight.
34.5 Turn right.
34.6 Stop sign. Turn right toward Union, Connecticut.
35.9 Turn right onto Cemetery Rd.
36.3 Park on right. (Private property.)

Stop 9. Leadmine Pond Gneiss. The objective of this stop is to see some of the rock types found together in the Leadmine Pond Gneiss here, and compare them with rocks in the Leadmine Pond Gneiss south of Webster Road (Stop 7). We have moved a short distance across strike to the east from Stop 8, still in the same belt of Leadmine Pond Gneiss (fig. 2).

Walk up hill to west.

9A. Felsic Gneiss. Take a passing look at small outcrops near the top of the slope to see that they are foliated plagioclase-quartz-biotite gneisses.

Walk 500 feet north across fairly level ground to small, rocky knob.

9B. Rusty-Weathering Feldspar-Graphite Schist. This poorly exposed unit contains deeply rusty-weathering, crumbly, graphitic rocks. In some places it has sillimanite-rich folia, with pale biotite or sparse garnet, but in most places garnet and biotite are rare, sillimanite is sparse, and the predominant minerals are quartz, feldspar, and graphite. It is coarse-grained, with K-feldspar megacrysts. Graphite is ubiquitous, occurring in masses up to 5 mm across.

Walk 300 feet southwest to long, low outcrops sticking out of the south side of the slope.

9C. Mafic Gneiss Unit. This unit contains layered medium- to dark-gray gneisses. The most distinctive type is a foliated, streaky hornblende-biotite-clinopyroxene-plagioclase gneiss with small clumps of hornblende (± pyroxene) grains. Thinly layered biotite-plagioclase ± hornblende ± clinopyroxene schist is also common. A subordinate amount of leucocratic plagioclase gneiss is present in this unit as well. This unit is similar to the one we saw at Stop 7C.

Walk 100 feet southwest to hillside outcrop and rubble.

9D. Garnet Quartzite. Yellow- to red-weathering, layered quartzite peppered with garnets. Some layers contain thin boititic laminae. Small amounts of graphite are present. This unit is believed to be the same one we saw at Stop 7B.

The units here in the Leadmine belt are similar to the units south of Webster Road (Stop 7). A unit-for-unit correlation seems warranted for the following sequence: 1. Layered, predominantly felsic gneisses (stops 7A, 8A); 2. Garnet quartzite (stops 7B, 9D); and 3. Mafic gneiss and amphibolite (stops 7C, 9C). Other units of the Leadmine Pond Gneiss are present in the Leadmine belt (such as stops 8B, 9A, and 9B), but are presumably cut out by faulting near Webster Rd. Although the younging sense of this sequence is not known, it faces in opposite directions in the two belts.

The similarity of Leadmine Pond Gneiss units in two different strike belts is compelling evidence that there is a coherent lithostratigraphy in the Leadmine Pond Gneiss, and that the same units are structurally repeated on the ground. This is believed to represent the pre-Silurian (Precambrian ?) basement of the Merrimack belt at this latitude (Berry, 1988). Similar sequences of thin units are present in parts of the Putnam Group (Connecticut), Nashoba Formation (Massachusetts), and the Cushing Formation in the Falmouth-Brunswick sequence (Maine) (fig. 1).

Walk east and return to cars. Turn around and retrace route to the east on Cemetery Rd.

36.7 Stop sign. Turn left.
38.0 Turn left toward I-84.
38.1 Turn right onto I-84E.
38.4 Large roadcut along southbound exit ramp (to left) contains various rocks of the Leadmine Pond Gneiss.
38.6 Enter Massachusetts.
39.1-39.6 Large roadcuts of rusty-weathering schist and slabby, light purplish and greenish granofels of the Paxton Formation.
41.4 Outcrop to the right at north end of weigh station is purple Paxton granofels with white, popcorn pegmatites.
41.9 Take Exit 1, Mashapaug Rd., Southbridge.
42.0 Stop sign at end of exit ramp. Turn right.
42.3 Pass under I-84E.
42.7 Pass under I-84W.
43.1 Pull off to right, near dirt road. Walk up dirt road to outcrop in back of dumping area.

STOP 10 (OPTIONAL). EASTERN CONTACT OF LEADMINE POND GNEISS. We have seen two places where the western contact of the Leadmine Pond Gneiss is faulted against the Rangeley Formation (stops 3 and 7), based on truncation of units and inferred stratigraphic offset. In contrast, the eastern contact of the Leadmine Pond Gneiss is interpreted to be an unconformity based on the persistence of thin units in the basal parts of the Rangeley in the Brimfield-Sturbridge area (Berry, 1989). A thinly layered calc-silicate granulite unit that occurs at several places along the eastern contact of the Leadmine Pond Gneiss is interpreted to be at the base of the Rangeley Formation. It is assigned to the base of the Rangeley rather than the top of the Leadmine Pond Gneiss because similar rocks are also found higher in the Rangeley. The objective of this stop is to see this basal calc-silicate unit of the Rangeley.

10A. Base of the Rangeley. The rocks from west to east in the outcrop include: interlayered plagioclase gneisses, granitic gneisses, hornblende-biotite gneisses and amphibolites of the Leadmine Pond Gness; thinly interlayered light gray quartz-feldspar-biotite granofels and medium- to light-green diopside calc-silicate granofels of the basal Rangeley; and rusty-weathering, migmatitic sillimanite-garnet-biotite schist with interlayered pegmatites of the Rangeley.

Unfortunately, there is no evidence at this outcrop that would indicate the presence of an unconformity. It is by mapping along this contact and by considering the regional stratigraphic relationships that an unconformity is proposed. The rocks here, as elsewhere, have been strongly deformed and metamorphosed. West-side-up asymmetric fabrics of the backfold stage are well developed, especially in the migmatitic schist.

Return to road. Drive or walk ahead to:
43.3 Roadside outcrop on banking to the right.

10B. Rangeley Schist and Granofels. This outcrop of rusty-weathering sillimanite-garnet-biotite schist with interlayered feldspathic granofels is representative of a thin belt of rocks lying east of the Leadmine Pond Gneiss and west of the Paxton Formation (Stop 11). From lithologic similarity with rocks to the west, and from their apparently similar stratigraphic position above the Leadmine Pond Gneiss, the rocks here are assigned to the Rangeley.

Turn around and retrace route to the north.
43.9 Pass under I-84W.
44.0 Abandoned "Sturbridge Isle" service area on right. An exposure at the south end of the parking lot contains light gray, coarse-grained sillimanite-biotite-garnet-cordierite schist interlayered with quartz-feldspar-biotite-garnet granofels (Stop 6 of Thomson et al., this volume). Similar rocks on strike to the southwest were mapped as a member of the Bigelow Brook Formation (Peper et al., 1975). In the Westford quadrangle, Connecticut, the Kinney Pond fault separates the Bigelow Brook from the Hamilton Reservoir Formation to the west (Peper and Pease, 1975). It is not known whether the Kinney Pond fault extends northward into Massachusetts. If so, it would lie between stop 10 and here, approximately under the westbound lane of I-84. This gray schist and granulite unit was assigned to the Littleton Formation by Zen et al. (1983).
44.3 Outcrop of Paxton granulite and pegmatite lurking in the shadows to the right.
44.4 Pass under I-84E.
44.7 Turn left onto I-84E.
46.1 Take Exit 2, to 131 Sturbridge, Southbridge.
46.3 Stop sign. Turn left.
46.5 Stop sign. Turn left.
46.7 Park near telephone pole with "Shattuck Rd" sign.

Stop 11. Layered Granofels of the Paxton Formation. The main objective at this stop is to see an example of layered granofels characteristic of the Paxton Formation for comparison with the granofels at stops 2 and 7F. The quartz-plagioclase-biotite±garnet granulites have a distict purplish hue caused by the fine-grained reddish biotite. These purplish-gray granulites are interlayered with smooth-looking, fine-grained, green, diopside calc-silicate granulites. Conformable white pegmatite layers are common. Dome-stage minor folds deform the layering and pegmatites together, with hinges plunging about 10° toward the south and axial surfaces dipping shallowly to the west. A large pegmatite body at the southwest end of the outcrop contains enclaves of sillimanite-biotite schist with a west-plunging, backfold-stage lineation. Muscovite is present in this and other pegmatite bodies in the outcrop. The muscovite is thought to be of retrograde origin because it appears to surround sillimanite in some places, but the petrology of these rocks has not been studied.

The rocks here have been assigned to the Bigelow Brook member of the Paxton Formation (Zen et al., 1983, Robinson and Goldsmith, 1991). Sulfidic schist, which is common in the Bigelow Brook, is exposed at the northeast end of the outcrop and along the southbound entrance ramp. We are approximately on strike with the type locality of the Paxton Formation at Paxton Falls, Massachusetts to the north. Although the purple and green granofels that we see here is similar to the rock at Paxton Falls, the sulfidic schist here is not present at the type locality, suggesting intervening structural and stratigraphic complications (fig. 3).

End of trip.

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MOTION ON THE CLINTON-NEWBURY AND RELATED FAULTS AND MULTIPLE DEFORMATION OF THE MERRIMACK GROUP IN EASTERN MASSACHUSETTS: ASPECTS OF THE ALLEGHANIAN OROGENY IN SOUTHEASTERN NEW ENGLAND

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INTRODUCTION

The purpose of this trip will be to examine exposures of Carboniferous and pre-Carboniferous rocks in eastern Massachusetts to determine the timing of deformation in this region. The conclusions the author has reached and that he will attempt to convince you of are:

1. Carboniferous rocks contain two prominent cleavages and pre-Carboniferous rocks immediately adjacent contain three prominent cleavages. Thus, the formation of the final two cleavages in pre-Carboniferous rocks must be an Alleghanian (Permian) event.

2. The Alleghanian cleavages exist in a well defined zone 15 km. wide which has a major fault (Clinton-Newbury) as its eastern boundary and a western boundary which has steep mylonites locally developed along it. This wide zone is interpreted as a shear zone and the deformation and cleavage formation within it is interpreted in terms of motion along the shear zone. This demands early Alleghanian sinistral displacement suggesting an early Alleghanian northerly oriented shortening.

3. Motion on the Clinton-Newbury fault pre-dated the last Alleghanian cleavage but postdated the first. Thus, motion along this fault is also an Alleghanian event. Microscopic kinematic indicators and regional geologic relationships indicate that the Clinton-Newbury fault is a normal fault with a small sinistral component.

4. Motion on the Clinton-Newbury fault was preceded by motion on a low angle mylonite zone, herein named the Wachusett Mylonite Zone. The zone of mylonitization associated with this structure is several kilometers thick, is truncated by the Clinton-Newbury Fault and displays evidence of normal displacement.

These conclusions are, to some extent, speculative and it is hoped that the trip participants as well as those who may follow this trip on their own will be able to use the field trip stops to draw their own conclusions.

REGIONAL GEOLOGIC SETTING

The area covered by this trip includes, for the most part, the Merrimack trough. This region exposes rocks of unknown age and unknown affinity with the other tectonic zones of New England. The Merrimack trough rocks form a belt which includes much of eastern Massachusetts and southeastern New Hampshire and Maine (Figure 1). The Merrimack trough is bounded on the southeast and east by the Clinton-Newbury fault (Skehan, 1967) which separates it from the Nashoba terrane and contains a sequence of metasedimentary rocks composed of phyllites and a variety of quartz-rich rocks. In eastern Massachusetts these include the Tower Hill Quartzite, a nearly pure quartzite of local extent, the Oakdale Formation, a series of calcareous quartzites and siltstones, and the Worcester Formation, a series of phyllites and associated non-calcareous quartzites. Some would also include the Paxton Formation, a pelitic schist, which is in contact with the Worcester Formation. Also of controversial association is the Vaughn Hills Formation, a thinly laminated metasiltstone of local extent and with very poorly known contact relations with surrounding rocks. The Merrimack group is intruded by the Ayer granite,dated at 430 Ma by Zartman and Naylor (1984) and by the Fitchburg plutonic complex dated at 400 Ma also by

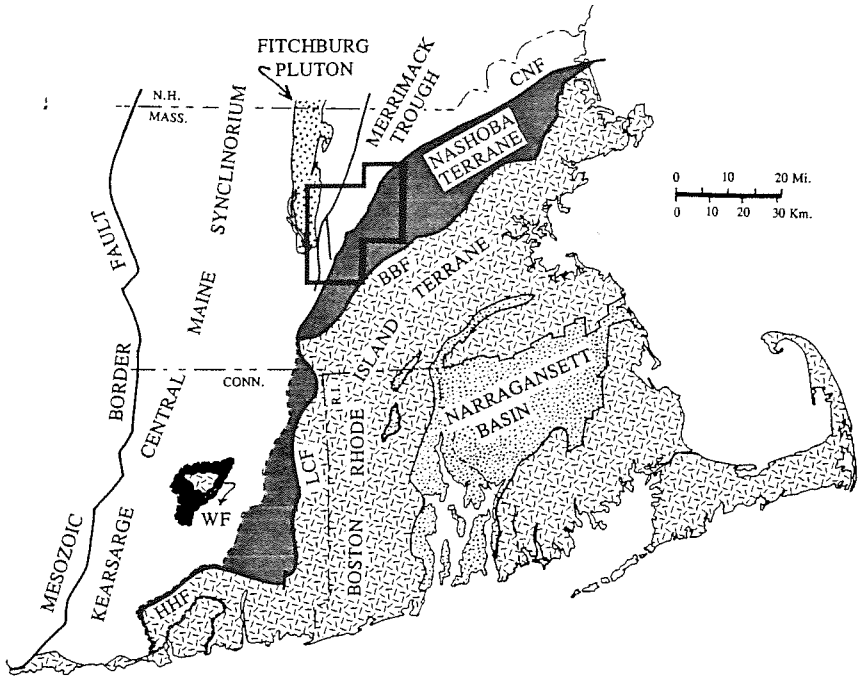

Figure 1. Generalized tectonic map of southeastern New England showing area studied. CNF = Clinton-Newbury fault; BBF = Bloody Bluff fault; LCF = Lake Char fault; HHF = Honey Hill fault; WF = Willimantic fault.

Zartman and Naylor (1984). In New Hampshire and southeastern Maine, the Merrimack trough is divided into the Eliot, Berwick and Kittery formations and, although considerable similarities exist between the two stratigraphies, no accepted correlation has been proposed.

East of the Clinton-Newbury fault lies the Nashoba terrane which contains rocks so strikingly different from those in the Merrimack trough that no connection is believed to exist between the two regions (Hepburn et al, 1987). The Nashoba terrane consists of a sequence of diverse metasedimentary and mafic metaigneous rocks all at very high metamorphic grade. The mafic schists and gneisses and quartzofeldspathic gneisses are of undetermined age but must be older than the Sharpners Pond diorite (446 Ma, Zartman and Naylor, 1984) which intrudes them.

The Clinton-Newbury fault in the area of Clinton, Mass. splits into two separate fault strands (Figure 2). The block which lies between the two strands is very similar to the Nashoba terrane, but is sufficiently different that Hepburn and Munn (1984) named it the western Nashoba block. This block contains biotite gneisses and muscovite-biotite gneisses similar to those in the Nashoba terrane, sillimanite-bearing schists, amphibolites and a reasonably large granite body named the Rocky Pond granite by Hepburn and Munn (1984). Zartman and Naylor (1984) had great difficulty obtaining a well-constrained age for the Rocky Pond granite (which they referred to as the muscovite granite at West Berlin), but claimed that the best interpretation was that the age of crystallization was 448 Ma and that the isotopes were disturbed by metamorphism at 367 Ma (with an error of plus/minus 97 Ma). The Rocky Pond granite contains xenoliths of the Vaughn Hill formation. If the Vaughn Hills belong with the Merrimack trough, as the author believes, then the Merrimack trough and the Nashoba terrane must have been juxtaposed sometime before 448 Ma. If, however, the Vaughn Hills is associated with the Nashoba terrane, a view expressed by Ashenden (personal communication, 1992), then the timing of juxtaposition of the two terranes cannot be constrained.

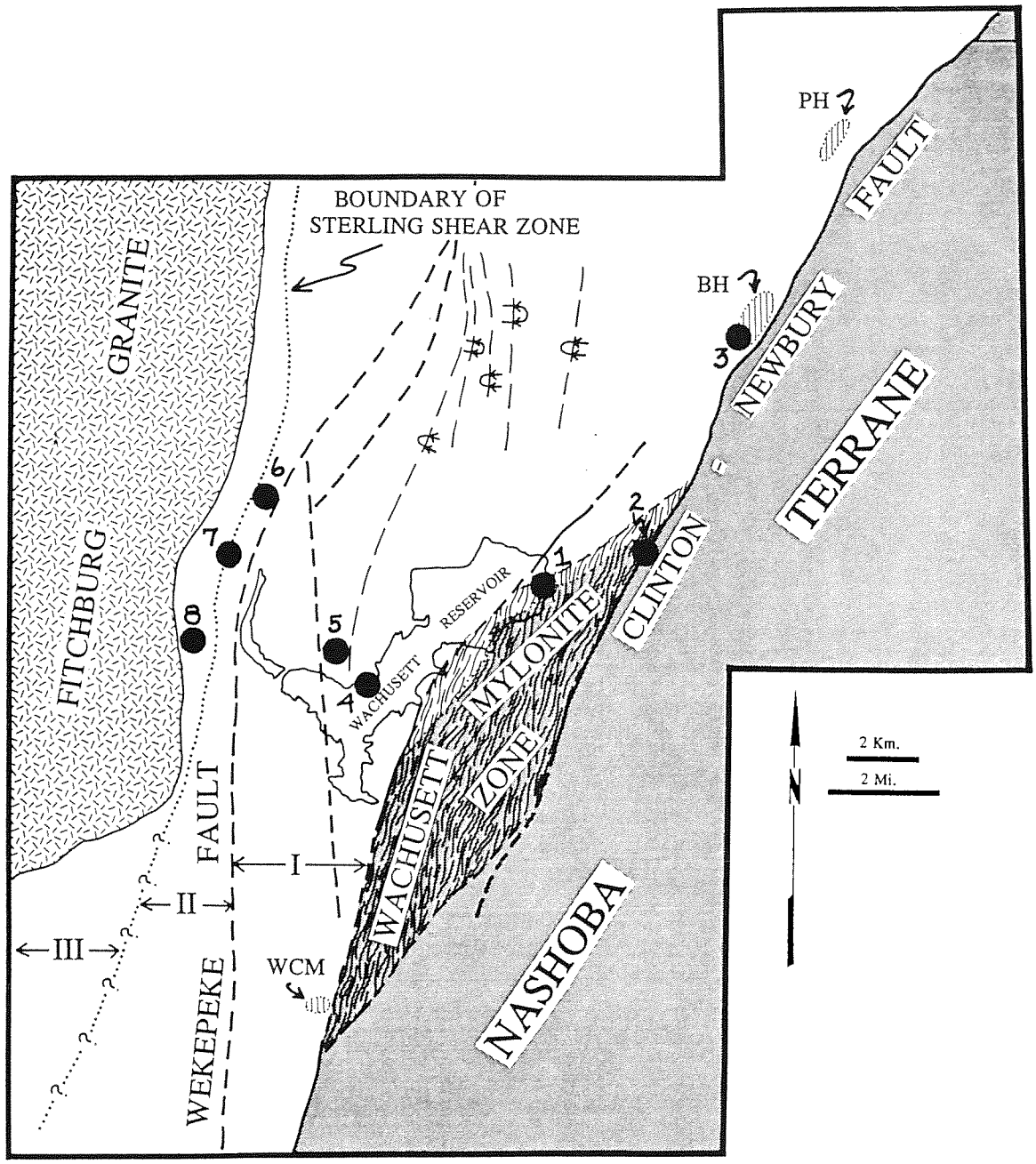

Figure 2. Tectonic map of the Worcester, Massachusetts area shown on Figure 1. Rocks of the Merrimack trough are unpatterned except in the Wachusett Mylonite Zone where they are given patterns indicating mylonitization. Carboniferous rocks are shown with a vertical rule pattern in three locations: PH = Pin Hill; BH = Bare Hill; WCM = Worcester Coal Mine. Roman numerals refer to structural domains discussed in the text and shown in Figure 4. Field trip stops are numbered.

STRUCTURAL GEOLOGY

Wachusett Mylonite Zone. The Wachusett Mylonite Zone (WMZ) occupies much of the western Nashoba block of Hepburn and Munn (1984). The top of the zone lies in the Carville Basin (Stop 1) within rocks of the Merrimack Trough. Most of the mylonites, however, are developed from rocks of the western Nashoba region (Figure 2). Within the WMZ, mylonites have very similar orientations, striking northeast and dipping moderately northwest, with elongation lineations trending to the northwest (Figure 3). Not every exposure of rock in this region is mylonitized, rather the structure of the zone consists of blocks of unmylonitized rock surrounded by shear zones. This is especially true in the two intrusive units within the WMZ, the Ayer granite and the Rocky Pond granite. The metasedimentary and metavolcanic rocks within the zone, however, appear to be thoroughly and pervasively mylonitized. Numerous oriented thin sections display unequivocal kinematic indicators which show that the WMZ moved as a normal fault. Some small regions adjacent to unmylonitized blocks display thrust sense indicators and this is interpreted as due to the patterns of flow around these resistant blocks. Conditions of deformation were such that quartz behaved ductily whereas feldspar behaved brittly, consistent with lower to middle greenschist facies conditions. The mylonites of the WMZ are truncated by the Clinton-Newbury fault.

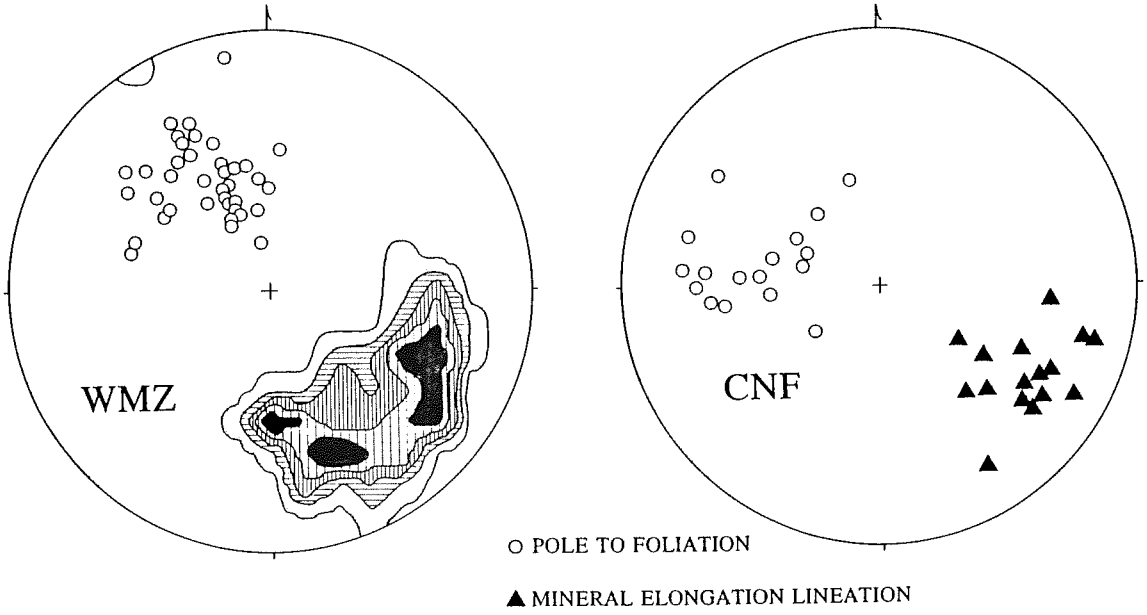

Figure 3. Lower hemisphere, equal area stereographic plots of structural data for the Wachusett Mylonite Zone (WMZ) and the Clinton-Newbury Fault (CNF). Mineral elongation lineations are shown as open circles and poles to mylonitic and phyllonitic foliation are contoured for the WMZ (contours 2, 4, 6, 8, 10, 12% per 1% area; 46 data points).

Clinton-Newbury fault. The Clinton-Newbury fault (CNF) is a very steeply dipping structure which can be traced from the coast of Massachusetts around Newbury southeastwards to Clinton and Worcester, Massachusetts. The fault dips steeply to the west or northwest and contains rocks of high metamorphic grade in the footwall and rocks of lower metamorphic grade in the hangingwall. Isolated inliers of Carboniferous rocks exist only immediately adjacent to the

fault in the hangingwall (Pin Hill in Harvard, Bare Hill in Bolton and the Worcester coal mine and related deposits in Worcester). All these regional geologic relationships suggest that the fault moved as a normal fault. Microscopic kinematic indicators in mylonites and phyllonites support these regional relationships. The CNF mylonites dip more steeply than those of the WMZ (figure 3), although there is considerable overlap in foliation orientations, but mineral elongation lineations in CNF mylonites and phyllonites invariably trend due west (Figure 3), indicating that the fault had a slight sinistral component along with the normal motion.

Pre-Silurian Rocks. Rocks of the Merrimack Trough display three prominent planar fabrics (cleavage, foliation) at the mesoscopic scale (Figure 4). The earliest is always parallel to bedding and is interpreted to be associated with isoclinal folding. Such folds have not been unequivocally identified, at any scale, however. The second prominent planar fabric is a crenulation cleavage which strikes north to northeast and dips steeply to the west or northwest. It is parallel to the axial planes of folds which exist at both the mesoscopic and macroscopic scale. Chlorite, which postdates the formation of high grade metamorphic minerals (andalusite, staurolite and garnet) has grown during the formation of this cleavage. Some chlorite grains clearly predate the cleavage whereas other grains in the same thin section clearly postdate the cleavage. This cleavage and associated folds have never been observed affecting mylonites of either the WMZ or CNF, but do affect rocks immediately adjacent to these faults. A third cleavage, also a crenulation cleavage, is commonly, though not ubiquitously, found in rocks of the Merrimack trough. This fabric is subhorizontal (Figure 4) and can be associated with kink-style folds with amplitudes of .5 meters, but rarely larger and more commonly smaller. There is a fourth cleavage which has been found only in a few locations and only affecting small volumes of rock. This subvertical fabric strikes WNW.

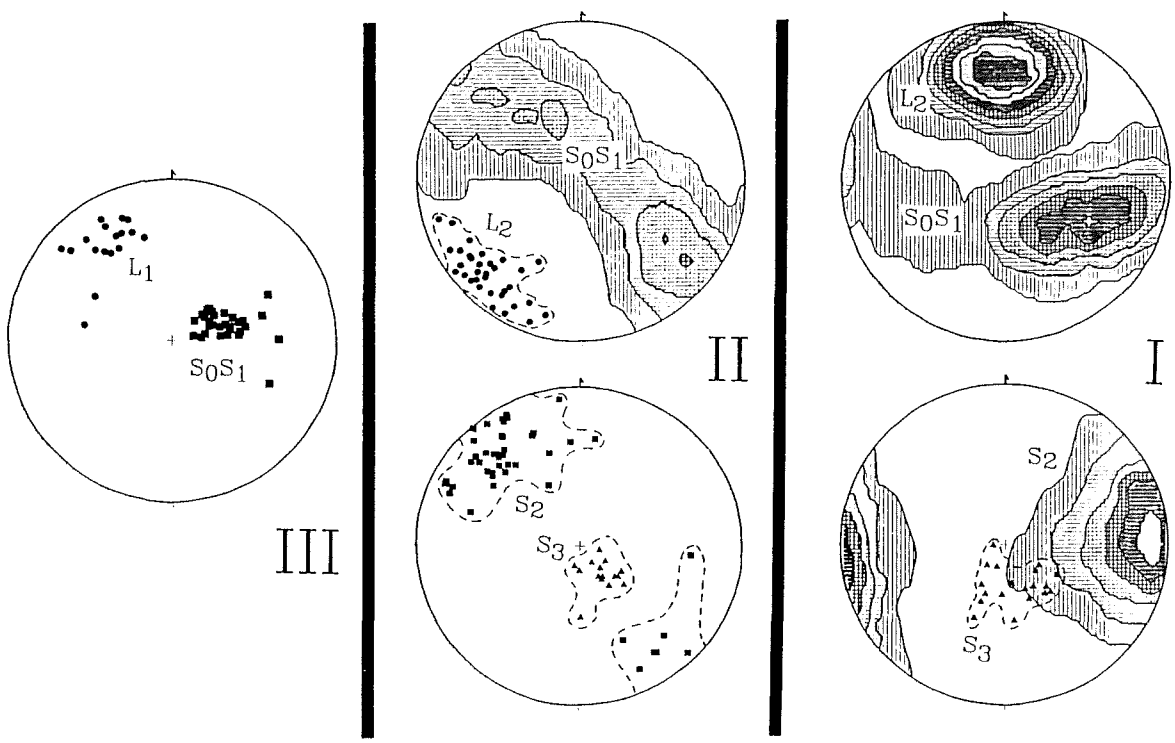

Figure 4. Lower hemisphere, equal area stereographic plots of structural data from three domains in the study area. All contours are in intervals of 3 sigma using the Kamb method. S_0S_1 are poles to bedding and pervasive bedding parallel foliation. L_1 are biotite and, locally, andalusite mineral alignment lineations. S_2 are poles to crenulation cleavage axial planar to D_2 folds. L_2 are intersection lineations and fold axes of D_2 folds. S_3 are poles to subhorizontal crenulation cleavage.

Carboniferous Rocks. Carboniferous rocks exist in a few small locations, always with poor contact relationships to surrounding rock. Most are believed to be in fault contact on all or most of their boundaries (Thompson and Robinson, 1976; Grew, 1973) In all localities, the interbedded metaconglomerates and phyllites display two cleavages, one parallel to bedding and a later crenulation cleavage cutting bedding at a high angle. No folds, other than small amplitude warps, have been found which are associated with these two cleavages. Phyllites in Carboniferous rocks contain chloritoid which grew during the formation of the earliest cleavage. Some splays are affected by cleavage whereas others in the same thin section overgrow the first cleavage. It is not possible to find cleavages in Carboniferous rocks with the same attitudes as in pre-Carboniferous rocks immediately adjacent. The most probable reason for this is that the Carboniferous deposits are all in small fault blocks (the most deeply downdropped during CNF normal motion) and, thus, have been rotated with respect to their surroundings after cleavage formation.

TIMING OF STRUCTURAL EVENTS

Because pre-Carboniferous rocks contain three prominent cleavages and Carboniferous rocks contain only two, it is reasonable to conclude that the two latest cleavages in pre-Carboniferous rocks formed during the Alleghanian orogeny. This is further corroborated by the timing of those two cleavages with respect to metamorphisms. The second cleavage in pre-Carboniferous rocks formed during the growth of chlorite and the first cleavage in Carboniferous rocks formed during the growth of chloritoid. There can be little doubt that the growth of these two minerals was approximately synchronous. The formation of the two Alleghanian cleavages can be used to determine the timing of motion on the WMZ and the CNF. The mylonites of neither of those dislocations have the first Alleghanian cleavage, despite being immediately adjacent to rocks that do contain that fabric. Thus, at least the last motion on both those features must postdate the formation of that cleavage. The last motion on both the WMZ and CNF must, therefore, be Alleghanian. The two motion episodes probably followed closely in time, but with a change in stress resulting in a different direction of motion. The second Alleghanian cleavage is present in many locations in both the WMZ and the CNF. Thus, the sequence of Alleghanina structural events in the Clinton, Massachusetts area is as follows: Formation of steep crenulation cleavage and folds; normal motion on the WMZ, normal motion on the CNF, formation of subhorizontal crenulation cleavage.

A SHEAR ZONE ORIGIN FOR ALLEGHANIAN CLEAVAGES

The two Alleghanian cleavages occur in a discrete zone which can be mapped (Figure 2). The zone strikes approximately NNE and has as its eastern boundary the CNF. The western boundary is a zone where one finds subvertical mylonites 10 to 20 cm thick, where the two cleavages disappear over a zone less than 500 ft. (Figure 2) and where the retrogression of early high grade metamorphic minerals also disappears. Considering the close association with major faults and mylonites, a shear zone is one reasonable interpretation of this zone. The cleavages which form within the zone can be viewed as strain accommodation features. For the bedding and parallel foliation outside the zone to have been folded in association with the first Alleghanian cleavage, they must have been favorably oriented, that is lying within the field of shortening of the strain ellipsoid for shearing. This would only have been the case if the first motion on the zone was sinistral. Sinistral motion on a zone striking NNE would have been produced from a subhorizontal shortening oriented approximately N-S. A major change in stress regime must have taken place after this because the next two events were normal displacement on the WMZ and CNF. Normal motion can also be deduced for the wide Alleghanian shear zone producing the second Alleghanian cleavage. Normal displacement in a steeply dipping shear zone would result in a strain ellipsoid with a subhorizontal XY plane, which would parallel the cleavage. The shear zone

model makes sense in terms of retrogression of early formed metamorphic minerals. The shear zone would have been a zone of fluid flow, with the fluids accomplishing the retrogression. Certainly, hydrous metamorphic minerals are characteristic of the zone of Alleghanian cleavages.

For many decade we have been attempting to understand the nature of the Alleghanian orogeny in New England. In the Clinton area the orogeny produced a complex series of events resulting from changes in stress directions. It may well be that other regions have experienced similar complex deformation and retrogression. It may be that the effects of the Alleghanian orogeny in New England is confined to such narrow zones rather than affecting a wide expanse of the mountain belt.

ACKNOWLEDGEMENTS

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GOLDSTEIN

ROAD LOG

Mileage

0.0 Assembly point: commuter parking lot at intersection of Interstate 495 and Route 62. Although the focus of this trip is not on the rocks of the Nashoba terrane, excellent exposures of sillimanite gneiss of the Nashoba Formation can be seen at the entrance ramp to Interstate 495N. Interestingly note the numerous vertical dip-slip ductile shear zones in these roadcuts. Exit from parking lot and proceed west on Route 62.

1.8 Town of Berlin, continue west on Route 62.

3.4 Blinking traffic light at railroad crossing. Continue west on Route 62.

4.9 "T" intersection with Route 70. Turn right onto Route 70N and Route 62W.

5.5 Turn right into parking lot for district Courthouse. Walk across street and down to shore of Wachusett Reservoir.

STOP 1. MYLONITIZED AYER GRANITE ON SMALL ISLANDS IN CARVILLE BASIN

This is not a public access area of the Wachusett Reservoir. You must obtain permission from the MDC police station located approximately 1 mile north of the Courthouse on Route 62. The large island in the small embayment known as the Carville Basin is composed entirely of Ayer granite. The south side of the island is typical undeformed Ayer granite with unoriented cm size feldspar megacrysts. The north side of the island, however, exposes intensely mylonitized Ayer granite with a foliation striking NE and dipping shallowly to the northwest. The lineation in these rocks, composed of ductilely elongated quartz and brittlely deformed feldspar megacrysts plunges down-dip. Kinematic indicators here include composite planar fabrics (S and C or S and C') and "bookshelf" fractured feldspars and indicate thrust motion. Nearly all other samples from the Wachusett mylonite zone, however, contain normal sense kinematic indicators. The unusual sense of displacement here is interpreted as being related, in some way, to the presence of the large undeformed block of Ayer granite. This can be viewed as related to either back rotating of this block or to the block existing as a remnant of undeformed rock trapped between R and R' Reidle shears in an overall normal sense shear zone. Exposures on the southern side of Carville Basin are variably mylonitized amphibolites of the Reubens Hill Formation, a unit grouped with the Nashoba terrane. Thus, the Nashoba-Merrimack terrane contact is present within the Carville Basin. Depending on water level, the small islands in the Carville Basin expose intensely mylonitized Ayer granite, Tower Hill quartzite, Worcester phyllite and Oakdale quartzite, all units of the Merrimack terrane. Therefore, the terrane contact is most likely very close to the southern side of the basin.

Return to vehicles and proceed west on Route 62.

5.7 Dam for Wachusett Reservoir on right. Eastern branch of Clinton Newbery Fault (CNF) passes beneath the dam and runs through the steep sided valley below the dam. Intensely fractured Ayer granite is exposed in the spillway for the reservoir and in the roadcut west of the dam.

6.4 Turn right onto Cameron Street.

6.7 Bear left around triangle and turn left along Chace Street.

7.8 Turn right and cross railroad tracks.

7.9 Turn right after railroad tracks onto Wataquadock Hill Road.

9.2 Pull off onto right side of road.

STOP 2. CLINTON NEWBURY FAULT AT WATAQUADOCK HILL

An excellent exposure of the Clinton-Newbury fault and its attendant mylonitization and phyllonitization can be seen at this locality. Follow the trail on the east side of the road southward

for approximately 80 feet. Turn left onto a trail which cuts to the east. After approximately 200 feet, locate large outcrops in woods to the right. Exposures here are intensely mylonitized plagioclase rich gneisses of the Reubens Hill Formation. Although these rocks are grouped with the Nashoba terrane, they are believed by the author to be in the hanging wall of the CNF. To the northeast of this locality, and continuing along the trace of the CNF, the Tadmuck Brook Schist (TBS), a rusty-weathering aluminous formation, is found immediately below the CNF. At this locality intensely mylonitized TBS is found immediately below the Reubens Hill Formation. Because the CNF splits into two fault strands, from around this point southward the eastern strand lies entirely within the Nashoba terrane (Figure 1). At this locality, one can examine the nature of the CNF faultzone. Initially, contrast the orientations of foliation and lineation here with those at Stop 1. Note the very steep foliation, dipping from 50° to 70° to the northwest. Also note that the mineral elongation lineations trend due east, distinctly different from those in the WMZ. Towards the southeast, one finds panels of dark grey phyllite intercalated with plagioclase gneisses and, finally, exclusively grey phyllite. The phyllite is interpreted to be strongly sheared and retrogressed TBS, which it grades into the southeast. The phyllite is, thus, a phyllonite, and the zone 10m wide in which one finds intercalated phyllonite and gneiss is interpreted to be the trace of the CNF. Kinematic indicators at this locality and many others show that the CNF is a normal fault. Most striking, but difficult to see unless one cuts an oriented sample, are extensional crenulation cleavages in the TBS phyllonite. Across from the parking area one finds a grey granite typical of many found in the Nashoba terrane. A small shear zone related to the CNF cuts this granite and one can see the steep foliation and east-trending lineation characteristic of the CNF. One also sees a strong, horizontal crenulation lineation associated with the subhorizontal crenulation cleavage, S_3. S_3, thus, postdates at least some motion along the CNF.

Continue north on Wataquadock Hill Road.

9.8	Turn right onto West Berlin Road for safe turnaround. Return in opposite direction in Wataquadock Hill Road.
11.8	Turn left across railroad tracks.
11.85	turn left onto Chace Street.
12.9	Bear right around triangle onto Cameron Street.
13.2	Turn right onto Routes 62 and 70. Follow Routes 62 and 70 through to center of Clinton.
13.8	Turn right in center of Clinton onto High Street.
14.1	Stop light. Go straight onto Route 110N.
17.6	Intersection with Route 117. Continue on Route 110.
19.8	Turn right onto West Bare Hill Road.
20.2	Turn right onto Scott Road.
20.8	Pull far off onto right shoulder.

STOP 3. HARVARD CONGLOMERATE AT BARE HILL

Exposures close to the road are typical Harvard conglomerate; phyllite with interlayered metaconglomerate. Here, one can clearly see two cleavages, one parallel to bedding and one at high angles to it. Several small folds with subhorizontal axes appear related to the second cleavage.

Much better exposures can be found in the woods. Walk S 30° W across the open field (first ask permission of the people in the house to the south). Follow trail southeast through the woods for approximately 1200 feet, or until the large outcrops on the right end. Walk westward, uphill, through the woods to the bare hilltop. Here you can clearly see bedding, defined by layers of metaconglomerates and phyllite. A prominent foliation is parallel to bedding and a second, strong

cleavage cuts across it. The arguments for the age of these rocks can be found in Thompson and Robinson (1976). Their carboniferous age is accepted here because of the presence of carbonaceous fragments in the rocks. Thus, both cleavages are interpreted to be Alleghanian. One finds identical relationships at the exposures of Harvard conglomerate at Pin Hill, in Harvard, several miles to the north and at the Worcester coal mine to the south.

Return to trail and proceed along trail to the southeast. Turn left at the second trail intersection. Follow the trail to the southwest to the bare top of Vaughn Hill.

STOP 3A. VAUGHN'S HILL FORMATION AT VAUGHN'S HILL

This lovely location is now the site of a spectacular home under construction. Special permission has been given by the architect to visit this now private locale. Those following this field guide on their own are urged to invest some time in obtaining permission from the owners who will move in when the house is finished, shortly after this NEIGC.

The Vaughn's Hill formation is a thinly interbedded, fine-grained quartzite and schist of unknown age. It is not aereally extensive nor does it contain fossils. It is considered to belong stratigraphically with the rocks of the Merrimack trough. A fault is interpreted to be present in the valley separating Vaughn's Hill from Bare Hill. Three prominent planar fabrics can be seen in these rocks. One is parallel to bedding; one is vertical, strikes N45°E and is axial planar to the numerous small folds; a third is subhorizontal. Considering the proximity of Vaughn's Hill and Bare Hill, the latest two cleavages in the Vaughn's Hill formation are interpreted as time correlatives of the two cleavages in the Harvard conglomerate and, thus, Alleghanian in age. The initial foliation can only be constrained to be pre-Alleghanian, but is most likely to be Acadian because it is believed to be present in the ~400 Ma Fitchburg granite.

Return to vehicles. Continue along Scott Road.

22.3 Stop sign. Turn right onto Bare Hill Road.
23.5 Turn right onto Route 117W.
24.1 Turn left onto Route 110. Bolton Orchards - Lunch supplies.
27.7 Stop light, right turn; stay on Route 110.
27.8 Stop light, turn left; stay on Route 110 (Main Street).
28.1 Stop light, continue straight on Route 110.
32.4 Pull over onto shoulder after farm buildings on right. Parking here is restricted, be sure that you are not parked in a no parking zone. Walk back and cross street. Walk along public access road north of house on east side of road. This may appear to be a driveway, but it is not. However, the land around this entrance is private and the rather unpleasant people who live in the house may attempt to restrict your access. Be advised that they cannot.

Walk along dirt road approximately 500 feet to the first dirt road on the right. Turn right and walk approximately 4000 feet nearly to the shore of Wachusett Reservoir. Follow the road as it curves to the left. Walk approximately 300 feet to the first dirt road on the right. Turn right and walk approximately 800 feet to outcrops on the shores of the reservoir.

STOP 4. SHOLAN POINT

The superb exposures at Sholan Point allow one to examine the evidence of multiple deformation in the Worcester phyllite of the Merrimack trough. The exposures here are extensive and there is much to see. However, there are three principal observations:

1) In the phyllite with thin, fine-grained quartzite interlayers there are many small folds, especially easy to see as folded quartz veins. Extreme care needs to be taken when interpreting these small buckles. One prominent and systematic fabric element which can be found in the phyllites is a subhorizontal crenulation cleavage (S_3), similar to that seen at Stop 3A.

2) North along the shores one can find quartzite beds within the phyllite. These beds are 30-40 cm thick and are light grey in contrast to the darker grey phyllite. The quartzite beds display a very strong cleavage (S_2) at high angles to bedding, suggesting close proximity to a D_2 fold hinge. Cleavage changes orientation abruptly form quartzite into phyllite sweeping in an opposite sense on the top and bottom of the bed suggesting close proximity to an anticline to the northeast or a syncline to the southwest.

3) At the southernmost end of the point, the lithology is different, consisting mostly of quartzite beds with minor phyllite. This lithology folds more competently than the phyllite and displays numerous D_2 folds. Note that there is a bedding parallel foliation (S_1) which is folded along with bedding, and that the axial plane foliation (S_2) is parallel to the prominent cleavage in the quartzite beds seen earlier. Fold asymmetry here is clockwise, indicating the same structural position as the sheared cleavage in quartzite and phyllite.

Return to vehicles. Continue west on Route 110.

32.6	Turn right into parking lot beneath power lines for safe turnaround. Exit parking area and turn left onto Route 110.
34.0	Turn left onto Chace Hill Road.
34.7	Intersection, cross road and continue on Swett Hill Road. Swett Hill Road merges with Kendall Hill Road.
36.1	Sterling Orchards. park and ask permission to leave your vehicle. Walk back to the power lines. Ask permission of people in house below power lines to walk across their property. Walk south along power lines.

STOP 5. Numerous outcrops of andalusite schist are present below the power lines. Bedding dips moderately to the southeast. Comparing these orientations with those at Sholan Point indicates that a 1st order syncline lies between the two locations. Andalusite at this locality has been almost entirely altered to white mica. In addition, one can find pseudomorphs of staurolite, now coffin-shaped patches of schist.

Return to vehicles and continue in same direction on Kendall Hill Road.

36.8	Bear left around "island".
37.3	Town of Sterling. Turn right on Route 12.
37.4	Intersection with Route 62, stay on Route 12.
39.2	Turn left on Interstate 190.
40.4	Park at south end of first roadcut on Interstate 190.

STOP 6. Pavement exposures on top of the island between the northbound and southbound lanes show two prominent foliations. One (S_1) is parallel to bedding and the other (S_2) is subvertical and parallel to the axial plane of small folds. Here, on the western side of the Wekepere Fault, S_2 has a more northeasterly strike than on the eastern side of the fault. Several small shear zones with sinistral displacement can be found and appear to be related to the S_2 cleavage. The shear zones are parallel to cleavage, are approximately 10cm wide and are marked by intense development of S_2 and by either truncation of bedding against the zone or by a discordance of bedding orientations on either side of the zone. Locally, S_3, the subhorizontal crenulation cleavage, can be seen.

40.9 Roadcut very similar to Stop 5.
42.2 Roadcut. Pull off onto wide paved shoulder.

STOP 7. This exposure is similar in many ways to the previous one. The lithology is the same and one can find bedding (S_0) with a parallel foliation (S_1) and a later crenulation cleavage (S_2) associated with small folds. Locally, S_3 can also be seen. One also finds very highly foliated zones with strong mineral elongation lineations unlike anything seen elsewhere, except in the WMZ. These zones are currently interpreted as mylonites. Locally, S_2 cuts the mylonites and, in other locations, is absent. It is concluded, therefore, that S_2 was roughly synchronous with the mylonitization. The mylonitic foliation is typically oriented N30°-35°E 45°-50°NW with lineations plunging 30°-40° to the NNW. S_2 is oblique to this orientation, typical striking N55°E with near vertical dips. As of this writing, samples of mylonite are being prepared for thin-sectioning. The results will be available for the field trip.

Return to vehicles. Continue on Route I190.

42.7 Roadcut. Pull over onto shoulder.

STOP 8. The rock fabric changes markedly from Stop 7 to Stop 8. Only locally does one see a weak S_2 cleavage and only at the northeastern edge of the roadcut. Associated with it are very low amplitude folds. No where is the subhorizontal crenulation cleavage seen only locally does one see smooth seen. Only locally does one see smooth foliation surfaces with strong mineral elongation lineations. For the most part, bedding and parallel foliation (S_1) at this locality are planar, striking N10°E to N10°W and dipping 30°-40° to the west. Thus, a prominent D_2 strain gradient is present between Stops 7 and 8, a distance of approximatley 500 feet. The significance of this is discussed in the accompanying article.

44.1 Exit from Interstate 190.
44.3 Turn right onto Route 140N.
44.5 Turn left onto Legg Road.
45.2 Park on shoulder next to large roadcuts.

STOP 9. The rocks exposed here are the typical green to purple calcareous quartzite of the western Merrimack trough, identical to those we have seen at the last two stops. However, from the northeast end of the roadcut these grade into a rusty-weathering aluminous schist bearing garnet, staurolite and andalusite. In contrast to the alteration experienced by these minerals only a mile and a half to the east, here the high grade metamorphic minerals are unaltered and pristine. Only one foliation, parallel to bedding, is present striking N10°-20°W and dipping 20°-30°SW. A prominent mineral elongation lineation, marked primarily by biotite, is present trending S25°W. Other features of interest include a sill of Fitchburg granite which has been boudinaged, indicating that at least the final aspects of deformation postdated the intrusion of that body.

Return to vehicles and turn around. Proceed back to intersection with Route 140 and turn right. I190 west will lead to Interstate 290 in Worcester which is the best route to the Massachusetts turnpike. Interstate 190 east can be taken to Route 12 south which intersects with Route 62. To return to the original meeting stop, take Route 62 through Clinton to the intersection with Interstate 495.

THE PELHAM DOME, CENTRAL MASSACHUSETTS: STRATIGRAPHY, GEOCHRONOLOGY, STRUCTURE AND METAMORPHISM

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PURPOSE OF TRIP

The Pelham dome (Figure 1) contains the only known exposure of Proterozoic (613 Ma) basement rocks, of possible Avalon affinity, in the Bronson Hill anticlinorium north of southern Connecticut. These are sharply bounded upward by a 200m-thick layer of Late Ordovician (455-443 Ma) calc-alkaline intrusive gneisses typical of the cores of the other domes. These are, in turn, cryptically overlain by Late Ordovician, Silurian, and Lower Devonian cover strata of the Bronson Hill sequence. The structure is not that of a massif, but of a series of east-directed recumbent fold and thrust nappes on which the dome structure is superimposed (Figure 2). The relations in the Pelham dome have commonly been cited as evidence that the Bronson Hill magmatic arc was founded on continental crust of Avalon affinity (Hall and Robinson, 1982) or that little metamorphosed Avalon continental rocks were thrust beneath older Acadian high-grade rocks during the late Paleozoic (Gromet, 1989; Getty and Gromet, 1992). The purpose of this trip is to give participants an insight into the field relationships that must provide the basis for tectonic reconstructions, and to call attention to the surprising results of recent isotopic dating by several methods that demonstrate how truly complex the geology is. Emphasis will be on seeing and trying to understand what is known, rather than discussing comprehensive stratigraphic-tectonic-metamorphic models.

THE "CLASSIC VIEW" (PRE-1988) OF PELHAM DOME STRATIGRAPHY AND STRUCTURE

Northern Part of the Dome

The classic view of internal Pelham dome stratigraphy and structure was established by Ashenden (1973) in the northern part, during construction of the Northfield Mountain Pumped Storage Project . Ashenden has summarized older research. The work began with surface mapping of the pumped storage reservoir site and surroundings, and examination of cores from diamond drill holes, some more than 900 feet deep, into the site of the underground powerhouse (Figure 3). This was followed by underground mapping of the access tunnel, underground powerhouse and associated draft tubes and penstocks by Ashenden and Robinson in 1968 and 1969, and finally mapping of the tailrace tunnel by Robinson in 1969 and 1970. From this developed the concept of a "mirror-image" stratigraphy with a slab of Dry Hill Gneiss in the center, overlain and underlain by Poplar Mountain Quartzite, in turn overlain and underlain by Poplar Mountain Gneiss. Thus, the Dry Hill Gneiss was conceived as occupying the core of a fold nappe with stratigraphically younger basal Poplar Mountain Quartzite and Poplar Mountain Gneiss in inverted position beneath and in right-side-up position above (Figure 4). Because of the common interbeds of quartzites and calc-silicate rocks in the Dry Hill Gneiss, it was envisaged initially as a series of rhyolitic volcanics, and this is supported by the geochemical work of Hodgkins (1985). The stratigraphically overlying Poplar Mountain Quartzite and Gneiss were considered as slightly younger sedimentary units, probably with a source in the volcanics. Layers of actinolite quartzite and of hornblende-leucogneiss in the Poplar Mountain clearly established its close affinity with the Dry Hill. The recumbent fold hypothesis was further supported by details of the Dry Hill Gneiss itself, in which a Hornblende Member in the core is underlain and usually overlain by a distinctive massif Biotite Member. An early U-Pb zircon analysis from the Dry Hill hornblende member (Naylor et al., 1973) established the unit as late Proterozoic (ca. 600 Ma), much older than any other unit then or subsequently dated in the Bronson Hill anticlinorium.

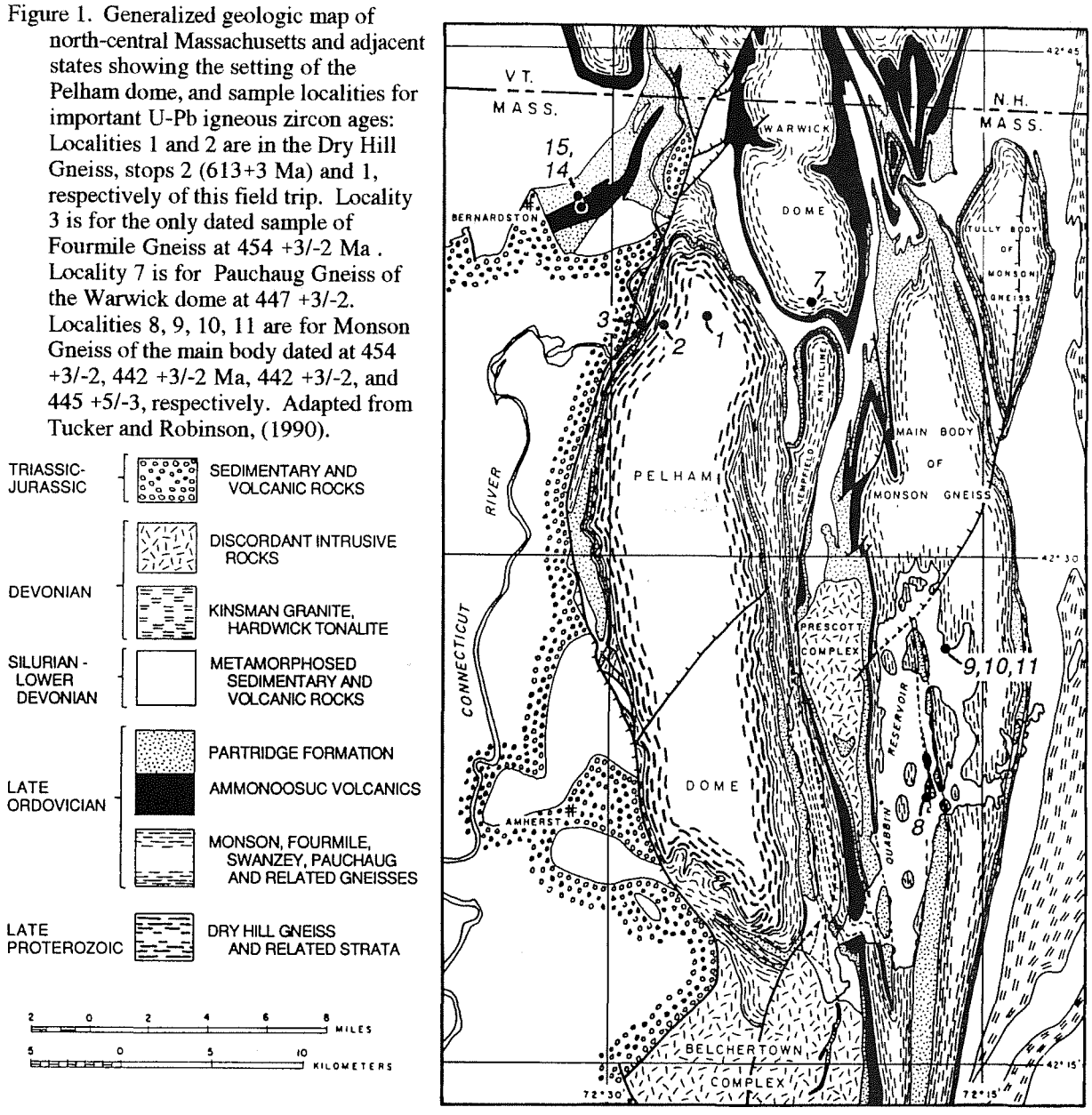


Figure 1. Generalized geologic map of north-central Massachusetts and adjacent states showing the setting of the Pelham dome, and sample localities for important U-Pb igneous zircon ages: Localities 1 and 2 are in the Dry Hill Gneiss, stops 2 (613+3 Ma) and 1, respectively of this field trip. Locality 3 is for the only dated sample of Fourmile Gneiss at 454 +3/-2 Ma . Locality 7 is for Pauchaug Gneiss of the Warwick dome at 447 +3/-2. Localities 8, 9, 10, 11 are for Monson Gneiss of the main body dated at 454 +3/-2, 442 +3/-2 Ma, 442 +3/-2, and 445 +5/-3, respectively. Adapted from Tucker and Robinson, (1990).

Although no hinges for the proposed fold nappe could be located, evidence developed by Robinson (1963) for early recumbent folds in basement-cover contacts farther north in the dome, and evidence from the Pelham-Shutesbury syncline (Michener, 1983), showed that the early recumbent folds are east-directed and so, an eastward closure of the nappe of Dry Hill Gneiss was suggested in cross section (Figure 4). A long recognized problem is the great asymmetry in thickness between the thick Poplar Mountain Quartzite and Gneiss below the Dry Hill and the thin Poplar Mountain Quartzite and Gneiss above, and the fact that no younger unit has been seen in the inverted sequence. Early on, Ashenden recognized that the predominant gneisses above the Poplar Mountain are gray plagioclase gneisses and amphibolites quite unlike the Dry Hill and much more like the plagioclase gneisses exposed in other domes. He named these Fourmile Gneiss for the excellent exposures on Fourmile Brook in Northfield Farms. Initially these were considered as probably metamorphosed volcanics, but more recent considerations by Robinson and Hollocher (Robinson et al., 1989), based on large wave-washed exposures near Quabbin Reservoir, suggest these are more likely a highly deformed complex of intrusive igneous rocks with only minor volcanics. Robinson and Ashenden discussed at length the significance of the base of the Fourmile, which Robinson had been able to study in detail where perfectly exposed in the tailrace tunnel, but only for one period of about two hours. All of of this came to little, when it was discovered that the basal Fourmile rock is in fact a sill of Devonian Belchertown Quartz Monzodiorite! Subsequently we have a U-Pb zircon age of 454+3/-2 Ma on true Fourmile

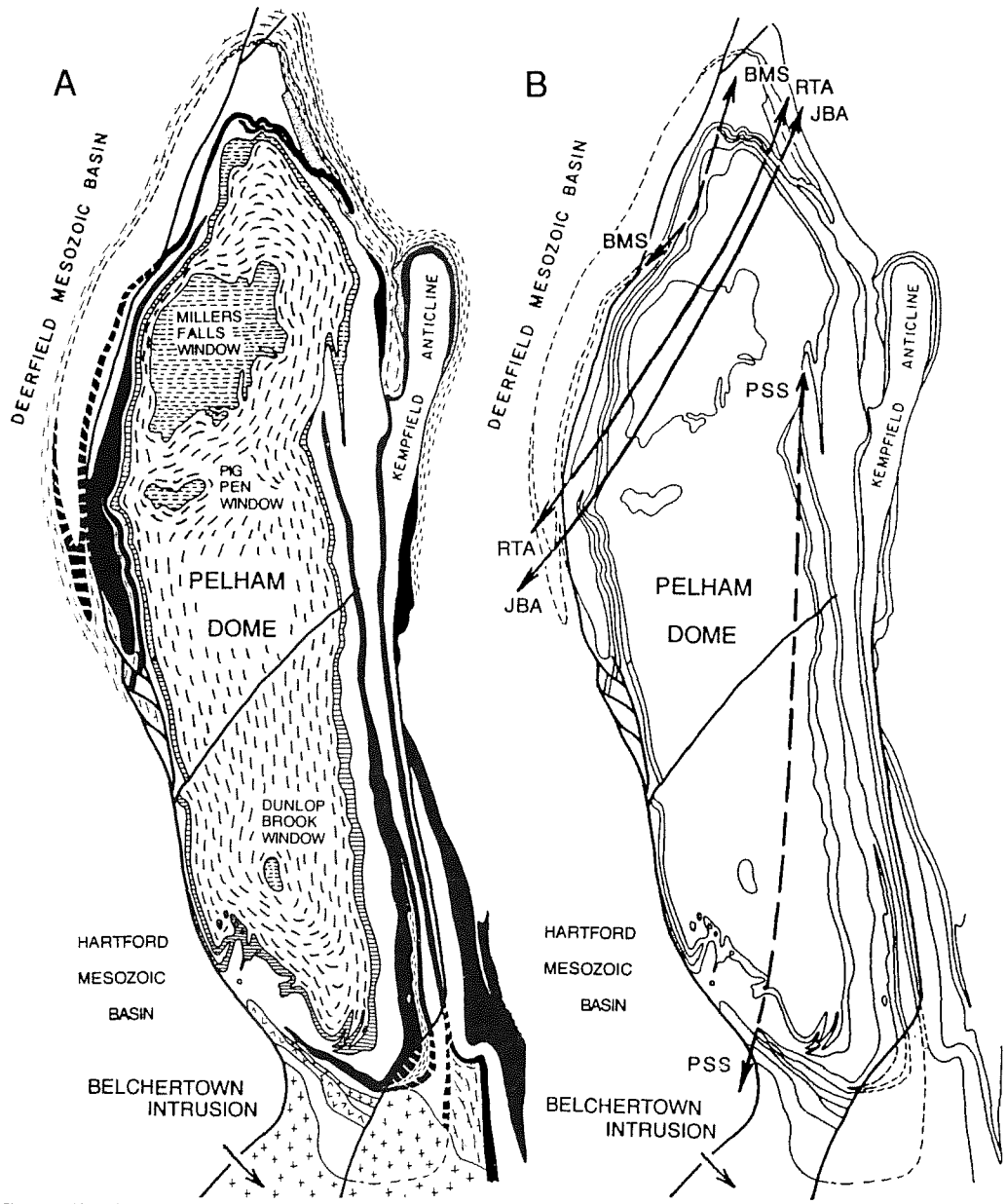


Figure 2. Generalized geologic maps of the Pelham dome. A) Map showing major stratigraphic units and structural features with geology restored west of Connecticut Valley border fault. B) Similar map showing eight hinges of early east-directed recumbent folds, all in the contact between Partridge Formation and Fourmile Gneiss, connected by axial lines above the present erosion surface,. PSS) Pelham-Shutesbury syncline (Michener, 1983); BMS) Beers Mountain syncline; RTA) Roman T anticline; JBA) Jacks Brook anticline. Other folds are represented only by axial surfaces and lack exposed hinges.

Gneiss from the tailrace excavation, which is very similar to ages of plagioclase gneisses in other domes (Tucker and Robinson, 1990). The nature of the contact is still an enigma, but one is strongly pressed to believe it is a major fault. If true, this may also be the explanation of the wildly different thicknesses of the two Poplar Mountain sequences.

The stratigraphic concepts developed by Ashenden were carried around the south and west sides of the Millers River Window (Figure 3) by Onasch (1973) and Laird (1974). Onasch also mapped separately several layers of quartzite-rich gneiss and Dry Hill - like gneiss within the Poplar Mountain Gneiss of the window. Laird mapped another window, the Pig Pen Window, southwest of Dry Hill, exposing inverted Poplar Mountain Quartzite from beneath surrounding Dry Hill Gneiss, and traced the right-side up sequence along the west flank of the dome south of Millers Falls.

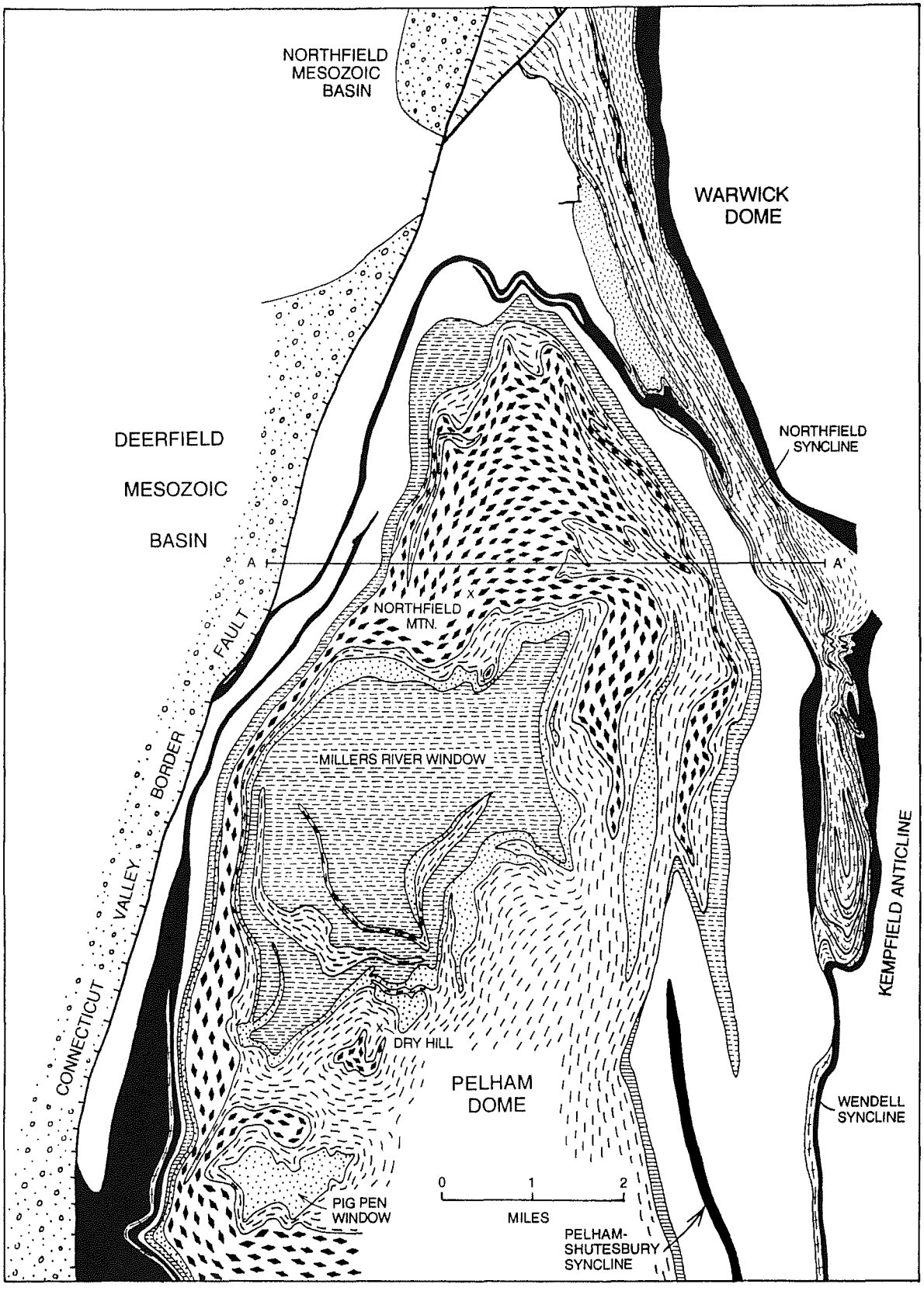


Figure 3. Generalized geologic map of the northern part of the Pelham dome after Ashenden (1973) showing detailed distribution of major rock units and line of cross section in Figure 4. For key to patterns see Figure 4.

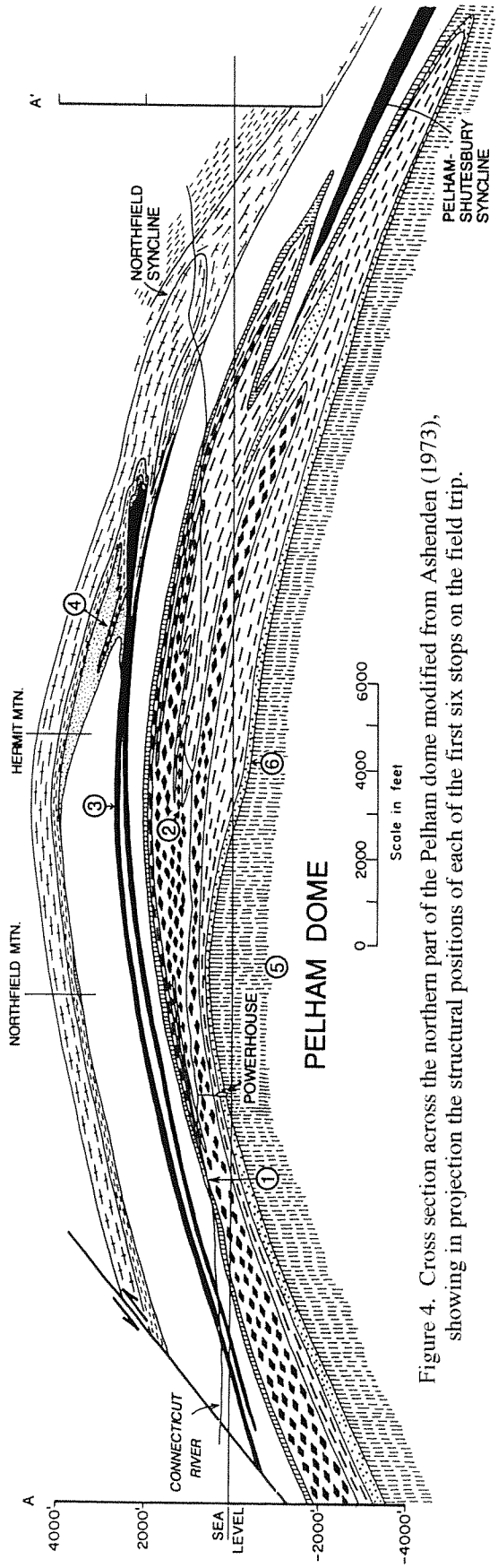
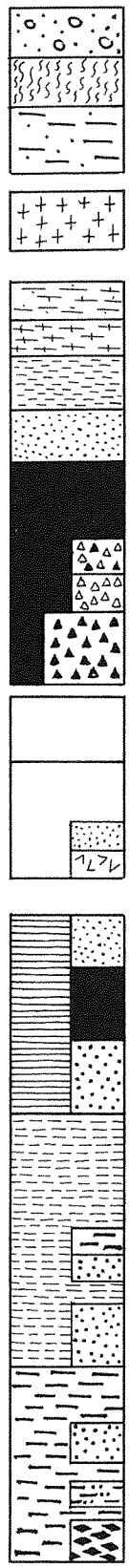


Figure 4. Cross section across the northern part of the Pelham dome modified from Ashenden (1973), showing in projection the structural positions of each of the first six stops on the field trip.

Key to patterns for units in Figures 2, 3, 4, 5, 11, and 14.



TRIASSIC-JURASSIC SEDMENTARY ROCKS

MESOZOIC SILICIFIED ZONE

SILURIAN RANGELEY FORMATION AND GRANITE WEST OF BORDER FAULT

DEVONIAN BELCHERTOWN QUARTZ MONZODIORITE

DEVONIAN ERVING FORMATION
GRANULITE MEMBER
AMPHIBOLITE MEMBER

DEVONIAN LITTLETON FORMATION

SILURIAN CLOUGH QUARTZITE

ORDOVICIAN PARTRIDGE FORMATION
MICA SCHIST AND AMPHIBOLITE

AMPHIBOLITE AND FELDSPAR GNEISS
FELDSPAR GNEISS

ORDOVICIAN AMMONOOSUC VOLCANICS

ORDOVICIAN PAUCHAUG GNEISS OF WARWICK DOME

ORDOVICIAN FOURMILE GNEISS

QUARTZITE
COARSE INTRUSIVE FACIES

MOUNT MINERAL FORMATION
SILURIAN ? UPPER QUARTZITE MEMBER

ORDOVICIAN ? MICA SCHIST AND AMPHIBOLITE MEMBER

PROTEROZOIC? BASAL QUARTZITE MEMBER

PROTEROZOIC POPLAR MOUNTAIN GNEISS

GNEISS MEMBER
JERUSALEM HILL GNEISS
JERUSALEM HILL QUARTZOSE GNEISS

QUARTZITE MEMBER

PROTEROZOIC DRY HILL GNEISS
BIOTITE MEMBER

PELHAM QUARTZITE MEMBER

SHUTESBURY QUARTZITE-SCHIST MEMB.

HORNBLENDE MEMBER

The most obvious structural features to be seen in the northern part of the dome are a strong foliation, normally quasi-parallel to compositional layering and a strong N-S trending mineral lineation defined by biotite aggregates, quartz lenses and ribbons, actinolite or hornblende needles, and elongated pebbles in conglomerates. In the cover-sequence strata the lineations are statistically parallel to a wide variety of minor fold axes (Robinson, 1963). Deeper in the dome, there is a variety of asymmetric minor folds that are not usually parallel to the lineation, but lie at any angle up to 90° to it. In a few exposures, the fold axes can be seen curving over as much as 140° (Ashenden, 1973, Figure 20). When these minor folds are plotted with their rotation senses on equal area diagrams, as in Ashenden, 1973, Figure 25 or Onasch, Figure 23 and Figure 30, they form two distinct groups on opposite sides of a separation angle that is 20° or less. This makes it obvious that these are sheath folds (Onasch, 1973, Figure 5) formed during an episode of north-over-south shearing. The direction of the mineral lineation lies directly within the separation angle (Onasch, Figure 25), showing that the lineation is parallel to the transport direction of the folds, and this is confirmed by observations of great circle rotation of the lineation by some of the folds. Coupled with these observations of asymmetric minor folds, there are rare observations of shear and fault surfaces, as well as a wide variety of asymmetric kinematic indicators, such as asymmetric tails on microcline megacrysts, as extensively studied by Reed (Reed and Williams, 1989; Reed, 1990, 1992) as discussed below. The sheath folds have not been successfully observed in cover-sequence strata. This could be because they are not there, but is more probably due to more severe shear in the cover, in which essentially all fold axes have been rotated into parallelism with the transport direction. An important recent discovery that supports this is the finding by Reed and Williams (1989) that the same strain field does indeed pervade the cover sequence northeast of the dome, as shown by rotated inclusion trails in growing garnet and staurolite porphyroblasts. This pattern of minor structural features in the northern part of the dome indicates there was a major phase of longitudinal transport parallel to the axis of the dome, with higher rocks sliding southward relative to deeper rocks. This pattern of shear appears to bear no relation to whether the rocks are right-side-up or upside-down due to the earlier recumbent folding.

Southern Part of the Dome

Mapping in the southern part of the dome was carried out by Advanced Mapping Classes at the University of Massachusetts between 1972 and 1978 (Figure 5). On the east flank, in the Shutesbury Quadrangle, (Robinson et al., 1973), a new stratigraphic sequence was worked out beginning with Dry Hill Gneiss at the base (see Stop 8), followed by Emerson's (1898) Pelham Quartzite, a relatively pure quartzite with minor feldspar and biotite, and commonly with minor actinolite. This is followed by more Dry Hill Gneiss (Rocky Run Gneiss of Robinson et al., 1973) and then by a new unit, the Mount Mineral Formation. This is characterized by a variety of mica schists, amphibolites, quartzites, and rare gneisses and ultramafic rocks. In some areas this can be differentiated into at least three members, a basal quartzite with actinolite like the Pelham Quartzite (mc), a middle member of kyanite-mica schist and amphibolite (msa), and an upper member of muscovite-garnet quartzite with subordinate gray mica-garnet schist (mq). The Mount Mineral Formation, commonly the upper quartzite member, is overlain by the Fourmile Gneiss (see Stop 8), which here is subdivided into a lower yellow-weathering member, commonly with abundant muscovite and minor garnet, and an upper gray member of biotite gneiss and amphibolite, more like the type Fourmile. Two curious features of the yellow member of the Fourmile are scattered lenses of garnet amphibolite and gedrite gneiss reminiscent of the Ammonoosuc Volcanics, and a few layers of muscovite quartzite closely resembling the upper quartzite member (mq) of the Mount Mineral Formation. Although some of the rock types in the Mount Mineral Formation seemed to resemble some of the cover units, particularly Partridge Formation and Clough Quartzite, their stratigraphic setting and their state of metamorphism (see below) strongly indicated that they belong to a Proterozoic sequence.

It was at this early stage in 1972 that Lewis Ashwal discovered coarse sillimanite schist within the middle member (Robinson, Tracy and Ashwal, 1975) with relics of an older granulite-facies metamorphism, studied in detail by Roll, 1987 (see also Robinson, 1991). At this time the common occurrence of coarse K-feldspar in this unit and also in the Dry Hill and Poplar Mountain Gneisses, and the absence of similar features in the overlying Fourmile Gneiss, strongly suggested that the old high grade metamorphism, best preserved in the pelitic schists, was probably Proterozoic.

The sequence developed in the eastern part of the Shutesbury Quadrangle was mapped around the southern end of the dome to the southwestern flank, where the metamorphosed harzburgite lenses including the "Pelham asbestos quarry" were found to lie within the schists of the Mount Mineral Formation. The map pattern defines a series of south- to southwest-directed recumbent folds. Some of these have exposed hinges, and some appear in pairs so that fold axes can be traced across the south-plunging axis of the dome itself. These are folds in foliation generally at an angle to lineation and appear to be giant versions of the late asymmetric folds found in the northern part of the dome.

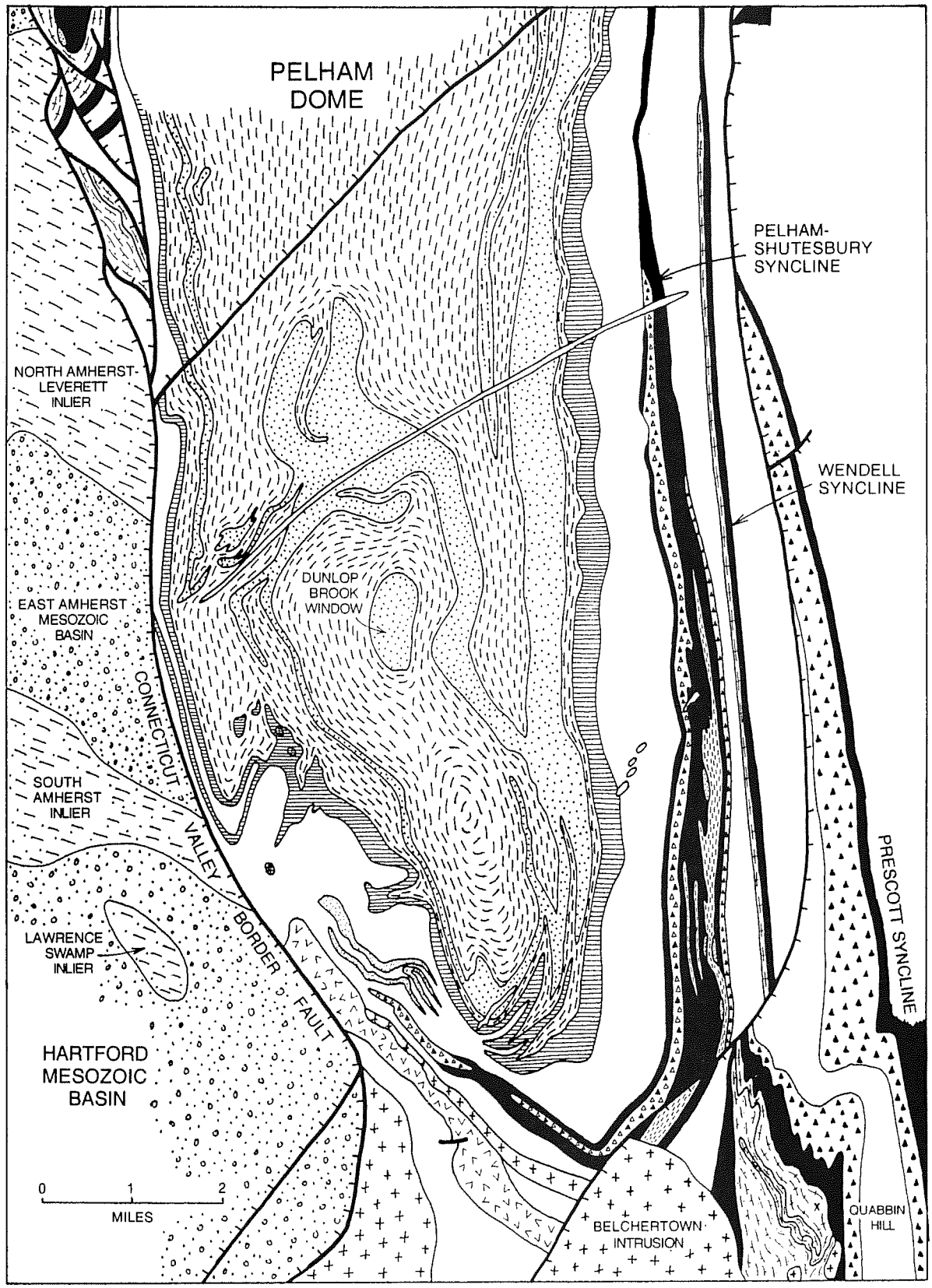


Figure 5. Generalized geologic map of the southern part of the Pelham dome based on Robinson et al., (1973), unpublished field data from Advanced Mapping classes 1972-1980, and Zen et al., (1983). For key to patterns see Figure 4.

Minor folds of the same type are also abundant. The Pelham Quartzite thins drastically, and becomes involved in a series of these giant, late-asymmetric folds and other folds that make the outcrop pattern baffling, especially near Mount Orient (see Figure 14). The deepest exposed structural level in the southern part of the dome is in the erosional window at Dunlop Brook, where quartzite is exposed that was tentatively correlated with the inverted Poplar Mountain Quartzite as exposed in the Pig Pen window. It is curious that although the Pelham dome is 44 km long, linear features plunge south along the crest only in the southernmost 4 km. This is explained by the large cumulative structural asymmetry produced by the late asymmetric folds as explained by Ashenden (1973, Figure 34).

Detailed mapping and correlation across the northwestern part of the Shutesbury quadrangle still remains to be accomplished. Detailed correlation between the northern and southern parts of the dome has been hampered by this gap and by the generally poor outcrop in the southeastern part of the Millers Falls quadrangle (Figure 3), but the following concepts were developed for the Massachusetts Bedrock Map (Zen et al., 1983). Pelham Quartzite is a large sandstone lens within the upward facing part of the Dry Hill Gneiss. The exposures at Dunlop Brook could be a downward-facing equivalent of the Pelham Quartzite or of the Poplar Mountain Quartzite as suggested on the State Map. The Mount Mineral Formation was considered to be a southward, more aluminous facies of the upper Poplar Mountain Quartzite and Gneiss. Strongly favoring this was the characteristic actinolite-bearing quartzite at the base of both units, the development of large K-feldspars in both, and the position of both below the base of the Fourmile Gneiss.

Ordovician-Silurian-Devonian Cover

The stratigraphy of cover units on the northeastern flank of the Pelham dome was established by Robinson (1963). The plagioclase gneisses, now called Fourmile Gneiss (Ashenden, 1973) are overlain by sulfidic mica schists and amphibolites of the Ordovician Partridge Formation. These, in turn, are overlain by the Lower Silurian Clough Quartzite, predominantly quartz-pebble conglomerate. The Clough is above a regional unconformity and locally cuts out the Partridge to rest directly on Fourmile Gneiss (see Stop 4). Above the Clough is the Lower Devonian Littleton Formation consisting of massive to well bedded coarse-grained garnet-staurolite-mica schists (see Stop 4). The Littleton is overlain with apparent unconformity by the type Erving Formation (not to be seen on this field trip), consisting of fine-grained well bedded quartz-plagioclase-biotite granulite, calc-silicate granulite, and rare kyanite schist with thin to very thick mappable layers of epidote amphibolite. Locally, the Erving Formation is in direct contact with the Fourmile Gneiss (Figure 3). The relations between the Erving and Littleton Formations are intimately related to a major stratigraphic dilemma and the question of a Whately thrust (Robinson et al., 1988; Trzcienski et al., this guidebook). The sequence on the northeastern flank is traced southward through the Wendell syncline and its extremely narrow southward extension to the northern part of the Great Hill syncline at Quabbin Hill (Figure 5) where Erving Formation directly overlies Partridge Formation. On the west flank of the dome, where not cut out by the Connecticut Valley border fault, the Fourmile is overlain by a substantial amount of Partridge Formation, in turn locally overlain by Clough. The Littleton is absent and the Clough or Partridge are directly overlain by Erving Formation (see map of Leverett area in Robinson et al., 1988).

Robinson (1967) rediscovered the cover stratigraphic sequence of Emerson's (1898) Pelham-Shutesbury syncline east of the southern part of the dome, and showed that these rocks occur in an early eastward and downward-facing isoclinal syncline completely surrounded by plagioclase gneisses of the dome. The entire Pelham-Shutesbury syncline was mapped in detail by Michener (1983), but will not be seen on this field trip. The east limb of the Pelham-Shutesbury syncline is the only place on the Pelham dome where the Ammonoosuc Volcanics occur between the Fourmile Gneiss and the Partridge Formation. Although the Ammonoosuc is thin, the characteristic two-fold stratigraphic sequence with lower amphibolites and gedrite-plagiolcase gneisses and upper garnet-muscovite gneisses is well defined (Schumacher, 1988). The Partridge Formation of sulfidic schists and amphibolites and rare hornblendites has a thin basal quartzite and, on the west limb, extensive separately mapped volcanics . The Clough is very thin with a characteristic pure basal quartzite and overlying mica schist and calc-silicate. The youngest unit, the Littleton, is gray kyanite-garnet-mica schist unlike the Littleton northeast of the dome, leading to much speculation as to how the strata of the Pelham-Shutesbury syncline connect at depth with the other parts of the cover sequence. No evidence of an older metamorphism, other than the "ambient" kyanite-staurolite grade metamorphism has ever been found in the Pelham-Shutesbury syncline, despite extensive study of large wave-washed exposures and electron-probe analyses of several specimens (Tracy et al., 1976; and R. J. Tracy, unpublished data).

Contact Relations with the Belchertown Intrusion

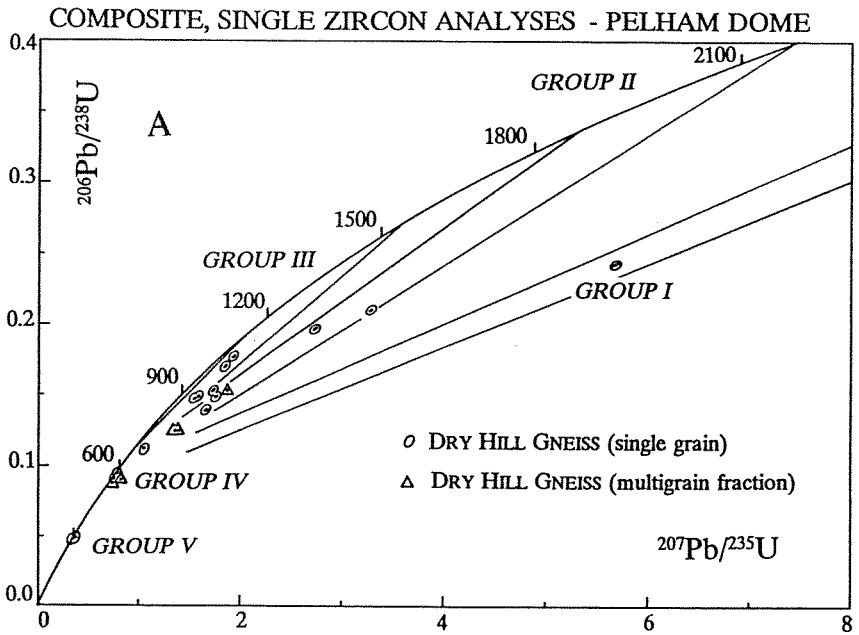
The Belchertown Quartz-Monzodiorite intrusion consists of a primary core of orthopyroxene-augite monzodiorite and a variably hydrated and metamorphosed rim of hornblende monzodiorite and hornblende monzodiorite gneiss. The core has yielded a concordant U-Pb age on zircon of 380 ±5 Ma (Ashwal et al., 1979) and pelitic inclusions within the intrusion show clear evidence of intense contact metamorphism. A major sill of the hornblende monzodiorite gneiss intrudes the Fourmile Gneiss in the southern part of the Pelham dome (Figure 5) and smaller sills are known farther north on the west flank of the dome, including the one to be seen at Stop 1A. The contact relations of the Belchertown intrusion with the southeastern corner of the Pelham dome are extremely critical for interpreting the history of the dome, but they are also very poorly exposed. The intrusion clearly cuts into the strata now forming the Great Hill syncline south of Quabbin Hill (Figure 5), completely truncating the Erving Formation, and a septum of schist from the west limb of the syncline now remains as a three-mile long contact-metamorphosed xenolith within the intrusion. The map pattern, which is the best interpretation of available outcrop information, strongly suggests, that the Pelham-Shutesbury syncline is truncated by the Belchertown intrusion. If this interpretation is correct, then the early recumbent folding represented by the Pelham-Shutesbury syncline and related folds is older than 380 Ma, and the Proterozoic rocks, also recumbently folded, must have been in contact with their cover before 380 Ma. The relations with the Belchertown intrusion are thus critical, because the present interpretation would preclude the possibility that the Proterozoic rocks were unaffected by Acadian metamorphism and deformation, and that they could have reached their present position only in the late Paleozoic.

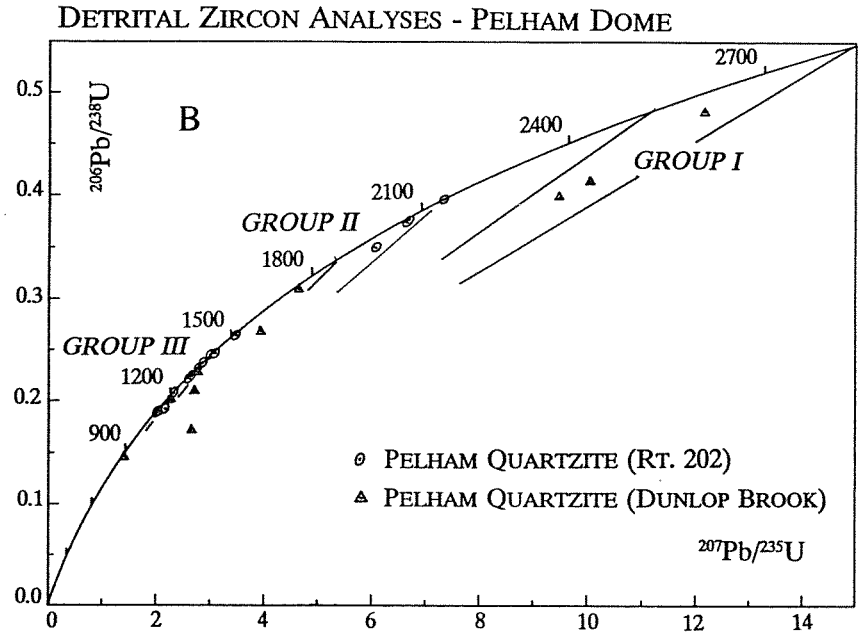
GEOCHRONOLOGY BEARING ON THE AGE AND PROVENANCE OF THE ROCKS

The geochronology of a variety of rocks in the Bronson Hill anticlinorium of Massachusetts and southwestern New Hampshire was reported by Tucker and Robinson (1990) and many of the localities and ages are summarized in Figure 1. Subsequent work has focussed on several topics, including the implications of the inheritance reported in the zircon populations of a sample of Dry Hill Gneiss at Stop 1, which is better dated at 613 ±3 Ma on a sample from Stop 2 that lacks this inheritance.. These features are illustrated in the concordia diagram in Figure 6A. Multigrain analyses of zircon suggest a *minimum age* of inheritance of 1407 Ma, the $^{207}Pb/^{206}Pb$ age of the most discordant analysis, which can only be considered as the *mean* age of the fraction analyzed. Optical examination of many grains, especially those polished and etched with HF vapor, reveled an internal, composite structure indicating that zircon cores within individual grains were the source of the inheritance. That is, the old radiogenic Pb was distributed within many grains of the fractions and not one or two xenocrysts in an otherwise uniform population. Analysis of 15 single composite grains demonstrates the presence of variably aged inheritance with age groupings of 2850-2550 Ma, 2150-1850 Ma, and 1510-1000 Ma (Tucker and Robinson, 1991). There are two other groupings at ca. 600 Ma and 300 Ma, the known times of igneous emplacement and regional metamorphism, respectively (Tucker and Robinson, 1990).

On the basis of the single-grain data in the Dry Hill Gneiss, a program of single detrital grain analyses was undertaken on some of the associated quartzites and one sample of Poplar Mountain Gneiss. Figure 6B shows 18 analyses of detrital single zircons from the Pelham Quartzite Member of the Dry Hill Gneiss on Route 202 (see Figure 14 and outcrop description following Stop 8,) and 12 analyses from the quartzite exposed on Dunlop Brook (see Figure 14) that had been tentatively correlated with the inverted Poplar Mountain Quartzite. Many of the analyses are nearly concordant and they replicate the age groupings defined by the inherited zircon components in composite igneous grains of the Dry Hill Gneiss as shown in Figure 6A. No detritus younger than 933 Ma was measured; thus the quartzites appear to lack detrital zircon from the enclosing Dry Hill Gneiss and other possible source rocks of late Proterozoic age. This data throws some doubt on the hypothesis that the Dry Hill Gneiss is volcanic rather than intrusive. One grain has a $^{207}Pb/^{206}Pb$ age of 2680 Ma, which is the minimum age of the grain. Single detrital zircon analyses from the inverted Poplar Mountain Quartzite at Stop 6 are shown in Figure 6C. The oldest grains overlap in age with the youngest population of the Pelham Quartzite, but none are as young as the 613 Ma Dry Hill igneous event. Taken together, these results suggest a Grenville and western Superior Province provenance (Corfu and Davis, in press) including also felsic plutons of early Proterozoic age (2150-1800 Ma). Recently, David and Machado (1992) have studied a similar zircon suite in the lower sedimentary sequence in the Miramichi terrane in New Brunswick, and have proposed a source in the Amazon shield.

Figure 6C also shows single detrital zircon analyses from the muscovite-garnet bearing upper quartzite member of the Mount Mineral Formation. Most of the analyses are concordant despite the fact that the rock has undergone up to sillimanite-orthoclase-grade metamorphism (see Stops 8,9). The source rock ages are distinctly different than for the quartzites in Figure 6B. The youngest grains in the Mt. Mineral Formation Quartzite are 459-439 Ma,





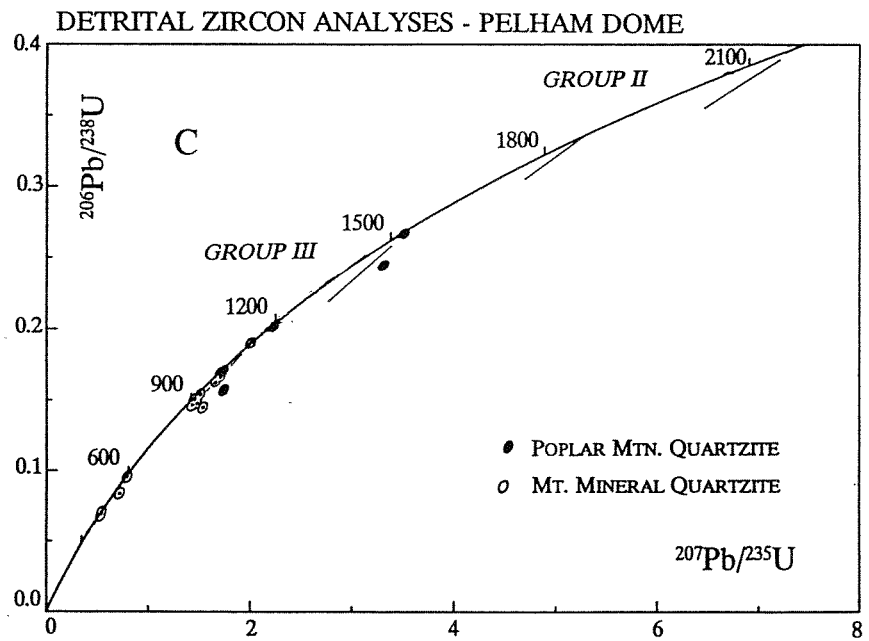


Figure 6. Concordia diagrams of U-Pb zircon analyses from the Pelham dome.

A) Multi- and single-grain zircon analyses from the Late Proterozoic (613 Ma) Dry Hill Gneiss at Stops 1 and 2 (see text).

B) Thirty analyses of detrital single zircons from the Pelham Quartzite Member of the Dry Hill Gneiss and the quartzite inlier on Dunlop Brook (see text).

C) Single detrital zircon analyses from the inverted Poplar Mountain Quartzite at Stop 6 and from the upper quartzite member of the Mount Mineral Formation at Stop 8. Greater than 85% of the analyses are concordant despite up to sillimanite-orthoclase-grade metamorphism, at least in the Mount Mineral sample.

demonstrating that the quartzite is younger than earliest Silurian and may correspond to the Clough Quartzite. The older grains in this sample include ~1000 Ma, ~900 Ma, and ~~600 Ma groups that could have derived from North American Grenville basement (McLelland et al. 1988; Ralcliffe et al., 1991), post-Grenville intrusives into it (Karabinos and Aleinikoff, 1990) or the Dry Hill Gneiss. The discovery that a presumed Proterozoic quartzite, beneath 200 m of Late Ordovician plutonic rock, is actually Silurian, has required drastic rethinking. However, two decades ago, the similarities of the Mount Mineral rocks to more familiar cover-sequence units were considered and then rejected because of the apparent physical improbability of such a correlation!

AGES OF METAMORPHISM

On published regional metamorphic maps of central Massachusetts (Robinson, 1991, Figures 2 and 25) the Pelham dome lies entirely within Zone I, the kyanite-staurolite-muscovite zone. This is succeeded by Zone II, the sillimanite-staurolite-muscovite zone, on the east limb of the Warwick dome and around the Keene dome and bodies of Monson Gneiss. Three or four occurrences of kyanite overgrown by fibrolite seemed to confirm that this is a prograde sequence (Robinson, 1963). In 1972 Ashwal discovered coarse relict sillimanite and orthoclase in mylonitic schist within the Mount Mineral Formation deep within Zone I (Robinson, Tracy, and Ashwal, 1975), and, because of the lithic association of these rocks, Robinson (1983, Zen et al., 1983) suggested that this older metamorphism was probably late Proterozoic. The idea of a Proterozoic granulite-facies metamorphism overprinted by an Acadian kyanite-staurolite-muscovite grade metamorphism was supported in a detailed study by Roll (1987). These ideas were investigated by radiometric dating of monazite and zircon from two of these outcrops. The concordia diagram in Figure 7A shows monazite and zircon from sillimanite-orthoclase pegmatite M22A, and monazite from adjacent schist. Zircon analyses from the pegmatite are concordant, suggesting crystallization at 350±2 Ma. Concordant analyses of monazite from the sheared but little reconstituted host sillimanite-orthoclase-garnet-biotite-rutile schist suggest high-grade metamorphism at 367±3 Ma, about 17 m.y. before pegmatite emplacement. These results were the first indication that the relict high-grade metamorphism in these rocks was Acadian rather than Late Proterozoic as believed previously (Robinson, 1983; Zen et al., 1983; Roll, 1987). The results further imply that the overprinting kyanite-muscovite grade metamorphism, thought to be the "ambient" Acadian grade, might actually be younger.

In the early stages of Tucker's program of isotopic dating, a sample of Dry Hill gneiss (Stop 1) from the core of the Pelham dome yielded sphene with concordant U-Pb ages of 292 ±5 Ma (Tucker and Robinson, 1990). Spear and Harrison (1989) obtained an Ar-Ar plateau age on hastingsite from the same outcrop of 287 ±1 Ma. In contrast, several samples from Zone II yielded monazite and sphene recrystallization ages around 360 Ma and these and other Late Devonian to Mississippian zircon, monazite and titanite (sphene) ages from a variety of rocks in a broad region of sillimanite-muscovite-staurolite zone and higher-grade rocks east of the Warwick dome and Kempfield anticline are shown in Figure 7B. The range of Acadian igneous emplacement and metamorphic ages spans 25 m. y. from 375 to 350 Ma. These ages are comparable to monazite ages in the granulite-facies center of the central Massachusetts metamorphic high determined by Barreiro (Thomson et al., this volume). They confirm that all of this high-grade metamorphism was Acadian as long-supposed, but they also show that it occurred substantially later than the peak metamorphism recorded in the highest grade rocks in central New Hampshire (Eusden and Barreiro, 1988; Smith and Barreiro, 1990; Zeitler et al., 1990.) The cross-cutting sillimanite pegmatite at "Iolite Hill" at the south end of the Keene gneiss dome (Figure 7B) is Late Mississippian.

In view of the apparently continous field of gneissic strain fabrics from the Pelham dome eastward, it was suggested early on that the young sphenes and hastingsites may have been subjected to a late Paleozoic heating event. However, Gromet (Gromet and Robinson, 1990) produced Rb-Sr and Sm-Nd mineral isochrons (see below) indicating the rocks underwent thorough recrystallization during gneiss formation at around 290 Ma. Particularly impressive was a Sm-Nd isochron that included euhedral metamorphic garnet. Subsequent studies by Tucker yielded a large collection of tightly constrained Late Carboniferous (Late Pennsylvanian) U-Pb metamorphic ages on sphene, monazite, and zircon as well as similar igneous ages on zircon and monazite from a variety of pegmatites ranging from severely to mildly deformed within the Pelham dome, the Northfield and Wendell synclines, and the Kempfield anticline.(Figure 7C). These include the ages of igneous zircons and monazites in pegmatites with mild to severe tectonic overprinting, that cut Poplar Mountain Quartzite, Fourmile Gneiss, and Clough Quartzite. The range of Pennsylvanian metamorphic and igneous emplacement ages spans 11 m. y. from 298 Ma to 287 Ma.

Gromet has investigated the question of late Paleozoic metamorphism and recrystallization rather extensively using the Rb-Sr and Nd-Sm as well as the U-Pb systems. His work gave the first good indication that the late Paleozoic ages are not merely cooling ages, but are the result of late Paleozoic recrystallization accompanying ductile

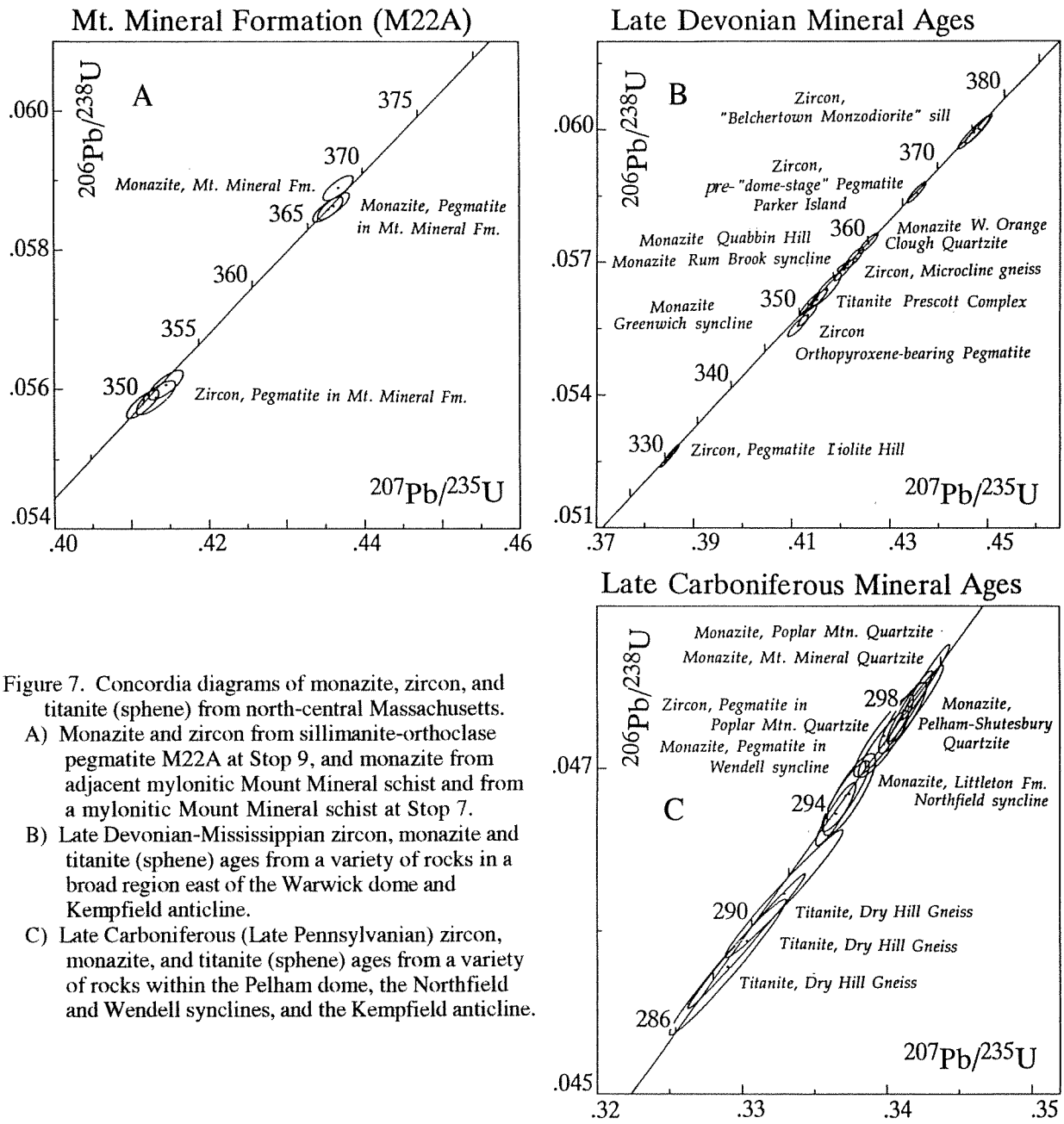
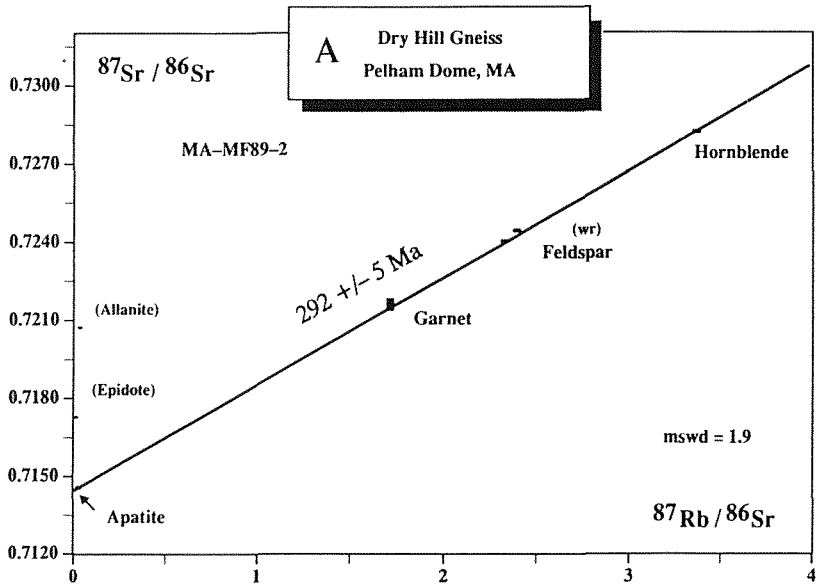
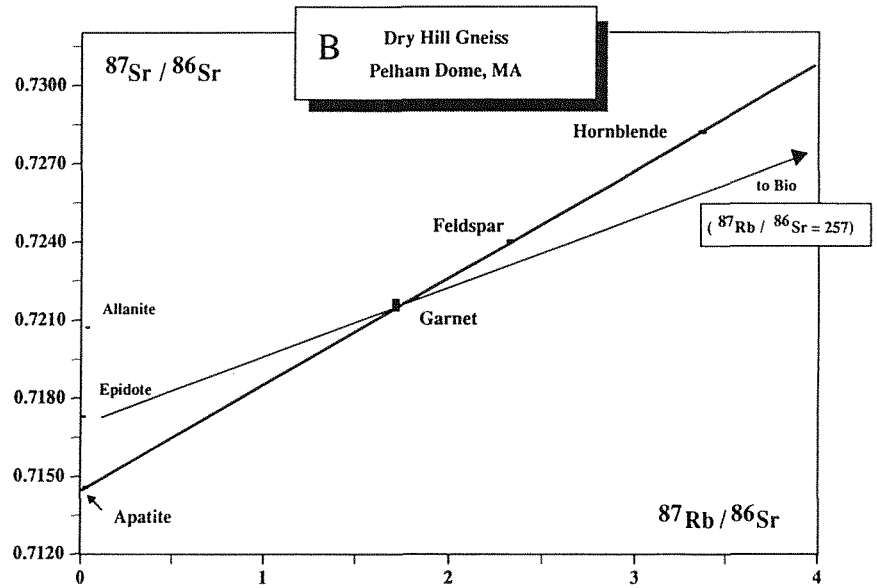
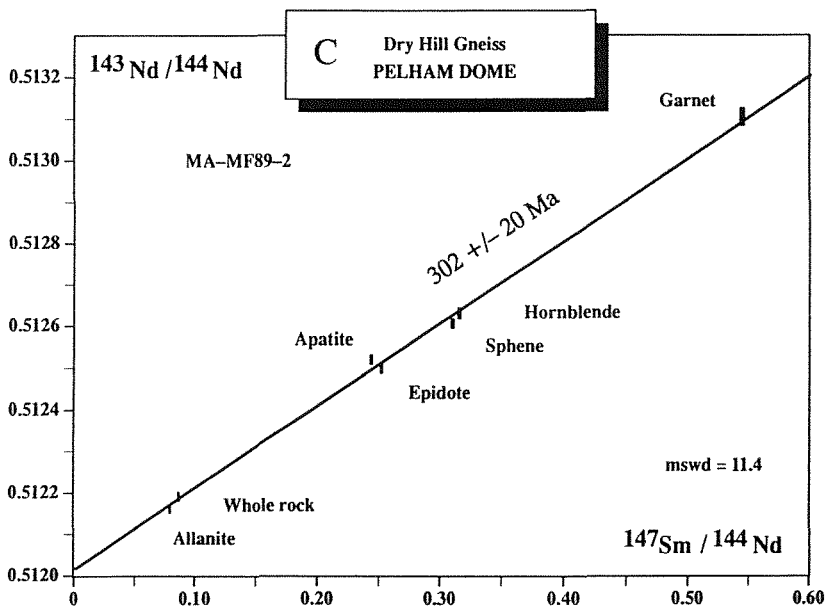


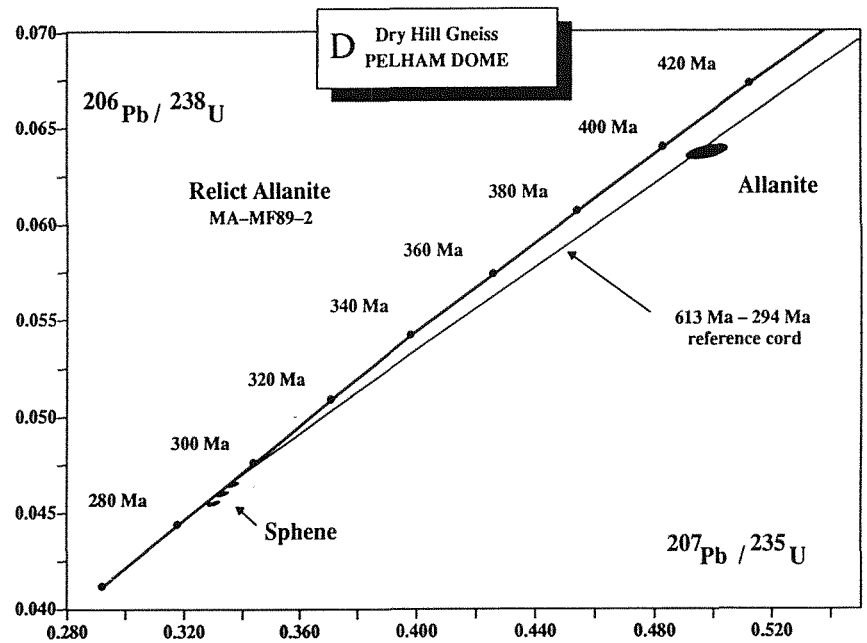
Figure 7. Concordia diagrams of monazite, zircon, and titanite (sphene) from north-central Massachusetts.
A) Monazite and zircon from sillimanite-orthoclase pegmatite M22A at Stop 9, and monazite from adjacent mylonitic Mount Mineral schist and from a mylonitic Mount Mineral schist at Stop 7.
B) Late Devonian-Mississippian zircon, monazite and titanite (sphene) ages from a variety of rocks in a broad region east of the Warwick dome and Kempfield anticline.
C) Late Carboniferous (Late Pennsylvanian) zircon, monazite, and titanite (sphene) ages from a variety of rocks within the Pelham dome, the Northfield and Wendell synclines, and the Kempfield anticline.

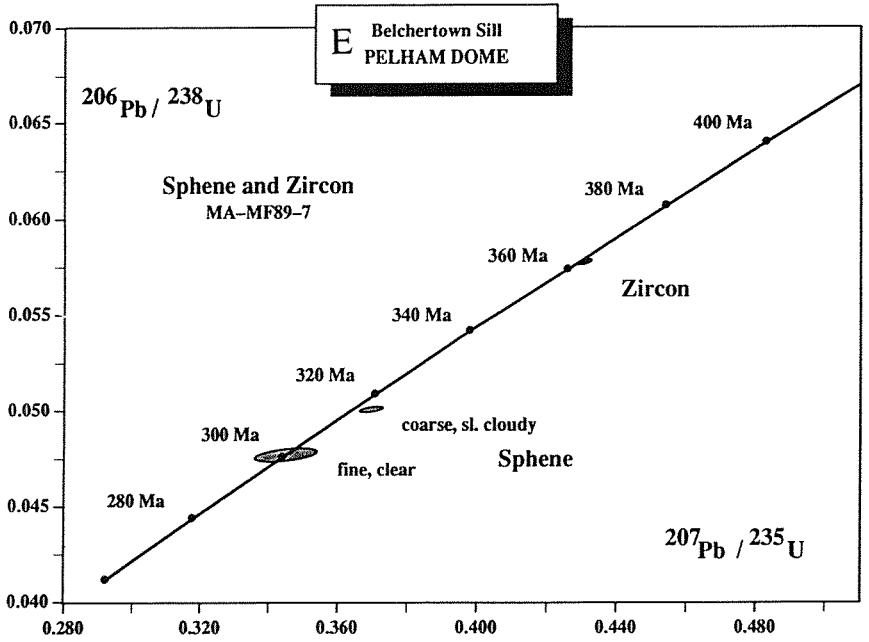
deformation (Gromet, 1989; Gromet and Robinson, 1990). Figure 8A shows a Rb-Sr isochron diagram of minerals and whole-rock from Dry Hill Gneiss at Stop 2, indicating isotopic homogenization between most minerals occurred at 292±5 Ma. In Figure 8B, allanite and epidote plot at elevated values of $^{87}Sr/^{86}Sr$ because they contain old, radiogenic ^{87}Sr, possibly gained from biotite after 292 Ma. Figure 8C shows a Sm/Nd reference line corresponding to an age of 302±20 Ma, through garnet, hornblende, sphene, epidote, apatite, whole rock and allanite from the same rock. Particularly impressive are the data on garnet which would not be expected to be reset easily by diffusion during cooling under low amphibolite conditions. Probe analyses by Hodgkins (1985) strongly suggest these garnets contain prograde growth zoning. In the concordia diagram in Figure 8D sphene analyses overlap or are near concordia at ca. 295 Ma, but the allanite plots about 1/3 up the reference chord. and is interpreted as a primary igneous grain that has lost about 2/3 of its radiogenic Pb during dynamic recrystallization/ metamorphism at about 294 Ma. Figure 8E is a concordia diagram of zircon and sphene in the Belchertown sill at Stop 1A indicating a crystallization age of approximately 370 Ma and a generation of clear sphene growth at 300 Ma. Coarse cloudy sphene lies near a chord between the fine sphene and zircon ages, suggesting that some older metamorphic sphene may have formed or that igneous sphene was only partially reset during the 300 Ma event. Figures 8E and 8F show

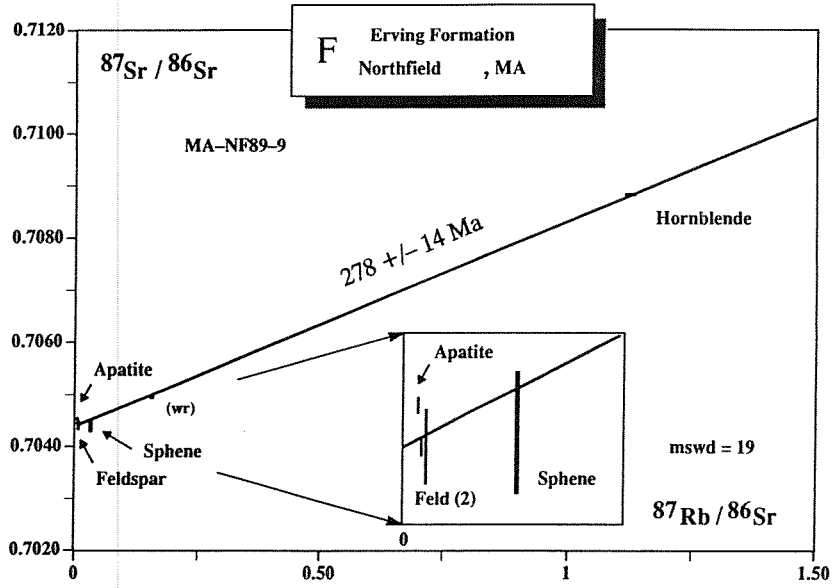

A
Dry Hill Gneiss
Pelham Dome, MA
87Sr / 86Sr
MA–MF89–2
Hornblende
(wr)
Feldspar
292 +/– 5 Ma
(Allanite)
Garnet
(Epidote)
mswd = 1.9
Apatite
87Rb / 86Sr
0.7300
0.7270
0.7240
0.7210
0.7180
0.7150
0.7120
0
1
2
3
4


B
Dry Hill Gneiss
Pelham Dome, MA
87Sr / 86Sr
Hornblende
to Bio
(87Rb / 86Sr = 257)
Feldspar
Allanite
Garnet
Epidote
Apatite
87Rb / 86Sr
0.7300
0.7270
0.7240
0.7210
0.7180
0.7150
0.7120
0
1
2
3
4


C
Dry Hill Gneiss
PELHAM DOME
143 Nd /144 Nd
Garnet
MA–MF89–2
302 +/– 20 Ma
Hornblende
Apatite
Sphene
Epidote
mswd = 11.4
Whole rock
Allanite
147Sm / 144 Nd
0.5132
0.5130
0.5128
0.5126
0.5124
0.5122
0.5120
0
0.10
0.20
0.30
0.40
0.50
0.60


D
Dry Hill Gneiss
PELHAM DOME
206Pb / 238U
420 Ma
400 Ma
Relict Allanite
MA–MF89–2
380 Ma
Allanite
360 Ma
340 Ma
613 Ma – 294 Ma
reference cord
320 Ma
300 Ma
280 Ma
Sphene
207Pb / 235U
0.070
0.065
0.060
0.055
0.050
0.045
0.040
0.280
0.320
0.360
0.400
0.440
0.480
0.520





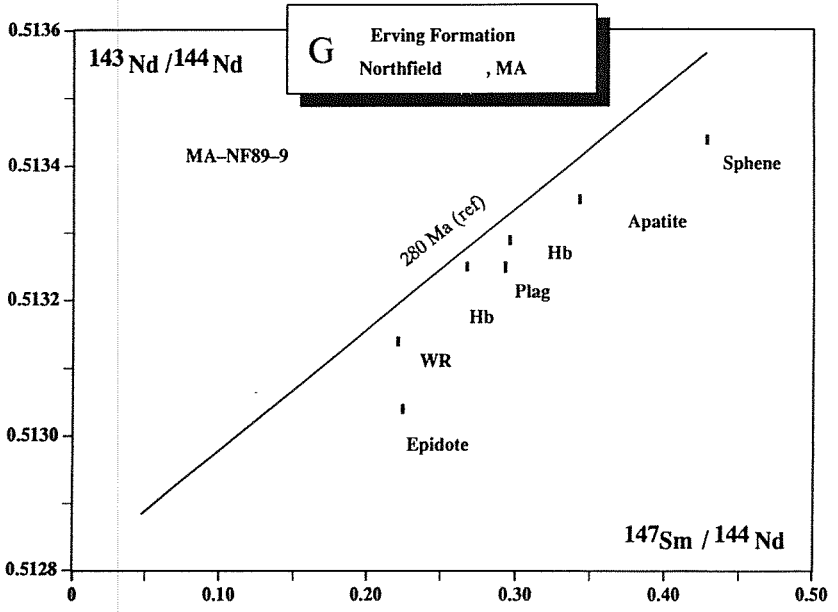


Figure 8. Isotopic results from in and near the Pelham dome.

A) Rb/Sr isochron diagram for minerals and whole-rock from the Hornblende Member of the Dry Hill Gneiss at Stop 2. Analyses represented by error boxes corresponding to the external 2-sigma-mean errors.

B) Rb/Sr isochron diagram with allanite, epidote, and biotite from same rock.

C) Sm/Nd isochron diagram of minerals and whole-rock from the same rock. Reference chord of 302 ± 20 Ma is defined by most minerals and whole-rock except apatite. Sphene and allanite are solution aliquots of minerals in D.

D) U-Pb concordia diagram showing allanite and three concordant analyses of clear faceted sphene from the same rock. Also shown is the reference chord defined by Tucker and Robinson (1990).

E) U-Pb concordia diagram of zircon and sphene from the sill of Belchertown Quartz Monzodiorite at Stop 1A..

F). Rb/Sr isochron diagram of minerals and whole rock in Erving hornblende schist in the Northfield Syncline.. For clarity, errors on the Rb/Sr axis are enlarged for the low Rb/Sr minerals. Reference line of 278 ±14 Ma is drawn through the whole rock and all minerals except apatite.

G) Sm/Nd isochron diagram of minerals and whole-rock in for the same rock. Total disequilibrium among minerals and whole-rock is indicated by scatter about a reference isochron of 280 Ma.

Rb/Sr and Sm/Nd data for a sample of Erving Formation hornblende schist totally lacking biotite from the Northfield syncline. Hornblende, plagioclase, sphene and the whole-rock suggest a resetting age of 278±14 Ma (Figure 8E) which compares favorably to a monazite U-Pb age of 295 in a nearby outcrop of Littleton Formation staurolite schist. In contrast to the Rb-Sr data, Sm-Nd analyses of solution aliquots from the same minerals (Figure 8F) indicate total disequilibrium among minerals and whole-rock.

DEFORMATION-FABRIC STUDIES

Kinematic indicators and timing relationships related to the dominant north-south trending mineral lineation are well exposed in both the core and cover rocks of the northern part of the Pelham Dome, and have been the subject of study by Reed and Williams (1989; Reed, 1990, 1992). Kinematic indicators in the Poplar Mountain and Dry Hill Gneisses include asymmetric pressure shadows ("strain shadows"), S-C fabrics, asymmetric extensional shear bands, mica fish, asymmetric folds, and others. The most distinctive and common indicators are asymmetric strain shadows associated with feldspar megacrysts. The term "strain shadow" is preferred over pressure shadow because these domains do not appear to be sites of preferred precipitation of material, but instead are domains of slightly reduced finite strain relative to that typical of the matrix. The feldspar megacrysts, some several centimeters across, are dispersed in a fine- to medium-grained biotite quartz feldspar matrix. They are typically rectangular to slightly rounded in cross section with straight, sharp margins. Locally, some megacrysts have a narrow, less than 1 mm, pink, sugary-textured, recrystallized mantle, but most show little evidence of internal ductile deformation. In the field, asymmetric, triangular strain shadows occur adjacent to virtually all feldspar megacrysts. In the subhorizontal foliation of the northern part of the Pelham Dome, the shadows invariably step upward from north to south indicating top-to-the-south shearing. In hand specimen, the shadows are defined by their lighter color, coarser grain size, and a subtle curvature of the matrix foliation. In thin section, matrix phases inside and outside of shadows are moderately to strongly annealed. Quartz and feldspar grain boundaries are generally straight and smooth but undulose extinction and subgrains are common. The shadows are subtle in thin section. They are typically defined by a slightly reduced modal proportion of biotite and a systematic curvature of the biotite-defined foliation. Adjacent to megacrysts, biotite makes an angle of approximately 30-45°with the matrix foliation. Away from the megacrysts the foliation curves progressively into parallelism with the matrix fabric.

The strain shadows are interpreted to represent low-strain domains, relative to typical matrix, that formed in the lee of the rigid feldspar megacrysts. Matrix quartz and feldspar crystals are slightly coarser and more annealed than typical matrix, presumably because the ratio of recrystallization rate to strain rate was slightly greater than that experienced by the matrix. The mode of biotite is slightly reduced in the shadows because the domains experienced a lesser degree of strain-related dissolution of quartz and feldspar. The oblique orientation of biotite crystals presumably documents a smaller amount of shear strain compared to typical matrix biotite, but it is interesting to note that many biotite crystals are nearly parallel to the theoretical infinitesimal flattening direction at 45° to the matrix fabric. The deformational behavior of the gneisses is distinctive in that the matrix phases, including feldspar, display abundant evidence for dynamic recrystallization, and the megacrysts are essentially undeformed. The deformational conditions were such that dislocations were mobile in the finer, probably wetter matrix, but were relatively immobile in the large dry feldspar megacrysts. This suggests that temperatures of deformation were on the order of 400 to 600°C. Below 400°C, matrix feldspars would be expected to behave brittlely; above 600°C, the megacrysts would be expected show significantly more deformational microstructure. It should be noted that high strain rates might be expected to increase the contrast in behavior between megacrysts and matrix, but the extreme homogeneity of the deformational microstructures throughout a thickness of more than 1 km of gneisses suggests that strain rates were probably not unusually great.

Kinematic indicators are also well developed in the cover rocks, especially the Erving, Partridge, and Littleton Formations, on the northeastern limb of the dome. They include S-C fabrics, asymmetrical extensional shear bands, mica fish, asymmetric folds, asymmetric pressure shadows ("strain shadows"), offset pegmatites, and others. All indicators consistently suggest dextral shear associated with the generally east-dipping foliation. As with the underlying gneisses, asymmetric strain shadows are probably the most important because of their abundance and the implications for the timing of deformation and metamorphism. In the Littleton Formation, quartz-rich shadows with coarser grain size and curving foliation are present on many garnet, kyanite and staurolite porphyroblasts. Again, the shadows are interpreted to represent domains of lower finite strain, and thus greater quartz/mica ratios, but some precipitation of new quartz may have occurred in these domains. Some garnet porphyroblasts contain curving inclusion trails that either represent an earlier tectonic fabric or an earlier stage of the shear-related fabric. Several staurolite porphyroblasts, with strain shadows, contain subhedral to euhedral garnet porphyroblasts which are themselves associated with shadows defined by quartz and ilmenite inclusions in the host staurolite porphyroblast.

These relationships indicate that ductile shearing occurred during and after the development of garnet, kyanite, and staurolite porphyroblasts in the Littleton Formation Thus, at least a component of the deformation occurred at middle amphibolite facies conditions, compatible with temperatures of deformation interpreted for the Poplar Mountain and Dry Hill Gneisses.

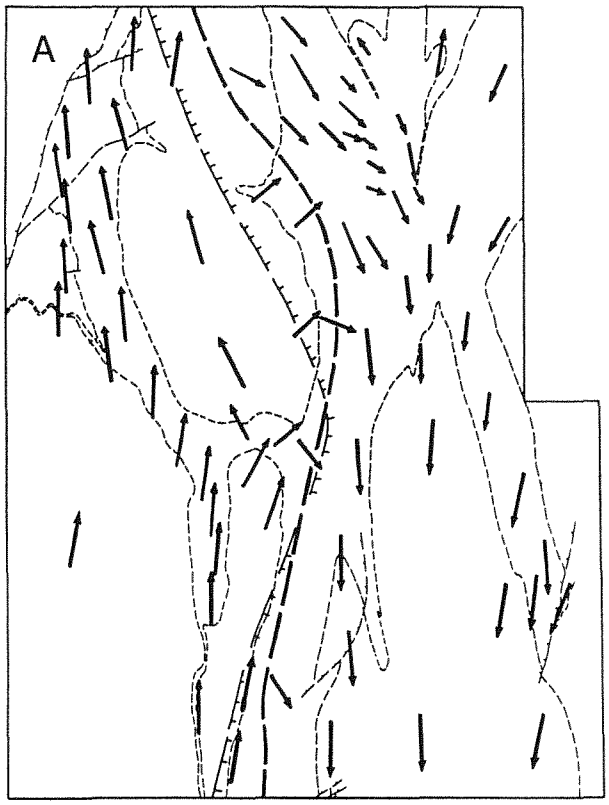
DelloRusso and Robinson (1989) made a pilot study of kinematic indicators parallel to the lineation within the central part of the Warwick dome, where the lineation forms a swirling pattern (see Figures 9 and 10). The rocks are much more homogeneous than in the Pelham dome, with very few porphyroclasts, but enough could be seen, mainly in S-C fabrics, to make estimates of shear sense in 16 out of 17 specimens. Overall these results indicate relative southward or upward movement of the gneisses in the core relative to cover. At first this might seem opposite to what has been observed in the Pelham dome. However, Reed has found a consistent east-side-south sense entirely across the Northfield syncline to the overturned southwest flank of the Warwick dome. When translated into terms of the Warwick dome, this indicates southward movement of the core relative to the cover, exactly as described above.

In response to the new geochronology, in the last several months Robinson and Peterson have undertaken new fabric studies of the reconstituted and mylonitic schists (Roll, 1987) of the Mount Mineral Formation, with the following preliminary results. In the fully reconstituted schists, the north-over-south kinematic indicators described earlier by Reed and Williams (previous paragraphs) predominate, indicating that this shearing was in progress during growth of new garnet and staurolite, apparently at the same time that garnet and staurolite were undergoing growth in the cover schists of the Northfield syncline (Reed and Williams, 1989; Reed, 1990, 1992). However, in the mylonitic schists which successfully resisted this recrystallization due to local scarcity of water, there is clearly an older linear fabric trending roughly E-W which involves oriented sillimanite, highly elongate quartz ribbons, and tails around orthoclase and garnet, including evidence that the orthoclase was itself undergoing strong grain-size reduction with limited recrystallization. One of the samples from stop 7 is a sillmanite-orthoclase-garnet-rutile mylonite with subordinate biotite in which quartz, plagioclase and orthoclase all underwent severe grain-size reduction with only minimal recovery. These features are all evidence of an earlier and much higher-temperature phase of shearing, probably soon after the peak of Acadian granulite facies metamorphism. At several localities on the east limb of the dome (Stops 7 and 9) sawed slabs and oriented thin sections indicate an east-side-down sense of shear during this apparently Acadian phase. Such a shear sense would be consistent with westward underthrusting of the Proterozoic rocks with respect to the overlying Mount Mineral Formation, but the petrologic relations strongly suggest this thrusting was Acadian. Such a shear sense would also be consistent with the pattern of proposed Acadian east-directed fold and thrust nappes that appear to dominate the gross structure of the dome.

STRATIGRAPHIC AND STRUCTURAL IMPLICATIONS FROM NEW GEOCHRONOLOGY

The most devastating implication of the new geochronology is that at least part of the Mount Mineral Formation is not Proterozoic, but Silurian or younger. This means that it must exist either as a deep infold of cover between underlying Proterozoic rocks and overlying Ordovician "basement", or more probably as a sliver of cover caught beneath a thrust sheet of Ordovician "basement". Such narrow belts of cover between areas of "basement" are common in this part of Massachusetts, but none had been previously identified in direct contact with the Proterozoic sequence.The concept that the base of the Fourmile Gneiss is a thrust is compatible with earlier work, though not required by it. If, as seems possible, the Fourmile Gneiss and higher strata did not undergo the Acadian granulite-facies metamorphism, then the postulated thrust must have come after 350 Ma, and must be one that carried formerly cooler older rocks on top of formerly hotter younger rocks. If the Proterozoic rocks beneath the Mount Mineral Formation also did not undergo the Acadian granulite-facies metamorphism, then a second thrust must be postulated that also came later and that carried former hotter younger rocks over formerly cooler older rocks.

Another devastating implication of the new geochronology is that a large region in and east of the Pelham dome underwent strong ductile deformation and kyanite-staurolite grade regional metamorphism in the late Pennsylvanian, whereas a still larger area farther east, bounded approximately by the regional sillimanite-staurolite isograd, reached peak metamorphism in the Late Devonian (Figure 9). Both of these regions are characterized by strong deformation fabrics dominated by a late N-S trending mineral lineation. The similarity of lineation in these two regions had always been a strong argument that they are of the same age, which the new data disproves. Because the structural and metamorphic features associated with the New Salem retrograde zone about in the middle of Figure 9B (Robinson, 1963, Hollocher, 1981), appears to straddle the boundary between the Acadian and Pennsylvanian structural and metamorphic regions, it appears likely that this zone is even younger than Pennsylvanian.



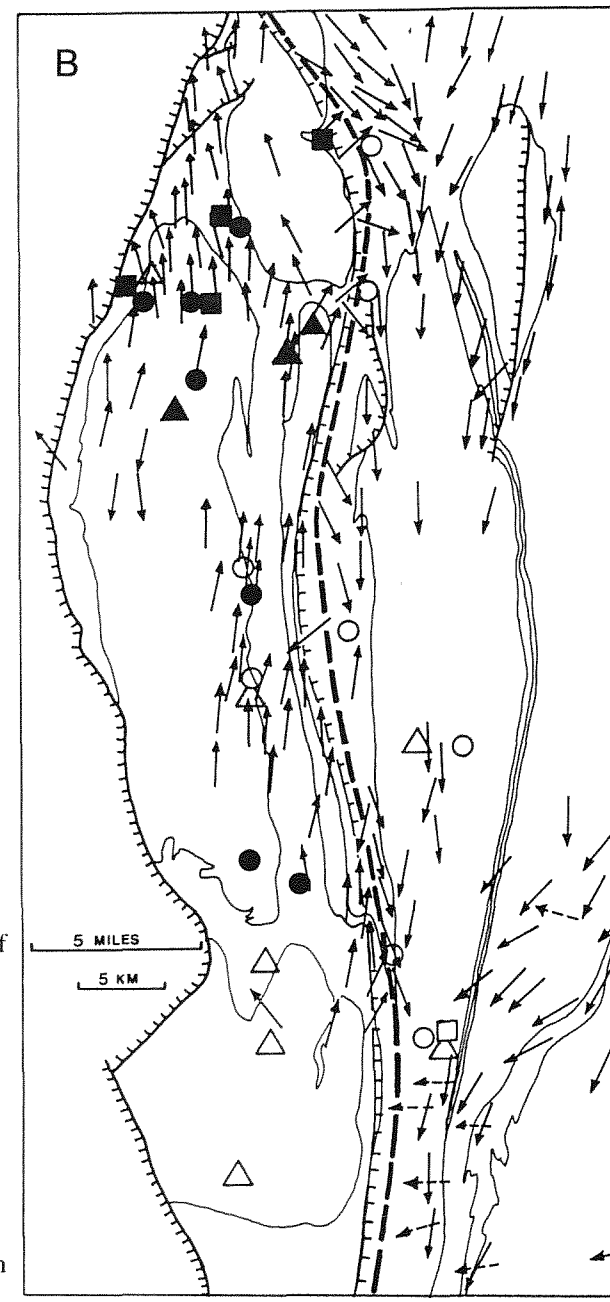


Figure 9. The "great swirl" of mineral lineations in north-central Massachusetts.

A) Distribution of mineral lineations in the Orange area (Robinson, 1963). Each arrow represents the mean of a subarea containing from tens to hundreds of lineation measurements. Length of arrows is inversely proportional to plunge as follows: long arrow - 0-30°; intermediate arrow - 30-60°; short arrow 60-85°. Heavy dashed line is the trace of the "great swirl" defined as the locus of points where lineation plunges due east. Adjacent hachured line is best estimated position of the fibrolitic sillimanite isograd.

B) Compilation of lineation data over a broader region combined with information on distribution of Acadian and Pennsylvanian metamorphic and igneous ages. Heavy dashed line is the trace of the "great swirl" defined as the locus of points where lineation plunges due east. Adjacent hachured line is the fibrolitic sillimanite isograd. Closed symbols indicate Pennsylvanian ages, open symbols Acadian ages. Circles - U-Pb ages on metamorphic minerals by Tucker; triangles - U-Pb ages on igneous minerals byTucker; squares - Rb-Sr or Sm-Nd isochrons on metamorphic minerals by Gromet. Both of the Acadian metamorphic ages west of the swirl are in relict Mount Mineral schist at Stops 7 and 9. Of the five Acadian igneous ages west of the swirl, one is in the relict sillimanite-orthoclase pegmatite at Stop 9 and the other four are from Belchertown Quartz Monzodiorite.

In the western region in the Pelham dome a north-over-south shear sense along the lineation was documented long ago (Ashenden, 1973; Onasch, 1973), and the same shear sense is now observed as right-lateral shear all the way across the isoclinal Northfield syncline (Reed, 1992). The only observations of shear sense along the lineation in the eastern region are those of Peterson (1992 and this volume) who found strong evidence for a late Acadian right-lateral shear along the eastern margin of the Monson Gneiss. However, a giant swirling pattern of the lineation recognized and puzzled over by Robinson (1963; DelloRusso and Robinson, 1989, Robinson et al., 1991)) between these regions (Figures 9, 10) can now be tentatively interpreted as due to the superimposition of a Late

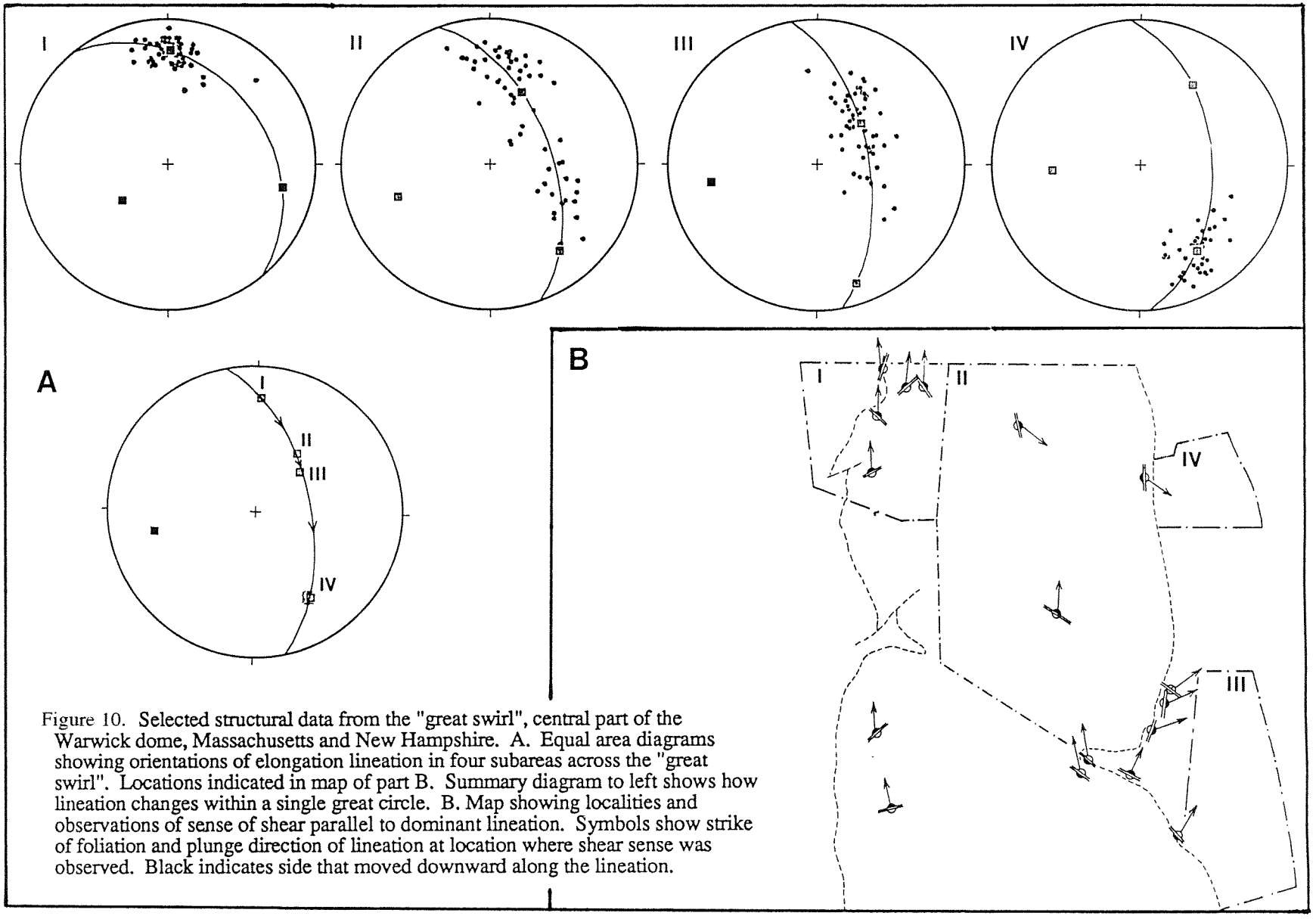


Figure 10. Selected structural data from the "great swirl", central part of the Warwick dome, Massachusetts and New Hampshire. A. Equal area diagrams showing orientations of elongation lineation in four subareas across the "great swirl". Locations indicated in map of part B. Summary diagram to left shows how lineation changes within a single great circle. B. Map showing localities and observations of sense of shear parallel to dominant lineation. Symbols show strike of foliation and plunge direction of lineation at location where shear sense was observed. Black indicates side that moved downward along the lineation.

Pennsylvanian strain field on previously strongly lineated rocks, with resultant reorientation of the lineation up to 180°. Figure 10 shows symbols for the 17 samples in the Warwick dome across the swirl for which a determination of shear sense was attempted (DelloRusso and Robinson,1989). The westernmost 9 samples show a shear sense consistent with upward movement of the core of the dome relative to the cover, and this is consistent with the shear sense in the Northfield syncline observed by Reed. Of the easternmost 7 samples, 3 are consistent with those to the west but are left-lateral due to warping over the crest of the dome, 3 are inconsistent with it and one is indeterminate. When the patterns of Pennsylvanian kinematic indicators determined by Reed (1992) are combined with those determined by DelloRusso and Robinson (1989) in the Warwick dome, a story emerges of a layer of rock about 2-4 km thick including much of the Warwick dome and Kempfield anticline that flowed southward during Pennsylvanian deformation relative to the rocks of the Pelham dome structurally below and relative to the preserved sillimanite-grade Acadian rocks above. The conditions that permitted this strain field to develop, particularly in the part of the Ordovician "basement" and its Bronson Hill cover closest to the Pelham dome, are not clear. It could have resulted from earlier Pennsylvanian underthrusting of a hot Avalonian basement, if evidence against late emplacement contained in structural relationships and in contact relations of the Belchertown intrusion, can be successfully deflected.

BASIS FOR TECTONIC MODELS

A successful tectonic model must explain how the various rock layers came to be in their present positions, how the various observed major and minor structural features were formed, and how the various rocks came to be metamorphosed when they did. Before discussing any models, a series of salient facts need to be reviewed and evaluated.

Situation of the Proterozoic Rocks

The Proterozoic rocks in the inner part of the dome appear similar to rocks of the Hope Valley part of the Avalon zone in southeastern Connecticut and western Rhode Island. In southeastern Connecticut Gromet (1989) has indicated that these rocks show no evidence of metamorphic recrystallization between the time of Avalonian igneous crystallization and late Paleozoic gneissic deformation associated with various shear zones. In the Pelham dome, Robinson had felt that the coarse microcline megacrysts typical of the Dry Hill and Poplar Mountain Gneisses and associated quartzites were a reflection of the same granulite-facies metamorphism shown by the schists of the Mount Mineral Formation. He further explained the preservation of high orthoclase megacrysts in the Mount Mineral versus maximum microcline in the Dry Hill and Polar Mountain as due to the highly aluminous nature of the former, causing protective muscovite reaction rims to form in the Mount Mineral. However, despite several efforts, there is nothing in our present data base, with the exception of one zircon at Stop 5, to prove that the Dry Hill and Poplar Mountain Gneisses were subjected to the same Acadian granulite-facies metamorphism as the overlying Mount Mineral schists, and several isotopic arguments against it. With this in mind, Gromet is strongly inclined to believe that the Proterozoic rocks were structurally emplaced beneath the Ordovician "basement" and Ordovician-Silurian-Devonian cover after the Acadian metamorphism, i.e. in the late Paleozoic. Robinson and Tucker, on the other hand, still hold out the hope that better isotopic evidence for an Acadian event will yet be found, and cite indirect structural arguments that make a late Paleozoic juxtaposition implausible.

The east-directed fold nappes of the Pelham dome clearly involve Lower Devonian cover. While it cannot be absolutely proved that this folding is also present in the Proterozoic rocks, the structural relations of the Dry Hill Gneiss strongly imply that it lies in the core of an east-directed fold nappe of the same type. In the northern part of the dome there is no proof that this family of recumbent folds is not as young as Pennsylvanian. In the southern part, unfortunately in an area of poor outcrop, it appears geometrically most probable that these recumbent folds are truncated by the Belchertown Quartz Monzodiorite intrusion with an igneous zircon age of 380 ±5 Ma. If correct, then this whole recumbent fold system must be earlier in the Acadian, probably synonymous with the backfold stage (Hall and Robinson, 1982), and the Proterozoic rocks must have been already in place.

While Gromet's arguments that the Proterozoic rocks are from Avalon are attractive, one must not rule out the possibility that they correspond to the Yonkers and Pound Ridge Gneisses of the North American side (Hodgkins, 1985). In support of this view is zircon detritus of generally Laurentian affinity. In favor of an "eastern" origin are the recently discovered west-over-east shear fabrics in the Mount Mineral schist, but these are high-temperature fabrics most probably associated with waning phases of Acadian metamorphism.

Contacts of the Mount Mineral Formation

The contacts below, within, and above the Mount Mineral Formation are critical to all models. At present the formation is viewed as a stratigraphic-tectonic amalgam of several elements, a basal quartzite that may well belong to the underlying Proterozoic section, a middle member of sulfidic schists, amphibolites, and plagioclase gneisses that most closely resemble the Late Ordovician Partridge Formation with a few slices of Late Ordovician intrusive "basement", and an upper member of muscovite-garnet quartzite with minor gray schist that is best assigned to the Lower Silurian Clough Quartzite, possibly with some thin layers of Lower Devonian Littleton Formation. So far, only the "Partridge" and "Clough" parts of the Mount Mineral have yielded evidence for Acadian granulite-facies metamorphism.

Major questions about the Mount Mineral Formation include the following:

A) What is the nature of the contact near the base separating rocks of the Acadian granulite facies from the underlying Proterozoic with its apparent lack of Acadian metamorphism?

B) What is the nature of the contact at the top of the Mount Mineral Formation separating the Acadian granulite-facies Silurian cover from the Ordovician intrusive basement apparently lacking evidence for the granulite-facies metamorphism? This contact can be an inverted unconformity, but is more probably a thrust or detachment fault, which could explain both the stratigraphy and the apparent metamorphic discontinuity.

Indeed, the situation is so complex that it may not be worthwhile to speculate at this time on the necessary sequence of events to provide these juxtapositions.

Contact between Late Ordovician Basement and Ordovician-Silurian-Devonian Cover

The contact between the Late Ordovician plutonic basement and its Late Ordovician Ammonoosuc and Partridge cover has its own distinct problems as discussed at length by Schumacher(1988) and more recently by Tucker and Robinson (1990). Radiometric ages appear to preclude the possibility that this is an unconformity as proposed earlier, and Robinson (Tucker and Robinson, 1990) postulated a late Taconian detachment fault or an early Acadian fault. More recently, Kohn and Spear (1991) have proposed this as a late-metamorphic Acadian thrust or detachment surrounding several domes in western New Hampshire. This contact occurs over a large distance around and within the Pelham dome, and is intricately deformed by the east-directed recumbent folds. Its origin must be factored into any model.

Cover Stratigraphy and Facies Relations

The cover sequence of the Pelham dome features extensive areas of Erving Formation at the top of the sequence (Robinson et al., 1988), but locally cutting down to come in contact with the Fourmile Gneiss. The stratigraphic sequence exposed in the Pelham-Shutesbury syncline lacks Erving Formation and is dominated by a variety of Littleton Formation not seen nearby. Any successful model should explain the subsurface connections between these different facies of cover strata and attempt to explain their relative positions of primary deposition.

Metamorphic Zones, and Acadian and Pennsylvanian Recrystallization

The pattern of Acadian isograds before Pennsylvanian overprinting is extremely difficult to discern. The relict Acadian metamorphism in the Mount Mineral Formation was at as high a temperature and at a higher pressure than anything that has been observed in the central Massachusetts granulite-facies high, based on the high pyrope content of relict garnets and the abundance of garnet-rutile assemblages (Roll, 1987). Yet coming out of the Pennsylvanian overprint toward the east, one passes through a classic prograde sequence beginning with the sillimanite-staurolite-muscovite zone. Is it possible that the cover-sequence strata were even lower grade before the Pennsylvanian overprint, and that only the Mount Mineral strata were involved in an Acadian high-grade metamorphism and achieved their present outcrop position as a result of complex faulting? One creative possibility is that the Mount Mineral schists are actually a tectonically trapped deep-seated western tail of the central Massachusetts high that has been overridden by an enormous east-directed thrust sheet of Bronson Hill basement along a western equivalent of the middle to late Acadian Conant Brook shear zone (Peterson, this guidebook), which has been arched subsequently over the crest of the Pelham dome. The evidence that the dominant metamorphism in the Pelham dome was Pennsylvanian, also requires a complete reevaluation of the status of metamorphic overhangs along the east margin of the Connecticut Valley synclinorium (Robinson et al, 1991).

"Dome-Stage" Lineation

Both the Acadian (Peterson, this guidebook) and the Pennsylvanian lineations are parallel to the axis of the orogen and imply transport and elongation parallel to the axis of the orogen during two widely spaced episodes. The "great swirl" is most easily explained as the superimposition of a Pennsylvanian strain gradient quasi-parallel to the older Acadian lineation, rather than a swirl within a single field of deformation. The concept of such a reorientation is better understood by examining and extrapolating the relationships on a grander scale, as shown by Peterson (this guidebook, Figure 8) between an earlier Acadian west-over-east thrusting followed by a later Acadian right-lateral shear.

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ROAD LOG

Trip will assemble at University of Massachusetts Football Stadium at 8 A. M. and an attempt will be made to consolidate into as few vehicles as possible. The formal road log begins on Route 63 at the entrance to the Northfield Mountain Pumped Storage Hydroelectric Project and. To reach this location from U. Mass. Stadium, the following notes are provided.

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.
14.8	Cross Millers River in Village of Millers Falls (Figure 11).
15.2	Cross beneath Route 2.
17.3	Turn right into Northfield Mountain Project and **begin formal road log.**
0.1	Turn right (south) into Visitor Center
0.2	Stop at Visitor Center parking lot to pick up escort. This is about 25 minutes drive from Amherst. The next 6.7 miles and STOPS 1, 1A, and 2 will be within the grounds and on roads of the Project not normally open to the public. Turn around and return north to stop sign.
0.3	Stop sign at exit of Visitor Center. Turn right (east) on private road.
0.5	Fork. Bear left toward warehouse. Right turn leads to gate entrance to powerhouse 1/2 mile underground.
0.6	Three-way junction. Bear right (south) toward warehouse.
0.7	Turn around in warehouse lot and park for Stop 1A (Figure 11).

STOP 1. DRY HILL GNEISS HORNBLENDE MEMBER OVERLAIN BY POPLAR MOUNTAIN QUARTZITE AND GNEISS (25 MINUTES) The man-made outcrops extend from the parking lot east and then south around the warehouse toward the portal of the Access Tunnel to the underground powerhouse. At this location we are at the top of a west-dipping layer of Dry Hill Gneiss 750 feet thick (Figure 4). The bottom of the layer is exposed in the roof of the underground powerhouse. Both here and there, the Dry Hill Gneiss is indirect contact with well bedded quartzite. The next unit upward here and downward there is the Poplar Mountain Gneiss. This apparent "mirror image" stratigraphy led Ashenden (1973) to propose that the Dry Hill Gneiss forms the core of a giant fold nappe, although a hinge has not been identified. Such fold nappes had already been identified by Robinson (1963) at a higher level in the dome, involving Fourmile Gneiss and cover units. Because these folds are presently east-directed, the postulated fold involving the Dry Hill Gneiss is also tentatively shown that way (Figure 4). The lower side the Dry Hill Gneiss is dominated by the biotite member, but this appears to die out by facies change and at Stop 1 the Hornblende Member is in direct contact with the overlying Poplar Mountain Quartzite. The contact is exposed in the drainage area at the bend in the fence. For a detailed description of the Dry Hill Gneiss Hornblende Member see Stop 2, for the Polar Mountain Quartzite see Stop 6, and for the Poplar Mountain Gneiss see Stop 5. The contact between the Quartzite and the overlying Gneiss is near the Warehouse Parking lot and it is locally repeated by a northwest-plunging late asymmetric fold.

Particularly interesting from the point of view of this trip are multi- and single-grain zircon analyses from a sample of Dry Hill Gneiss collected behind the Warehouse. Unlike the sample from the mountain top (Stop 2), the zircons from this sample suggest a complex history of igneous growth with inheritance, and possible secondary zircon growth or Pb-loss in Pennsylvanian time (Figure 6A). The multigrain analyses form a poorly defined discordia, with concordia upper- and lower-intercept ages of 1747 Ma and 500 Ma, respectively. The upper-intercept age of 1747 Ma could be interpreted as the *mean* age of an inherited zircon component, if all grains consist of a two-component mixture without appreciable secondary (post-600 Ma) Pb-loss or new zircon growth. A U-Pb analysis of faceted sphene from this sample produced a concordant age of 293 ±5 Ma (Tucker and Robinson, 1990), which is clearly younger than the zircon lower-intercept age and best interpreted as the time of metamorphic sphene growth.

Examination of polished zircon, etched in HF vapor, reveals that many grains are composite in structure, consisting of small turbid and cracked cores mantled by clear faceted overgrowths. Figure 6A demonstrates that zircon inheritance, manifest as cores, comprises many ages as seen by the resolvable and very different $^{207}Pb/^{206}Pb$ ages in 15 single-grain analyses. Two grains yielded near concordant (<2% discordant) analyses at the presumed eruption age of the rhyolite protolith (ca. 600 Ma, Stop 2), whereas another two yielded near concordant analyses at the known time of sphene growth (ca. 300 Ma). The latter grains have depleted Th/U ratios (0.02) suggesting that they are nearly pure metamorphic zircon which grew in a thorium-depleted environment. All other analyses are discordant with old $^{207}Pb/^{206}Pb$ ages suggesting, that they consist of an inherited core-component and an igneous or metamorphic overgrowth. Because the presumed purely igneous grains have experienced only minute amounts of Pb-loss or new zircon growth at ca. 300 Ma, it is possible *to approximate* the inheritance age of each composite-grain analysis by constructing a chord from the presumed time of eruption (ca. 600 Ma) through the discordant analysis to its upper intersection with concordia. Based on this method of estimation, three groups of inheritance ages are identified: Group I, 2850-2550 Ma; Group II, 2150-1850 Ma; and Group III, 1350-1220 Ma (Figure 6A, Tucker and Robinson, 1991)..

After stop return north to three-way junction.

0.9 Stop in paved road near three-way junction to examine road cut on right.

STOP 1B. SILL OF BELCHERTOWN QUARTZ MONZODIORITE GNEISS AT BASE OF FOURMILE GNEISS (5 MINUTES) The rock at this outcrop, composed of plagioclase An22, quartz, microcline, biotite and hornblende with minor epidote, allanite, sphene, apatite, zircon, ilmenite, and chlorite, was originally mapped by Ashenden (1973) as part of the Fourmile Gneiss very close to its base. As such, it was sampled by Robinson and G. W. Leo as part of an attempt to date the Fourmile Gneiss by R. E. Zartman. Zartman reported a preliminary U-Pb age on zircon of 370 Ma, which caused an immediate re-evaluation. Robinson reviewed notes of his underground work in the Tailrace Tunnel and suggested that the rock is a homogeneous sill of Belchertown Quartz Monzodiorite, 34 feet thick, exactly on the contact between the Fourmile Gneiss and the underlying Poplar Mountain Gneiss. Major- and trace-element analyses of surface and subsurface samples by Hodgkins (1985) confirmed the correlation of this rock with other gneissic rocks in the outer recrystallized part of the Belchertown intrusion. Larger sills of similar rock in Fourmile Gneiss have been mapped in the southern and western parts of the dome.

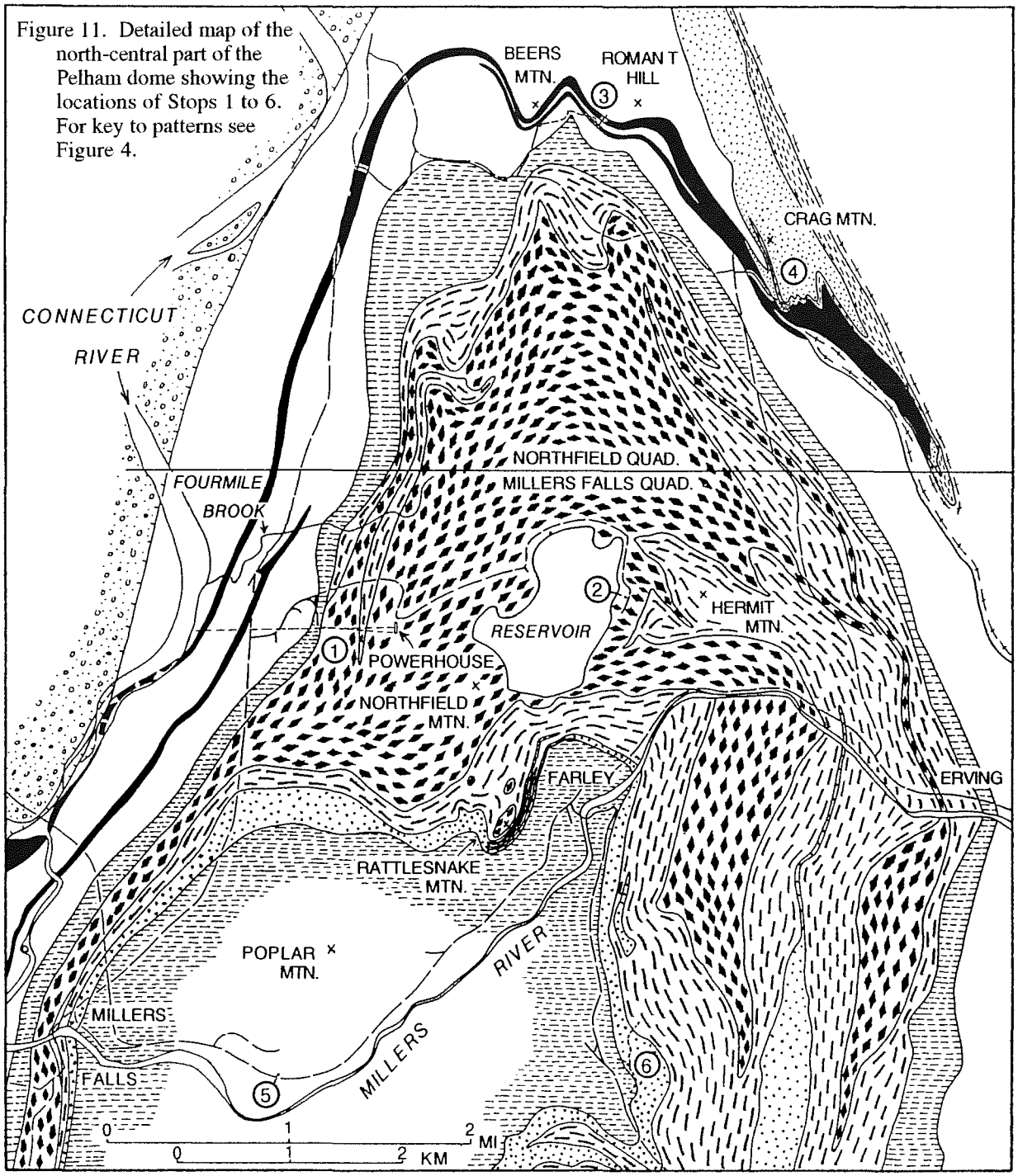

Figure 11. Detailed map of the north-central part of the Pelham dome showing the locations of Stops 1 to 6. For key to patterns see Figure 4.

The rock contains the same north-south trending mineral lineation as all other rocks in the Pelham dome, and like many other nearby rocks contains metamorphic sphene dated at 300 Ma (Figure 8E). Gromet has obtained an essentially concordant analysis of igneous zircon indicating a crystallization age of approximately 374 Ma, in good agreement with Zartman's determination. In addition, Gromet has analysed two very different varieties of sphene from this sample: one consisting of coarse (>150 μm), cloudy anhedral crystals, and the other of fine (< 75 μm) clear crystals. One interpretation of these data (Figure 8E) is that the coarse, cloudy sphenes are igneous crystals that experienced partial radiogenic Pb-loss (or new growth) at ca. 300 Ma, the time of new metamorphic growth of the fine, clear variety. Alternatively, the coarse cloudy crystals may represent an older generation of metamorphic sphene that is discordant, in part, because of partial Pennsylvanian Pb-loss or uncertainty in the initial-Pb isotopic composition. In the main Belchertown intrusion (Ashwal et al., 1979), all sphene appears to be of metamorphic origin, and Gromet's analysis of coarse, cloudy sphene may provide evidence for Acadian metamorphism of this rock.

Spear and Harrison (1989) reported a 350 Ma total-gas Ar age for metamorphic hornblende in this rock. Although difficult to interpret, it probably represents a meaningless determination that is older than the true metamorphic age because of the presence of extraneous radiogenic Ar.

Proceed through locked gate and up mountain. Numerous low cuts in dip slope of Dry Hill Gneiss.

1.8 Sharp left bend in road with radio tower to south. This lies directly above the underground powerhouse. Cores from two diamond drill holes more than 900 feet deep were used by Ashenden and Robinson to establish a detailed stratigraphy for the lower contact region of the Dry Hill Gneiss. This was later studied in outcrop on the face of Rattlesnake Mountain and will be seen on the southwest face of Bear Mountain at Stop 6.

2.5 Top of hill where pavement turns right. Turn left on gravel road that runs outside the north dike of the Northfield Mountain Reservoir.

3.0 Northwest corner of the dike. Approximate location where Robinson and Ashenden in 1968 collected "The Blob", a 250 pound exotic block within the Dry Hill Gneiss in the bedrock surface that was exposed during construction of the dike. A sawed section of this was subsequently "mapped" at 1:1 by T. N. Lincoln (unpublished Senior honors thesis, 1977) and probe analyses of coexisting minerals were performed by Tracy et al. (1980, see also Robinson, 1980, Figure 42). "The Blob" consists of highly deformed well bedded quartz-garnet-cummingtonite rock with various other chain silicates including manganoan hedenbergite, rhodonite and pyroxmangite, as well as minor magnetite and pyrrhotite. There is a hornblende-rich reaction rind against the Dry Hill Gneiss. Photographs show clear asymmetric tails indicating the same north over south shear sense typically recorded around microcline megacrysts. "The Blob" is the most exotic of a whole series of strange blocks including hornblendites, amphibolites, and calc-silicates that are typically present both in the Dry Hill Gneiss and the Poplar Mountain Gneiss. Some may be boudinaged layers or dikes, others may be fragments of country rocks through which the Dry Hill magmas passed en route to the surface. "The Blob" most closely resembles gangue associated with ore bodies at Franklin Furnace, New Jersey and Broken Hill, Australia!

3.7 North side of broad bedrock spillway on right. Stop in road and walk 100 feet west to nearest outcrop

STOP 2. DRY HILL GNEISS, HORNBLENDE MEMBER, AT AXIS OF PELHAM DOME (30 MINUTES) This outcrop is slightly east of the crest of the Pelham dome, which lies approximately within the reservoir (Figure 11). It is approximately in the center of the Hornblende Member of the Dry Hill Gneiss where it is about 1000 feet thick in the core of an interpreted east-directed recumbent anticline. The rock is typical of the Hornblende Member, with interlayered gray-pink biotite-feldspar gneiss and pink hornblende leucogneiss. Its petrography, mineralogy and geochemistry are covered in detail by Ashenden (1973) and Hodgkins (1985). The rock typically contains 25-30% quartz, 45-50% microcline, 5-10% oligoclase, 5-15% green biotite, 5-15% hastingsite, and accessory allanite, sphene, epidote, calcite, garnet, apatite, zircon, and pyrite. It typically contains widely spaced megacrysts of pink maximum microcline (Laird, 1974) that may be as large as 15 cm across with local graphic quartz intergrowths suggesting the megacrysts may be fragments from tectonically dismembered pegmatites. In thin section, the hastingsite is dark blue-green in one direction and shows a negative 2V as low as 5°. The tiny garnets appear to show metamorphic growth zoning with cores Alm 52, Spess 15, Pyr 1.5, Gross 28, And 3; and rims Alm 58, Spess 11.5, Pyr 1.5, Gross 26, And 3.

The bulk composition is typical of an A-type alkalic granite such as the Pikes Peak or Cape Ann, or an equivalent rhyolite, and is very close in most respects to the Potter Hill Gneiss in the gneiss domes of the Hope Valley terrane in southeastern Connecticut (Hodgkins, 1985). A characteristic of all these rocks is an extremely high FeO/MgO ratio, in some samples so high that MgO is below detection limits, although still above 1% in probe analyses of the hastingsite, which however contain up to 30% total Fe as FeO. On the basis of its layering and interbedding with quartzites and calc-silicates, the Dry Hill is considered to be a volcanic sequence produced in an extensional environment. The hornblende leucogneiss layers give the impression of being some kind of partial melt segregation produced during high-grade metamorphism. However, analyses show they have higher normative anorthite; lower Na_2O; higher K_2O; over 100 ppm more Ba and Zr; higher Ce, Nb, and Y; and lower Ti and V; all features not to be expected in partial melts. Hodgkins argues that the leucogneiss layers may be beds of crystal tuff where phenocrysts of orthoclase, quartz, hornblende or pyroxene, zircon, and sphene concentrated with respect to glass representative of the liquid from which they crystallized.

This outcrop is near the locality that was sampled by Naylor et al. (1973), and is the exact location of the samples studied by Gromet (Gromet and Robinson, 1990); Harrison et al. (1989) and Tucker and Robinson (1990). Unlike the results reported for Stop 1, the multigrain zircon analyses from this locality form a well defined discordia indicating that inherited zircon components were successfully avoided during grain selection. All analyses are discordant over a moderate range (15-31%) and two of the analyses of clear faceted sphene are concordant at 292 ±5

Ma. If regressed with the sphene analyses, the data define intercept ages of 613 ±3 Ma and 289 ±4 Ma, interpreted by Tucker and Robinson (1990) as the time of igneous crystallization and metamorphic recrystallization, respectively.

Evidence of Pennsylvanian isotopic resetting as a result of recrystallization accompanying strong ductile deformation also comes from the Rb-Sr, Nd-Sm and U-Pb studies of Gromet. A Rb-Sr mineral and whole-rock isochron (Figure 8A) suggests recrystallization at 292±5 Ma, in good agreement with the U-Pb sphene ages reported above, although biotite and epidote appear to have exchanged radiogenic ^{87}Sr at a later date (Figure 8B). Apparently allanite did not fully participate in the isotopic exchange at this time, an observation also borne out in the U-Pb data which clearly identifies an old radiogenic component in what is interpreted as relict allanite (Figure 8D). The Sm-Nd data indicates near complete homogenization of the rare earth isotopes and growth of new minerals, such as garnet, at this time, as confirmed by the close approximation of all minerals to a 302 ±20 Ma reference line (MSWD =11). A hornblende ^{40}Ar-^{39}Ar plateau age of 287 ±1 Ma (Harrison et al., 1989) is also consistent with the view that Late Pennsylvanian metamorphic recrystallization took place at this outcrop.

The prominent mineral lineation in this outcrop trends north-south parallel to the dome axis and to the mineral lineation in mantling strata of the dome. In many outcrops the mineral lineation has suffered great circle rotation by a series of variably oriented late asymmetric folds that have been studied in detail by Ashenden (1973) and Onasch (1973). High on the eastern end of this outcrop there is a well exposed late asymmetric fold in foliation trending northwest and overturned to the southwest. On the top of the outcrop, it can be seen that an axial plane foliation transects the fold and the strong north-south mineral lineation lies within the foliation plane. Thus, development of the lineation preceded development of some asymmetric folds and followed others, and appears to have been part of the same process. At several outcrops, asymmetric folds were measured that are consistently overturned to southeast, south, or southwest and vary in orientation over an angle of about 160°, so that they show a "separation angle" of 20°. The mineral lineation bisects the separation angle, indicating that the mineral lineation is parallel to the transport direction of the late asymmetric folds. The folds were in the process of being rotated into parallelism with the lineation, but did not achieve parallelism so completely here within the dome as they apparently did in the surrounding mantling strata. The consistent pattern of asymmetry of these folds throughout the dome, regardless of structural or stratigraphic position, implies relative southward sliding of the cover of the dome relative to the core during doming. Such a shear sense parallel to the lineation is also consistently demonstrated by a wide range of asymmetric microfabrics described in detail by Reed and Williams (1989) and Reed (1990,1992). These include, in particular, asymmetric tails on microcline megacrysts in the Dry Hill and Poplar Mountain Gneisses, but also asymmetric growth fabrics in garnet, staurolite, and kyanite in the Littleton and Erving Formations of the cover. These studies combined with the isotopic dating of metamorphic minerals and minerals in pegmatites implies that this major strain field is of Pennsylvanian age. Also on the eastern end of the outcrop there is a late cross-cutting pegmatite dike truncating a late asymmetric fold, but itself slightly deformed and foliated. Isotopic dating of a similar dike is in progress.

After stop continue south to junction where direction can be reversed.

4.0 Pass by stop 2 again.
4.2 View of Mount Greylock ahead to left. Return along north dike.
4.3 Return to pavement and turn sharp right down the mountain.
6.5 Three-way junction. Bear right.
6.7 Yield sign. Stay right.
6.8 Entrance to Visitor Center. Leave off escort and go straight.
7.0 Return to Route 63. Turn right (north).
7.6 Bridge over Fourmile Brook in Village of Northfield Farms. Type locality of Fourmile Gneiss is just down the brook.
9.5 Sharp left (east) on narrow, steep, poorly marked, paved road, just in view of large silo and barn on the left.
9.8 Soccer field on right.
9.9 Pavement ends.
10.1 T junction. Turn right (east) on pavement of South Mountain Road at Linden Hill School.
10.3 Road cut on left in Fourmile Gneiss showing road is on dip slope.
10.6 Fourmile Gneiss on left at top of very steep section. Beyond road flattens, then begins descent.
10.7 Driveway to left (Obscure sign- Dresser's Sunnydale Farm) leads 0.35 mile to farm and beginning of Stop 3 . Depending on parking conditions, it may be necessary to ferry participants to the farm in vans.

STOP 3. RECUMBENT SYNCLINES OF PARTRIDGE FORMATION IN FOURMILE GNEISS AT ROMAN T HILL QUARRY(45 MINUTES) From north side of farmhouse, walk east across fields (see Figure 11), in the process crossing the axis of the Pelham dome, and cross barbed wire into edge of woods. Walk north along edge of woods to old quarry road that runs east across swampy ground. At junction on east side of swamp go straight onto very steep ramp trail that bears up and to the right. When trail is within 30-40 feet of quarry cliff on left, climb straight up to base of cliff.

Cliff is composed of gray Fourmile biotite-feldspar gneiss, dipping gently northeast, that lies in the core of an east-directed recumbent anticline (Roman T Hill anticline) or thrust sheet (Figure 4). Note strong gently north-plunging lineation of elongate quartz and biotite aggregates. There is an unconfirmed rumor that the abutments for the Brooklyn Bridge came from these quarries. Follow animal trail southeast around base of cliff to slightly lower overhang exposing 10-20 foot layer of Partridge Formation biotite-muscovite schist with pegmatite that forms an isoclinal syncline (Beers Mountain syncline) between layers of Fourmile Gneiss. This very thin isoclinal layer has been followed to an apparent hinge on the west face of Beers Mountain. A second hinge appears near Fourmile Brook and the layer has been traced thence southward through the Northfield Project to the vicinity of Montague, a total distance of 10 miles. Note late asymmetric minor folds and indicators for north-over-south shearing.

Return back toward gneiss quarry and then climb up steep dirt ramp to top of cliff. Walk about 100 feet northeast to small cliff band in an upper layer of coarse muscovite-biotite schist of the Partridge Formation. This is also an isoclinal syncline (Tailrace syncline) and has been traced from the Northfield syncline near Crag Mountain southeast of here continuously around the entire northern part of the Pelham dome to the vicinity of French King Bridge, where it is truncated by the Connecticut Valley border fault, a total distance of 8 miles. This syncline is named for its exposure in the Tailrace excavation of the Northfield Mountain project, where it may still be seen just above river level. The Partridge rocks in this syncline, commonly no more than 20 feet thick, completely isolate the area of Fourmile Gneiss to the north from the rest of the Fourmile Gneissin in the dome. This isolated area of Fourmile Gneiss is itself a recumbent anticline, the Jacks Brook anticline. The type locality of the Fourmile Gneiss, as defined by Ashenden (1973) in the bed of Fourmile Brook at Northfield Farms, lies in the gneissic belt between the Beers Mountain and Tailrace synclines. Climb above schist ledges to smaller outcrops of Fourmile Gneiss above the schist syncline near summit of Roman T Hill.

Return by **exactly the same route** to quarry to farmhouse and end of Stop 3. It is very easy to become lost in cliffs and steep loose blocks on this hill.

Return from farmhouse to South Mountain Road. Proceed east on South Mountain Road.
11.4 Fork in road. Stay left on pavement and cross beneath powerline.
12.1 Four-way junction. Go straight up hill on Sky Farm Road.
12.4 Crest of hill. Park to right and walk up driveway on left (north) side for outcrops of Stop 4.

STOP 4. CLOUGH QUARTZITE AND DEFORMED CONGLOMERATE, LITTLETON SCHIST AND FOURMILE GNEISS IN EARLY RECUMBENT FOLDS AT CRAG MOUNTAIN (1 HOUR 15 MINUTES INCLUDING LUNCH) Crag Mountain and Brush Mountain are held up by massive Clough Quartzite and conglomerate involved in a series of east-directed recumbent folds that involve all stratigraphic units from the Fourmile Gneiss to the Lower Devonian Erving Formation. The recumbent folds were apparently earlier than the doming and their axial surface were arched by the doming. Data on the axial trends of the recumbent folds is scanty, but, on the basis of widely spaced hinge locations, they appear to trend north-northeast at a moderate angle to the dome axis (see Laird, 1974, Fig. 20).

The lowest of the recumbent folds is the Beers Mountain recumbent syncline of Partridge Formation seen at Stop 3. This is followed by the Roman T. Hill recumbent anticline and the Tailrace recumbent syncline, also seen at Stop 3. Above these is the Jacks Brook recumbent anticline described for Stop 3. The hinge of Fourmile Gneiss in this fold is located close to where we will walk on Crag Mountain (Figure 12), and it also has a hinge in the Erving Formation just northeast of Erving. There is ample evidence that the lower limb of the anticline is attenuated and may be thrust-like in places. Thrusting is also suggested where the Clough of the upper limb is dragged around the hinge of Fourmile Gneiss south of Crag Mountain. The thick zone of highly folded Clough of Crag and Brush Mountains is a series of isoclinal recumbent folds developed on the upper limb of the Jacks Brook recumbent anticline. The thin Crag Mountain recumbent syncline of Littleton Formation (Figure 12) occurs between layers of Clough south and west of Crag Mountain and hinges out northwest of the summit. Isolated

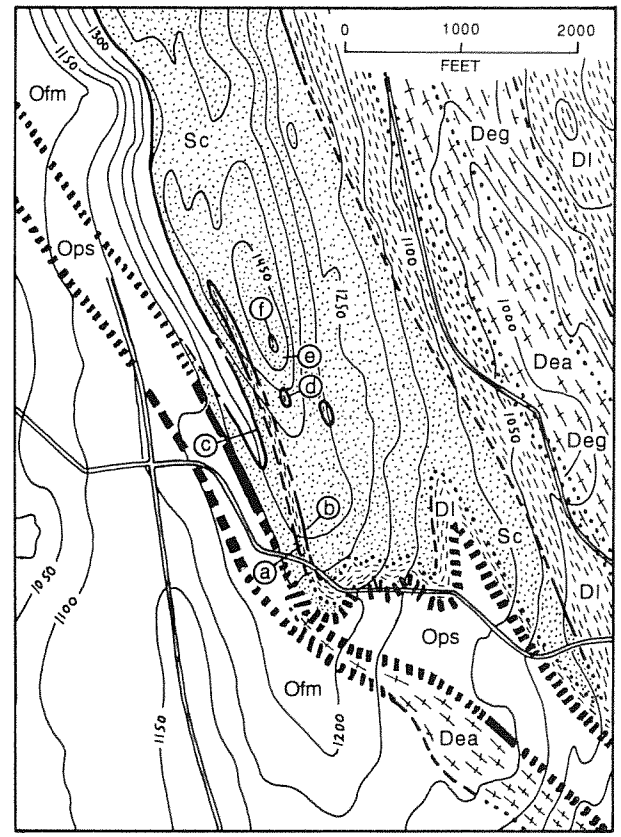

Figure 12. Detailed map of the Crag Mountain area showing stops on the walking log. Ofm - Fourmile Gneiss; Ops - Partridge Formation; Sc - Clough Quartzite; Dl - Littleton Formation; Deg - Erving Formation Granulite member; Dea - Erving Formation Amphibolite Member.

outcrops of Littleton in Clough on the mountain are probably tectonic inclusions along axial surfaces of higher recumbent synclines. Follow driveway, which is also the Metacomet Trail (white blazes) northward from road to first outcrop on left.

Station a. Typical coarse gray muscovite-biotite-garnet-staurolite-graphite-ilmenite schist of Littleton Formation in Crag Mountain syncline (Figure 12). North-plunging crenulation and mica lineation. The high Fe ratio explains the absence of kyanite in this unit as compared to nearby schist beds in the Erving Formation (Hall, 1970). This rock is identical to Littleton on the east limb of the Northfield syncline, which contains typical growth-zoned garnets (Tracy, Robinson and Thompson. 1976; Robinson, 1991, Figure 6). Reed and Williams (1989) have shown that garnets, and staurolites were growing in these rocks during north-over-south shearing. Recently Tucker has obtained a concordant monazite age from the same rock of 295±3 Ma, strongly suggesting that the staurolite grade metamorphism and concurrent shearing were Pennsylvanian.

Station b. Foliated and lineated pegmatite sill with broken feldspars in Clough on east limb of Crag Mountain syncline. Slightly to north and east is deformed fine conglomerate. Folds outlined by thin garnet-bearing beds have a strong east-dipping axial plane foliation formed by the flat direction of sword-like stretched pebbles with long axes parallel to fold axes.

Proceed north on driveway noting Clough on east and Littleton with pegmatite to west. Where driveway turns left and Metacomet Trail turns right, go straight on older trail. Where older trail turns right (east) up steeper slope, turn sharply west off trail to outcrops on brink of hill 75 feet away.

Station c. Clough Quartzite containing a foliated pegmatite sill is here in direct contact with Fourmile Gneiss of the Pelham dome, in the core of the Jacks Brook recumbent anticline (Figure 12). The contact, well exposed for three miles north of here, is interpreted as an unconformity (Emerson, 1898; Balk, 1956; Robinson, 1963). The Fourmile Gneiss is a dark variety abundant near the north end of the dome, that consists of quartz, andesine, microcline, biotite, and hornblende with accessory epidote, garnet and apatite. It shows the same north-south lineation common in this region. Return to trail and proceed up hill.

Station d. Tectonic inclusion of Littleton in Clough (Figure 12). Pieces under trail have slumped slightly to give south-plunging lineation.

Station e. The Crag. Conglomerate of Clough Quartzite. View to east, southeast, and south. Valley at east base of mountain is eroded in Erving Formation in the center of the Northfield syncline. First Bald Hills (heavily wooded) beyond are held up by Littleton on the east limb of the syncline. According to Reed (1992), the north-over-south shear sense in the Pelham dome continues unchanged by the axial surface of the syncline, so that, on the overturned east limb,. the rocks of the Warwick dome were moving south over the cover. The dark mountain to the east-northeast is Mount Grace (1617'), formed of Ammonoosuc Volcanics up to 4000' thick in a cross fold on the east flank of the Warwick dome. The depression between First Bald Hills and Mt. Grace is eroded in the south lobe of the Warwick dome, consisting of massive intrusive-looking gneiss, and a nearly continuous rim of high ground can be seen connecting around the south end of this lobe. A linear valley south of the Warwick dome is eroded in the gneiss of the Kempfield anticline. To the southeast, the Orange airport lies in the center of the broad valley

eroded in the main body of Monson Gneiss. This valley is virtually continuous from northern Massachusetts to southern Connecticut. Mt. Monadnock (3165') with its satellites Little Monadnock (1883') and Gap Mountain (1862') lie to the right of Mt. Grace. Farther on the eastern horizon are the Pack Monadnock Group, Watatic (1832') and Wachusett (2006') on the west contact of the Fitchburg plutons.

Station f. Summit of Crag Mountain (1503'). Best examples of stretched pebbles and cobbles, now believed to be oriented as a result of Pennsylvanian shearing. This is the type area for "northfieldite" a siliceous border of the "Pelham granite" (Emerson, 1915. 1917) here in its "pseudoconglomeratic" facies. (No Hammering Please - best samples are collected on trail south of station e.)

Extensive view to west and southwest. Partridge schist in Beers Mountain and Tailrace synclines is exposed on face of Beers Mountain behind farm. To the southwest, the north-plunging foliation arch of the Pelham dome is expressed in topography around the Northfield Mountain Reservoir (Stop 2). Beyond can be seen Mt. Toby, Mt. Sugarloaf, and Mt. Tom. The Mesozoic border fault runs from Mt. Toby northward near the west edge of the Pelham dome, passing about one mile west of Beers Mountain. Beyond the Connecticut Valley rise the Berkshire Hills and the Green Mountains including Haystack and Stratton Mountains. On the far horizon is Mt. Greylock (3,491'), the highest point in Massachusetts, and an eastern Taconic slice.

The Clough conglomerate poses two interesting problems, one concerning the origin of the rock and one concerning the deformation of the pebbles. The pebbles are more than 90% vein quartz with minor quartzite and black quartz-tourmaline vein material. A single pebble with sand-size clasts may be seen near the summit bench mark. What were the conditions under which such a virtually monolithologic conglomerate could form?

The pebbles and cobbles have the form of sword-blade-like ellipsoids or cigars with extreme elongation parallel to fold axes. On one excellent outcrop, about five miles north, on the east limb of the Northfield syncline, average axial ratios of 1 : 5 : 27.5 were measured. Compared to an original sphere of equal volume and radius 1, the statistical pebble ellipsoid (Rosenfeld, 1954) at this location has axial ratios of 0.19 : 1.01 : 5.24, indicating elongation parallel to fold axes by a factor of five. Earlier we have seen that this lineation is indeed the transport direction of a group of asymmetric folds inside the Pelham dome that are strongly rotated toward the lineation. Here at Crag Mountain, all folds seem to be coaxial, suggesting either that they formed parallel to the transport direction or were almost total rotated into parallel to it. These features combined indicate both shearing and elongation during the Pennsylvanian, parallel to the axis of the orogen.

Return quickly down the trail and driveway to Sky Farm Road.

Following Stop 4 do U-turn at driveway and return to four-way junction.

12.8 Four-way junction. Turn sharp left (south). Pavement ends.

13.7 Erving Town Line (sign gone). Pavement begins. Descend Mountain Road to valley of Millers River.

15.7 Junction with Route 2. Turn right (west).

17.6 Village of Farley, Town of Erving. Cliffs of Rattlesnake Mountain to right expose base of Dry Hill Gneiss with structurally underlying Poplar Mountain Quartzite and Gneiss; a classic but strenuous traverse.

18.2 Junction with Old State Road. Stay left on Route 2.

19.6 Begin passing lane.

19.8 Large road cut on both sides. Pull off to right as far as possible on soft shoulder near beginning of exposure. LEAVE FLASHERS ON. STAY COMPLETELY OFF PAVEMENT Best access to top of exposure is short trail at east end

STOP 5. POPLAR MOUNTAIN GNEISS AT DEEPEST EXPOSED LEVEL IN THE PELHAM DOME (20 MINUTES) This major road cut lies almost exactly on the axis of the Pelham dome and at the lowest structural level easily accessible to a field trip (Figure 11, Figure 4). It lies below the steep slopes of Poplar Mountain and contains rocks utterly typical of the Poplar Mountain Gneiss, particularly of the thick zone exposed in the Millers River window beneath the Dry Hill Gneiss.

The dominant rock type is a dark gray biotite-quartz-plagioclase-microcline gneiss, commonly with minor muscovite and garnet. Typical of the unit are large and very large irregular megacrysts of white maximum microcline (Laird, 1974), as contrasted with pink maximum microcline megacrysts in the Dry Hill Gneiss. The association of coarse quartz-microcline intergrowths suggests the feldspars may be sheared relics of pegmatites. In addition, the biotites are red-brown as contrasted to green biotites in the Dry Hill Gneiss. The bulk composition is more variable than in the Dry Hill Gneiss (Hodgkins, 1985), and Ashenden (1973) speculated that the unit is mainly

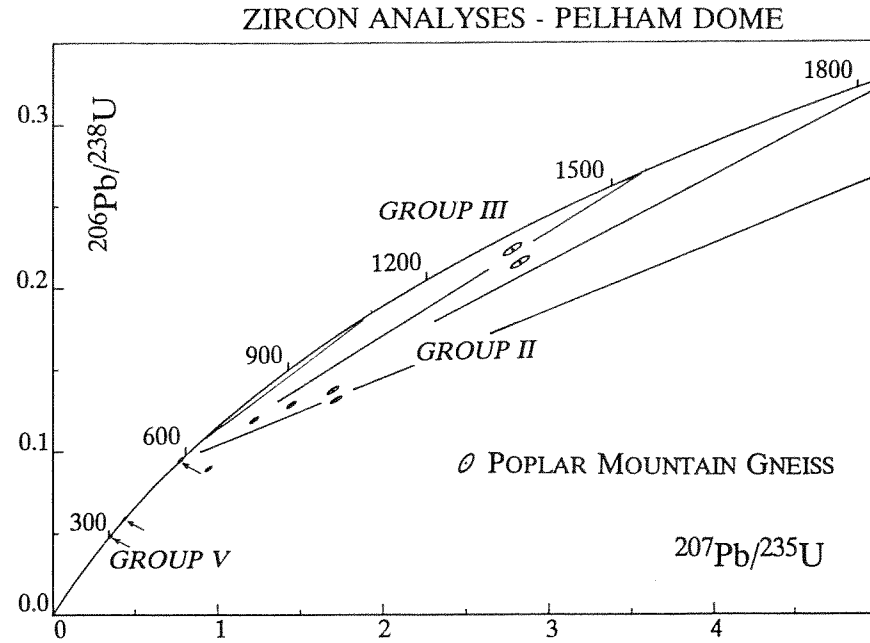

Figure 13. Concordia diagram of ten single-grain zircon analyses from the Poplar Mountain Gneiss at Stop 5. Most grains are very discordant, plotting generally within age groups II and III. Two other analyses (see arrows) are concordant at 370 Ma and 585 Ma, and a third (arrow,Group V) has a $^{206}Pb/^{238}Pb$ age of 302 Ma (Th/U = 0.032) suggesting that it is of metamorphic origin.

of sedimentary derivation, possibly arkoses derived by weathering of the protolith of the Dry Hill Gneiss, and deposited in a more reducing environment. Locally there are thin layers of quartzite and calc-silicate rock that support this sedimentary model. There is also a variety of blocks and boudins of unusual rocks including amphibolite, hornblendite, and calc-silicate rock in various locations at this outcrop.

The structure is dominated by subhorizontal foliation and a strong mineral lineation plunging gently north. On vertical surfaces parallel to the lineation there are superb examples of the types of asymmetric shear fabrics described in detail by Reed (1992), showing a north-over-south shear that is consistent with the abundant late asymmetric folds elsewhere in the dome (Stops 1A, 2, 3, 6, 7, 8). Typically there is an asymmetric pattern of foliation and strain shadows around the microcline megacrysts, but there is little evidence of ductile deformation of the microcline itself. This is consistent with petrologic evidence that the Pennsylvanian deformation and recrystallization took place at 550-600°C.

Most detrital zircons extracted from the Poplar Mountain Gneiss at this outcrop (Figure 13) have yielded very discordant analyses (11-53%), thus minimizing their usefulness for establishing the sedimentary provenance. However, two analyses of single, clear, faceted zircon are concordant within uncertainty (±1% of the $^{207}Pb/^{206}Pb$ age), indicating crystallization ages of ca. 585 and 370 Ma. Another analysis is ca. 12% discordant with a $^{206}Pb/^{238}U$ age of 302 Ma. The youngest grain clearly grew in a Th-depleted metamorphic environment (Th/U = 0.032, U = 80 ppm), thereby suggesting that Late Pennsylvanian metamorphic zircon growth may have produced the strong discordance observed for these detrital grains. Particularly interesting, however, are the two concordant zircon analyses at 585 Ma and 370 Ma. If interpreted as detrital zircon, the youngest of these analyses defines the *maximum sedimentation age* of the protolith, and hence a Late Devonian or younger age for the Poplar Mountain Gneiss. Equally possible, however, is that a sheared Acadian pegmatite, possibly represented in outcrop by the large microcline megacrysts, was the source of the youngest concordant zircon, in which case an Acadian metamorphic history for the gneiss is implied. This scenario is compatible with isotopic evidence gathered elsewhere (Stops 7 and 9), where an Acadian (367-350) sillimanite-orthoclase grade metamorphism is clearly demonstrated in the Mount Mineral Formation.

Continue west past outcrop. Note that foliation dip changes from northeast to north to northwest.
20.0 Outcrop ends.
20.4 Turn right on Old State Road.
20.6 Do U-turn at Erving Mobil Home Park.
20.7 Return to Route 2. Turn left (east) on Route 2. Look both ways for fast traffic.
21.4 Return past Stop 5, noting the crest of the dome exposed in better perspective.
21.5 Bear right into Rest Area. We will pause momentarily here so all can catch up following U-turn.
23.5 Maple Avenue to right and motorbike shop in Village of Farley.

23.6 Take VERY SHARP right turn (south) toward Millers River.
23.7 Cross bridge over Millers River. Parking spot on south bank is alternate lunch stop in case of bad weather.
23.9 Cross grade crossing on main line of Boston and Maine Railroad. Turn right after crossing.
25.2 Outcrops of Poplar Mountain Gneiss to left under powerline.
25.4 Y junction. Bear left toward Wendell.
25.5 Driveway to Allen Walsh farm. Park on right hand side as far off pavement as possible. Walk up driveway for Stop 6 (Figure 11).

STOP 6. INVERTED SEQUENCE OF POPLAR MOUNTAIN GNEISS, POPLAR MOUNTAIN QUARTZITE, DRY HILL GNEISS BIOTITE MEMBER WITH PENNSYLVANIAN ASYMMETRIC FOLDS (1 HOUR 15 MINUTES) The purpose of this stop is to observe the inverted stratigraphic sequence beneath the isoclinal fold nappe of Dry Hill Gneiss. These rocks were first studied in detail by Ashenden (1973) and was part of the area of Onasch's (1973) detailed fold study, all in inverted rocks. We will begin in the Poplar Mountain Gneiss and traverse upward through the Poplar Mountain Quartzite Member with its typical structurally lower zone of amphibolite, and into the Dry Hill Gneiss Biotite Member that forms the roof of the Northfield Mountain underground powerhouse. En route through the Quartzite Member, we will observe a variety of asymmetric folds that, taken together, document north-over-south shear parallel to the north-south lineation.

Walk up driveway to house of Allen Walsh. Step across electric fence and brook into field under powerline. Bear left toward west edge of field and left of ledges under power poles, and walk past first outcrops to trees north of powerline. Cross fence into trees, then walk west to get around juniper thicket and back out on powerline for about 300 feet to first west-facing ledge of typical Poplar Mountain Gneiss.

Return eastward to ledge under power poles to observe a variety of amphibolites structurally low in the Poplar Mountain Quartzite. One of these is a gneiss composed of hornblende, scapolite, andesine, quartz and sphene; another is a hornblendite with brown euhedral sphene crystals up to 2 cm long. Follow northward around lower side of west-facing ledges of spectacularly folded quartzites, then up slope to north end of field. In one of these quartzite outcrops, asymmetric folds with opposite shear sense can be found with a separation angle of less than 20° (Onasch, 1973, Figures 23 and 30) showing consistent north-over-south longitudinal shear. We will also pass the locality where a micaceous quartzite yielded coarse, detrital zircons with ages of 1540, 1206 and 1028 Ma (Figure 6C). On this limited basis, the Poplar Mountain Quartzite appears to lack the older detrital components in the Pelham and Dunlop Brook quartzites, but like these and unlike the upper Mount Mineral Quartzite, it also appears to lack zircons from the adjacent Dry Hill Gneiss. Metamorphic monazite, dated at 298 ±2 Ma, is also present in this outcrop. In the same quartzite layer 1.2 miles to the southwest, a highly sheared and boudinaged pegmatite has yielded igneous zircons dated at 299 ±2 Ma.

Follow logging road from field north about 200 feet, then walk northeast across wet groove to base of steep slope with cliffs. These begin with the highest part of the Quartzite Member (see fold, Onasch, 1973, Figure 7), succeeded by the Dry Hill Biotite Member, locally with some hornblende. From Biotite Member drop back to logging road and follow back to Walsh house and cars.

Continue on paved road toward Wendell.
27.8 Stop sign, Turn right (south) toward center of Wendell.
28.5 Center of Wendell, the second of four Pelham dome towns to be visited on this trip. Go straight.
29.1 Wendell Country Store.
31.8 Shutesbury Town Line at Camp Anderson. Turn sharp right at four-way junction.
32.4 Lake Wyola beach.
32.6 Junction near lake outlet. Turn sharp left on Locks Pond Road toward Shutesbury A. C.
36.5 T junction at Shutesbury Post Office and Town Hall. Turn left (east) toward Route 202. Pass by Moses N. Spear (Frank's great great uncle) Memorial Library on left. Begin steep descent.
36.9 Small exposure of Mount Mineral Formation on right.
37.5 Pelham-Shutesbury syncline (Michener, 1983) exposed in ditch on left.
37.7 T junction and stop sign. Turn right (south) on Route 202.
38.3 Road cut in Fourmile Gneiss east of Pelham-Shutesbury syncline.
38.4 Road cut in Partridge schist of Pelham-Shutesbury syncline on left.
38.7 Cuts in Fourmile Gneiss between Pelham-Shutesbury syncline and Mount Mineral Formation.

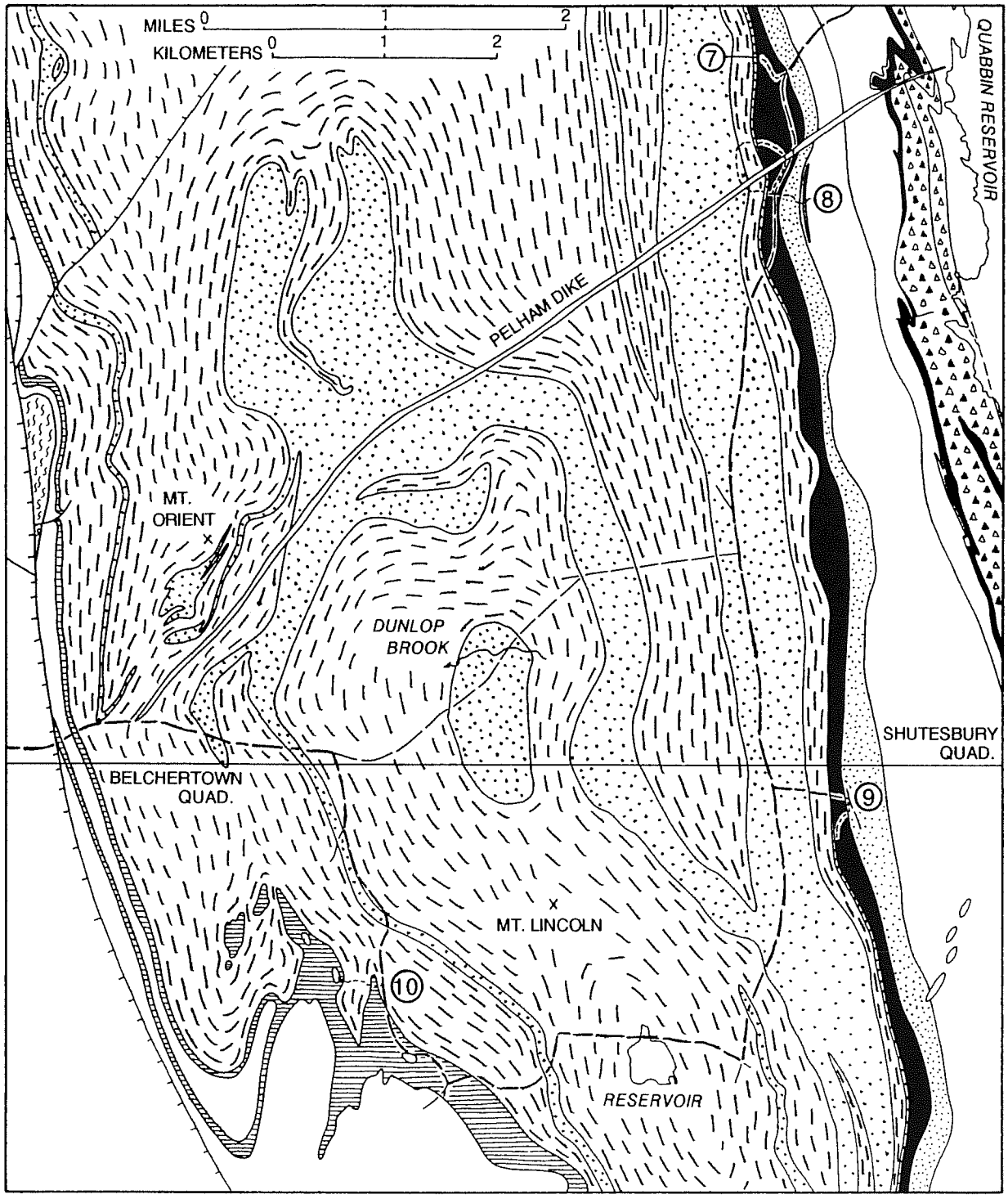

Figure 14. Detailed map of the south-central part of the Pelham dome showing the locations of Stops 7 to 10. For Key to patterns see Figure 4.

39.2 Descend long straight incline and pull off into paved ditch on right just before steep gravel driveway. LEAVE FLASHERS ON. Walk up driveway about 30 feet, turn sharp left (west) on old road and descend to Atherton Brook. Cross brook on stepping stones and walk about 200 yards northwest to outcrops.

STOP 7. SCHIST OF MOUNT MINERAL FORMATION WITH RELICT ACADIAN GRANULITE FACIES MINERALS OVERPRINTED BY PENNSYLVANIAN KYANITE-MUSCOVITE ASSEMBLAGE (20 MINUTES) Cross Atherton Brook and walk northwest about 200 yards (Figure 14) to small outcrops about 100 feet southwest of stream bank (Ashwal Locality 160). The western outcrop consists of coarse mylonitic sillimanite-orthoclase-garnet-biotite schist with north-plunging Pennsylvanian dome-stage anticlinal fold superimposed on an older west-over-east shear fabric. Locally there are sillimanite lineations parallel to the fold and to the earlier shear fabric. This was the sillimanite locality deep within the kyanite zone discovered by Lewis Ashwal in fall 1972 (Robinson, Tracy, and Ashwal, 1975) that set off a whole line of new research. The garnet with extreme resorption zoning described by Roll (1987, cover illustration, see also Robinson, 1991) came from this outcrop, as did samples of orthoclase of high structure state. The relict assemblage in this rock includes sillimanite, orthoclase, biotite, garnet (pyrope 35, spessartine 1.1) and rutile, indicative of a temperature of about 700°C and pressure of 6.8 kbar. The garnet rims fall off to a pyrope content of 10-11%, and are commonly surrounded by a matrix of muscovite, biotite, and kyanite indicative of conditions of re-equilibration under more hydrous conditions at about 580°C and 6 kbar. The orthoclase is commonly rimmed by a fine-grained intergrowth of muscovite and sodic plagioclase against the aluminous matrix. A U-Pb age of 367 ±2 Ma was obtained from metamorphic monazite in this rock (Figure 7A).

The mylonitic schists at this outcrop successfully resisted the Pennsylvanian recrystallization due to local scarcity of water, and they contain an older linear fabric trending roughly E-W which involves oriented sillimanite, highly elongate quartz ribbons, and tails around orthoclase and garnet, including evidence that the orthoclase was itself undergoing strong grain-size reduction with limited recrystallization. One of the samples from the granular top surface of the outcrop is a sillmanite-orthoclase-garnet-rutile mylonite with subordinate biotite, in which quartz, plagioclase and orthoclase all underwent severe grain-size reduction with only minimal recovery. These features are all evidence of an earlier and much higher-temperature phase of shearing, probably soon after the peak of Acadian granulite facies metamorphism. Preliminary thin section observations indicate a top-side east shear sense in this fabric.

The eastern outcrop consists of mica schist containing kyanite, staurolite, muscovite, biotite, and euhedral garnets, that is believed to be chemically equivalent to the western outcrop, but has undergone hydration and complete chemical reconstitution during Pennsylvanian metamorphism and deformation. The garnets have the characteristics of ones that have undergone prograde growth zoning. In this outcrop there is strong evidence of north-over-south shearing parallel to a subhorizontal Pennsylvanian mineral lineation.

Continue south on Route 202.

39.6 Right hand bend with outcrops of Pelham Jurassic dike on right.

39.7 Road cut on right in amphibolite and gneiss of Mount Mineral Formation (see Stop 8).

40.0 Medium-sized parking space on left bounded by steel rails. WATCH TRAFFIC AND FIT VEHICLES IN EFFICIENTLY. Follow walking route for Stop 8.

STOP 8. STRATIGRAPHIC SEQUENCE FROM PELHAM QUARTZITE, THROUGH DRY HILL GNEISS, MOUNT MINERAL FORMATION, AND FOURMILE GNEISS ON EAST-CENTRAL SIDE OF THE DOME (70 MINUTES) From parking place walk north along Route 202 to culvert over Briggs Brook. Cross to west side of highway and follow flagged trail up brook to area of low outcrops of Pelham Quartzite, about 600 feet thick in this vicinity (Figure 14).

Eighteen detrital zircons from this layer of Pelham Quartzite collected on Route 202, 1.3 miles to the south, and another twelve detrital zircons from a mineralogically similar quartzite at Dunlop Brook, were analysed as individual crystals to determine if the range of inheritance ages in composite zircons from the Dry Hill Gneiss (Stop 1, Figure 6A) is consistent with the age of zircon detritus in the Pelham Quartzites. Seventeen of the 18 analyses from the Pelham Quartzite on Route 202 are less than 6% discordant, indicating that secondary Pb-loss or new zircon growth has been nearly eliminated, and that the $^{207}Pb/^{206}Pb$ ages are reliable ages of their source rocks, not minimum ages. The twelve analyses of detrital zircon from the quartzite at Dunlop Brook are, in general, more discordant than those from the Route 202 locality, but three important points can be made (Figure 6B).

1) Three detrital grain analyses from Dunlop Brook yielded $^{207}Pb/^{206}Pb$ ages between 2616 and 2679 Ma, thus confirming the estimate of Group I inheritance ages defined by the composite grain work (Figure 6A). Also, detrital grain ages closely approximating age Groups II and III were measured in both widely spaced quartzite samples, implying that all age groups (I, II, III) are reliable indicators of sedimentary provenance,
2) The youngest analysis in the detrital grain population has a $^{207}Pb/^{206}Pb$ age of 933 Ma, thereby establishing a *maximum* depositional age of the quartzite. This grain has a low Th/U ratio (0.02) and high U abundance (1791 ppm U) suggesting it was derived from a pegmatite or possibly high-grade metamorphic terrane of late Grenvillian age. In contrast to the Mount Mineral and Poplar Mountain Quartzites (Figure 6C), the Pelham and Dunlop Brook quartzites do not commonly contain monazite, but instead sphene and actinolite. All the quartzites and intercalated gneisses and schists were strongly recrystallized and metamorphosed in the Late Pennsylvanian, but in contrast to the Mount Mineral Formation, no evidence points to an Acadian metamorphic history for the Pelham and Dunlop Brook quartzites.
3) Late Proterozoic, ca. 700-600 Ma, detritus is apparently lacking in the Pelham, Dunlop Brook, and Poplar Mountain Quartzites, implying that the Dry Hill Gneiss was not a major source of detritus nor were any other rocks of Late Proterozoic (Avalonian) affinity. The good agreement between the age groups defined by the detrital-grain and composite-grain studies suggests strongly that the source of the inheritance in the Dry Hill Gneiss (Stop 1) is detritus at the site of eruption, whose sources are dominantly Early and Middle Proterozoic rocks. Although the provenance of these quartzites is in doubt, both South American and Laurentian sources would seem to be indicated.

The top contact of the Pelham Quartzite against Dry Hill Gneiss follows the brook for some distance to a swampy area, east of which are cliffs of Dry Hill Gneiss with conspicuously coarse hornblende and small protruding microcline megacrysts suggestive of a porphyry. Ascend eastward to largest cliff where preliminary U-Pb zircon analyses suggest an igneous crystallization age of ca. 611 Ma.

Continue up cliffs to lower outcrops at top consisting of actinolite quartzite of basal member of the Mt. Mineral Formation. Petrographically this quartzite is identical to the most common type of Pelham Quartzite and other quartzites associated with Proterozoic rocks. It was thus natural to compare this with the Poplar Mountain Quartzite and to consider the rest of the Mount Mineral Formation as a facies of the upper layer of Poplar Mountain Gneiss. Robinson is still inclined to think that this quartzite is Proterozoic. Tucker is presently working on detrital zircons from this outcrop to throw light on this question.

Walk about 200 feet east across crest of hill to low outcrops of highly sheared kyanite schist and pegmatite. Although the matrix of the schist is a completely reconstituted kyanite-biotite-muscovite rock, the rock contains sheared relict megacrysts of orthoclase, sillimanite inclusions in garnet, and rutile, all features of the formerly granulite-facies Mount Mineral schists. A large sample of sheared orthoclase-rich pegmatite with E-W trending lineation from this outcrop is under study.

Walk down steep southeast slope of hill with limited outcrop to west side of Route 202 and then walk south to a series of small road cuts. These are in amphibolites and plagioclase gneisses within the middle member of the Mount Mineral Formation (msa) as mapped by Robinson et al., 1973. They contain superb exposures of late asymmetric folds in foliation, many with dome-stage lineation that has undergone great-circle rotation (Unpublished data of Robinson, 1967, reproduced by Michener, 1983, Figure 14). The fold axes generally plunge northeast, whereas the lineation trends more northerly. The intersections of the great circles of rotation of the lineation with the fold axial planes yields a transport direction of about N20E, consistent with other features of the Pennsylvanian deformation, and fold asymmetry indicates east-side-south transport. One layer bears striking resemblance to Monson Gneiss or gray Fourmile Gneiss in the local Ordovician basement and it has yielded a single concordant U-Pb zircon analysis of 456 Ma. Thus, the lower part of the Mount Mineral Formation, above the basal quartzite, lithically resembles Partridge Formation mica schist and amphibolites, and also Ordovician "basement" rock types, and may be a sliced-up amalgam of both.

Beyond the road cut, cross to east side of Route 202 and descend east down logging road behind locked gate. Cross low ground eastward to right angle bend of logging road, and immediately ascend west-facing bluff on logging track cut through laurels. At top of bluff is typical outcrop of the basal part of the yellow-weathering member of the Fourmile Gneiss, typically a fine- to medium-grained biotite-muscovite-feldspar gneiss with minor tiny garnets and magnetite. The origin of the yellow- and red-weathering is uncertain, but is being investigated by M. S. student, Jon Bull at the University of Massachusetts. In this one respect the unit resembles the Upper Member of the Ammonoosuc Volcanics.

Walk south about 200 feet to outcrops of muscovite quartzite on face of the bluff and slightly west of outcrops of yellow-weathering gneiss. This is the typical quartzite of the upper member of the Mount Mineral Formation as originally mapped (Robinson et al., 1973) and is the outcrop collected by Tucker that is responsible for the recent dramatic developments. Eighteen detrital zircons from this quartzite display a range of near-concordant U-Pb analyses between approximately 1362 Ma and 439 Ma (Figure 6C). Three grains, with habits closely approximating those of the Monson Gneiss (Tucker and Robinson, 1990), yielded concordant analyses indicating ages of 459, 441, and 439 Ma. These ages agree well with the known period of magmatism of the local Bronson Hill magmatic arc, and they demonstrate conclusively that his quartzite can be no older than Late Ordovician and may well be the early Silurian Clough Quartzite. Other detrital grains have ages that approximate established ages of Grenvillian and post-Grenvillian rocks in the Green Mountain and Berkshire massifs (Ratcliffe at al., 1991; Karabinos and Aleinikoff, 1990), as well as the local Dry Hill Gneiss, indicating a completely different provenance from the Pelham and Dunlop Brook quartzites, though a similar Middle Proterozoic component as the Poplar Mountain Quartzite. Replicate analyses of metamorphic monazite from this upper Mount Mineral quartzite have yielded a concordant age of 298 ±2 Ma, in agreement with Late Pennsylvanian metamorphic ages in many nearby rocks.

From the quartzite walk northeast past several outcrops of Fourmile Gneiss and across another poorly exposed belt of quartzite "within" the Fourmile to low knobs with blocks and small outcrops of garnet amphibolite and coarse gedrite gneiss. This rock, like the yellow-weathering gneiss also bears superficial resemblance to parts of the Ammonoosuc Volcanics and lead to speculation about the nature of the contact at the base of the Fourmile against the quartzite. Is it a thrust of Ordovician basement over Silurian cover, or is it an unconformity with Ordovician basement overfolded onto Silurian cover?

If time permits, we will walk south into the gorge of Briggs Brook where the "Clough Quartzite" and a second quartzite "within" the Fourmile are well exposed. The second quartzite is probably the Clough repeated by complex folding. However, note on the southwest flank of the dome there is an extensive zone of quartzite apparently in the middle of the Fourmile Gneiss.

Return on logging road to Route 202 and walk south on east side back to vehicles.

Continue south on Route 202.

40.1 Pelham Town Line.

40.7 Road cut on right in Pelham Quartzite. Turn out on left with view of Quabbin Reservoir and Mt. Wachusett (2006'). Like the Poplar Mountain Quartzite (Stops 1A and 6) and the basal quartzite of the Mt. Mineral Formation (Stop 8), the Pelham Quartzite commonly contains actinolite indicative of a former dolomite cement. See Stop 8 for discussion of detrital zircons from this outcrop.

40.9 Flashing yellow light at turn off for Amherst Road. Stay on Route 202.

41.2 Another cut in Pelham Quartzite on right.

42.2 Quabbin Reservoir, Gate 10 on left. Turn left (east) through locked gate onto gravel road.

42.7 T junction at elevation 1069'. Park in available space and walk southwest to outcrops in clearings.

STOP 9. UPPER MUSCOVITE QUARTZITE (CLOUGH QUARTZITE?), SILLIMANITE PEGMATITE AND SILLIMANITE-ORTHOCLASE SCHIST OF MOUNT MINERAL FORMATION, SOUTHEAST SIDE OF DOME (30 MINUTES) Walk southwest from junction to first west-facing cliff of muscovite-garnet quartzite (Figure 14). This is the same layer as at Stop 8 that yielded detrital zircons indicating the rock is probably Clough Quartzite. Foliation dips gently east and lineation trends roughly north-south and plunges gently north. There are several indicators of north-over-south shearing.

Most important at this outcrop is the detailed petrologic information obtained in sample M21 by Roll (1987). This shows an Fe-rich garnet which has undergone extensive retrograde resorption zoning (Roll, 1987, Figure 23; Robinson, 1991, Figure 35) with a core composition of Pyrope 4.7, Almandine 85, Spessartine 2.7, Grossular 7.6 to a rim composition of Pyrope 4.4, Almandine 73.5, Spessartine 13.5, Grossular 8.6. Roll suggested that this zoning is a result of a retrograde continuous hydration reaction, GAR + KSP + H_2O = BIO + MUS + QZ. As in other samples from the Mount Mineral Formation, some biotites included in garnet contain higher Ti content than matrix biotite (Roll, Figure 11; Robinson, 1991, Figure 36), locally up to 0.209 Ti/11 oxygens, indicative of a high-temperature heritage. The important conclusion is that the upper quartzite member of the Mount Mineral Formation at this outcrop enjoyed the same Acadian granulite-facies metamorphism as middle member schists, although the relict features are less well preserved.

Continue southwest past outcrops of amphibolite to a low knob of coarse mylonitic schist and sillimanite pegmatite. These lie 10-20 meters above the contact with Dry Hill Gneiss. This is locality M22 (Roll, 1987), which yielded 367 ±2 Ma monazite ages in both schist and pegmatite and a 350 ±2 Ma zircon age in sillimanite pegmatite (Figure 7A). It represents one of the areas of best preserved original granulite facies metamorphic features, and also of early E-W trending lineation and shear fabrics. Asymmetric tails on orthoclase and garnet porphyroclasts in slabs from this outcrop cut parallel to this lineation show east-side-down shear sense. Such a sense would be consistent with westward underthrusting of the Proterozoic basement beneath the Mount Mineral Formation after 350 Ma, but also with a set of east-directed fold and thrust nappes. Roll (1987) as well as Robinson and Peterson have noted that in these outcrops orthoclase forms recrystallized tails on porphyroclasts, a clear sign that this deformation took place under higher-temperature conditions than the later north-over-south Pennsylvanian shearing, possibly just after the peak of granulite-facies Acadian metamorphism. Return by same route to vehicles.

Return to Route 202, again through locked gate.
43.2 Turn left (south) on Route 202.
44.3 Road cut on left in Dry Hill Gneiss.
44.7 Cross road at Quabbin Gate 8. Turn sharp right (west) off of Route 202 onto Parkardville Road.
45.0 Belchertown Town Line.
45.2 View of Knightville Reservoir on left. The southernmost "hump" of the Pelham dome lies approximately in the southwestern corner of the reservoir.
45.8 Turn left at three-way junction onto Gulf Road.
46.6 Junction by dam. Turn sharp right on North Gulf Road.
47.0 White house with red shutters on left overlies a lens of harzburgite with anthophyllite veins in Mt Mineral Formation.
47.3 Former Shaw's Turkey Farm. Large garage on right, outcrop in garden on left.

STOP 10. MOUNT MINERAL FORMATION AND DRY HILL GNEISS IN LATE ASYMMETRIC FOLDS AND HARZBURGITE WITH ANTHOPHYLLITE VEINS AT THE PELHAM ASBESTOS MINE (45 MINUTES) In the southern part of the Pelham dome, the late asymmetric folds reach giant proportions with amplitudes in excess of 5 km (Figures 5 and 14). They trend across the dome axis, plunging southeast to east on the east limb and northwest to north on the west limb. The garden outcrop is at the hinge of one such fold (Figure 14) in which mylonitic schist of the Mount Mineral Formation in an "anticline" plunges gently north beneath "structurally lower" Dry Hill Gneiss. The fold is a broad fold in mylonitic foliation superimposed on a variety of earlier features. To the northwest about 1.2 km (Figure 14) there is an exposed hinge of a very similar northwest- to north-plunging anticline. West of this hinge are three isolated patches of Mt. Mineral Formation. All three are erosional windows in a stream gully on the steep west face of the Pelham Hills in which Mount Mineral Formation is exposed from beneath Dry Hill Gneiss on the overturned short limb of this large late asymmetric fold.

The Mount Mineral schist in the garden is a typical mylonitic schist with porphyroclasts of garnet and orthoclase in a matrix of biotite, kyanite, and minor muscovite (Roll, 1987) Dark rims of biotite around the garnets are obvious. Two thin sections gave the following mineral proportions: quartz 50-60%; plagioclase An28-31 10-20%; relict orthoclase 0-1%; biotite 5-11%; garnet 8-20%; kyanite 5-8%, sillimanite inclusions in garnet 1%, and minor matrix muscovite, rutile, ilmenite, pyrrhotite and chlorite. One of the thin sections yielded a 1x3 mm garnet porphyroclast studied in detail by Roll (1987) and illustrated by Robinson (1991). This has a preserved relict core with Pyrope 34.7, Almandine 77.5, Spessartine 1.4; Grossular 0.6 grading rapidly to a rim with composition Pyrope 14, Almandine 63.4, Spessartine 3.4., Grossular 5.0. The preservation of the tiny pyrope-rich relict only 0.4 mm from the matrix strongly suggests that this garnet is a fragment from the core of a much larger garnet that was only tectonically exposed to the matrix late in the shearing process.

Pass through fence gate by house and follow farm road west across the field where there are outcrops of hornblende-bearing Dry Hill Gneiss in the "synclinal" belt west of the anticline. Follow road into woods where it bears left, to junction with obscure trail that turns sharp right (northwest) to dumps and the pit of the "Pelham Asbestos Mine" (Emerson, 1898) within one of four lenses of metamorphosed harzburgite within the Mt. Mineral Formation. The mine itself is a pit about 60 x 60 feet on the west side of which the contact between altered ultramafic rock and pegmatites and gneisses can be seen. Where least altered, the ultramafic rock consists of about equal proportions of 1-3 cm porphyroblasts of olivine and orthopyroxene in a retrograde matrix of talc, anthophyllite, chlorite and magnetite. The feature of former economic interest was thin to thick cross-cutting veins of cross-fibre anthophyllite asbestos with a fibre length locally reaching 30 cm. This was specifically investigated

by D. R. Veblen, with the possibility of finding jimthompsonite or other multiple-chain silicates. However, the asbestos proves to be aggregates of fine prisms of anthophyllite sharing a common *c*-crystallographic axis, but completely randomly oriented with respect to *a* and *b* axes, and with the intersices between the prisms filled by talc, serpentine, and chlorite (Veblen, 1980). This rock is the only one of true mantle composition that has been observed in the Bronson Hill anticlinorium in Massachusetts, and its location within the already intriguing Mount Mineral Formation lends further fuel to speculation. The coarse olivine and orthopyroxene are believed to have grown during Acadian granulite-facies metamorphism, and the anthophyllite veins during the Pennsylvanian shearing and recrystallization.

Walk south from the south edge of the pit about 100 feet to a steep east-facing outcrop in the woods. This is typical Mount Mineral mylonitic schist, rather more reconstituted than most. It contains sheared relics of orthoclase and tiny sillimanite needles inside garnets, but only kyanite in the matrix. Observations on sawed slabs show it contains the north-over-south Pennsylvanian shearing, but no clear evidence of earlier E-W trending features.

Continue north on Gulf Road.
47.7 Stop sign. Go straight (north) on Enfield Road.
48.0 Outcrop in field on left of thin Pelham Quartzite.
48.2 Junction of Butter Hill Road. Go straight.
48.7 T junction with Amherst Road. Turn left (west) toward Amherst (5 minute drive). END OF FIELD TRIP!

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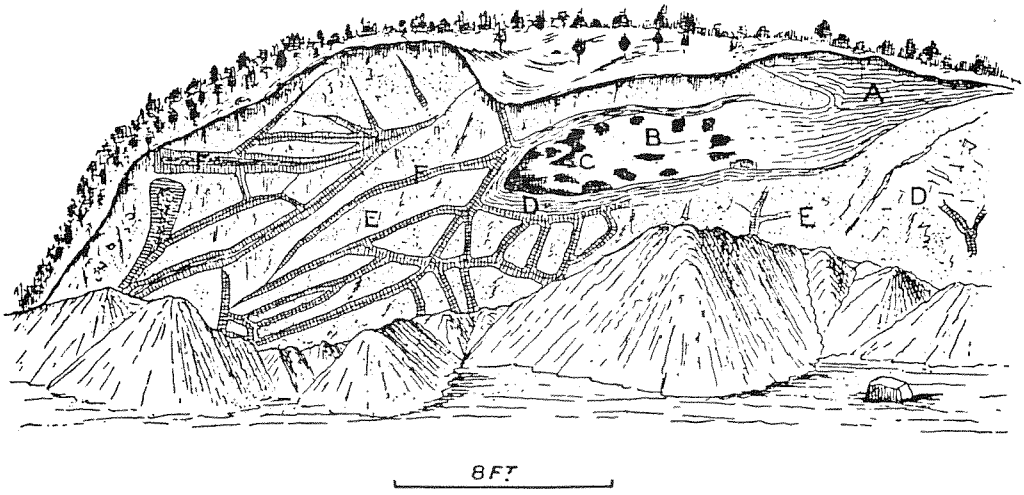

FIG. 3.—Southwest wall of Pelham asbestos quarry in 1890. A, Pelham gneiss; B, anorthite rock; C, black tourmaline masses; D, biotite layer; D', vermiculite layer (vermiculite and steatite from biotite and actinolite); E, saxonite; F, anthophyllite derived from saxonite.

MOHAWK TRAIL CROSS SECTION OF THE MESOZOIC DEERFIELD BASIN: STRUCTURE, STRATIGRAPHY, AND SEDIMENTOLOGY

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INTRODUCTION

An ancient, east-west trail of the Mohawk Indians led from the Boston Region across the Berkshires and into the Mohawk Valley of New York. This trail is now followed in part by Route 2 that crosses the Deerfield basin just north of Greenfield,Turners Falls, and Millers Falls. In this same area, the Connecticut River takes a large east-west bend, eroding a deep valley with excellent exposures across much of the basin. This trip makes use of this zone of good natural exposures and of the roadcuts along and near Route 2 to focus on the Mesozoic history of the basin.

The Deerfield basin, part of the Connecticut Valley rift system, is an east-tilted half graben situated between the Berkshire Mountains on the west and the Bronson Hill line of mantled gneiss domes on the east (Figures 1, 2, 3). The objectives of this trip include presentation of a cross section of the basin from the Berkshire foothills on the west to the Paleozoic "basement" just east of the eastern border fault. Our intent is to show the stratigraphic sequence and depositional environments of basin-fill units of Late Triassic to Early Jurassic age from oldest to youngest across the basin from west to east. Basin stratigraphic units, from oldest to youngest, are: Sugarloaf Arkose (1700 m); Deerfield Basalt (55 m) and Turners Falls Formation and Mount Toby Conglomerate (+/- 2000 m), the Mount Toby being a lateral facies equivalent of the Turners Falls Formation on the northern and eastern sides of the basin.The Triassic-Jurassic boundary is usually assumed to be in the upper part of the Sugarloaf Arkose. Another objective is to demonstrate a two-phase stress history starting with NW-SE extension that produced crustal stretching, rift sedimentation, and subsequent overall tilting of the basin and activation of normal faults. The second phase was marked by N-S to NE-SW compression that produced strike-slip faulting, possibly as a change to the drift stage and ridge pushing due to opening of the Labrador Sea and Atlantic Ocean.

Extensive stops and discussions will be held on the floor of the Connecticut River just below the Turners Falls Dam. At times of highest water these outcrops are inaccessable, but a reasonable trip can be run at moderate stage water levels. In most years, river levels are so low from June through October that a large pavement of exposures along the river floor is available for field trips. We reserve the right to change the schedule of stops in accord with the river stage at the time of the trip.

Many of the structural details of some of the stops on this trip have been given in a former NEIGC guidebook (Wise, 1988) and a NAGT guidebook (Wise and Belt, 1991) and will not be repeated here. The present guidebook includes the basic maps and figures needed for the stops plus a number of compilations and figures from six theses and two undergraduate reports done over the years at the University of Massachusetts (Goldstein, 1975; Handy, 1976; O'Toole, 1981; Meriney, 1988; Loundsbury, 1990; Stopen, 1988; Taylor, 1991; Williams, 1979).

CHRONOLOGIC SEQUENCE OF EVENTS

Late Precambrian to Ordovician	....	Deposition of protolith of Fourmile Gneiss at French King Bridge, metamorphism, erosion.
Ordovician	....	Deposition of Partridge Formation black shales of Bronson Hill area, French King Bridge exposures (Stop 8).
Ordovician	....	Taconian orogeny, extent still debated for Bronson Hill area.
Devonian	....	Marine deposits of Connecticut Valley-Gaspe Synclinorium to produce protolith of eastern Berkshires (Stop 1) and much of the basin floor for the rift (exposures a few km to the north).
Devonian	...	Acadian orogeny: most of the metaporphism of the Berkshires and Bronson Hill mantled gneiss domes.
Late Paleozoic	....	Alleghenian orogeny: some deformation and metamorphism; extent under debate for this region; possible early motions on eastern border fault.
Late Paleozoic to mid-Triassic	....	Regional erosion for basement floor beneath Deerfield basin.
Late Triassic to earliest Jurassic	....	Deerfield basin initiated by sag and/or local faults as NW-SE crustal stretching begins. Sugarloaf Arkose deposited (Stops 2 and 3), mostly of fluvial origin with minor alluvial-fan, lacustrine, and playa strata towards the top. Precipitation of albite cement begins and continues as groundwater transfers Na from sodic plagiocase disintegration in the source area highlands.
Early Jurassic	.....	Deerfield Basalt lava flows into a lake on the rift basin floor (Stops 3, 4, 6A, B).
Early Jurassic	.....	Playa deposits of lower Turners Falls Formation, with aluvial fans to N and E (Stop 6C).
Early Jurassic	.....	Lacustrine beds and fluvial sandstone of Turners Falls Formation , just below the Turners Falls dam (Stop 6D).
Early Jurassic	.....	Barton Cove lacustrine mudstones deposited (Stop 5). Some gravity sliding and slumping. Mount Pisgah alluvial-fan gravels (Stop 7) grade from N and NE into lacustrine gray muds.
Early Jurassic	.....	Mount Toby Conglomerate deposited mostly from the NE and E. Basin edge steps eastward to border fault in Mount Toby area.
Early to Mid-Jurassic	.....	Overpressure of water in lake beds causes fibrous calcite to form sill-like veins that slid by gravity toward the SE (Stop 6D).
Mid-Jurassic (?)	.....	Major tilting of the basin contents, integration of lesser faults into main border fault. Normal faults at dam. Extensional veins in lacustrine beds; maturation and main migration of hydrocarbons (Stops 3 and 6D)
Mid to Late Jurassic	.....	Change in the major stress field from NW-SE extension to compression first N-S, then shifting to NE-SW. Strike-slip faulting; thrusting; spaced cleavages; compressional (?) kink folds at Barton Cove (Stops 4, 5, 6A).
Late Pleistocene to Recent	.....	Glaciation, glacial lake Hitchcock 15,600-12,400 B. P. Subsequent river downcutting though the Hitchcock deltas of sand and gravel.

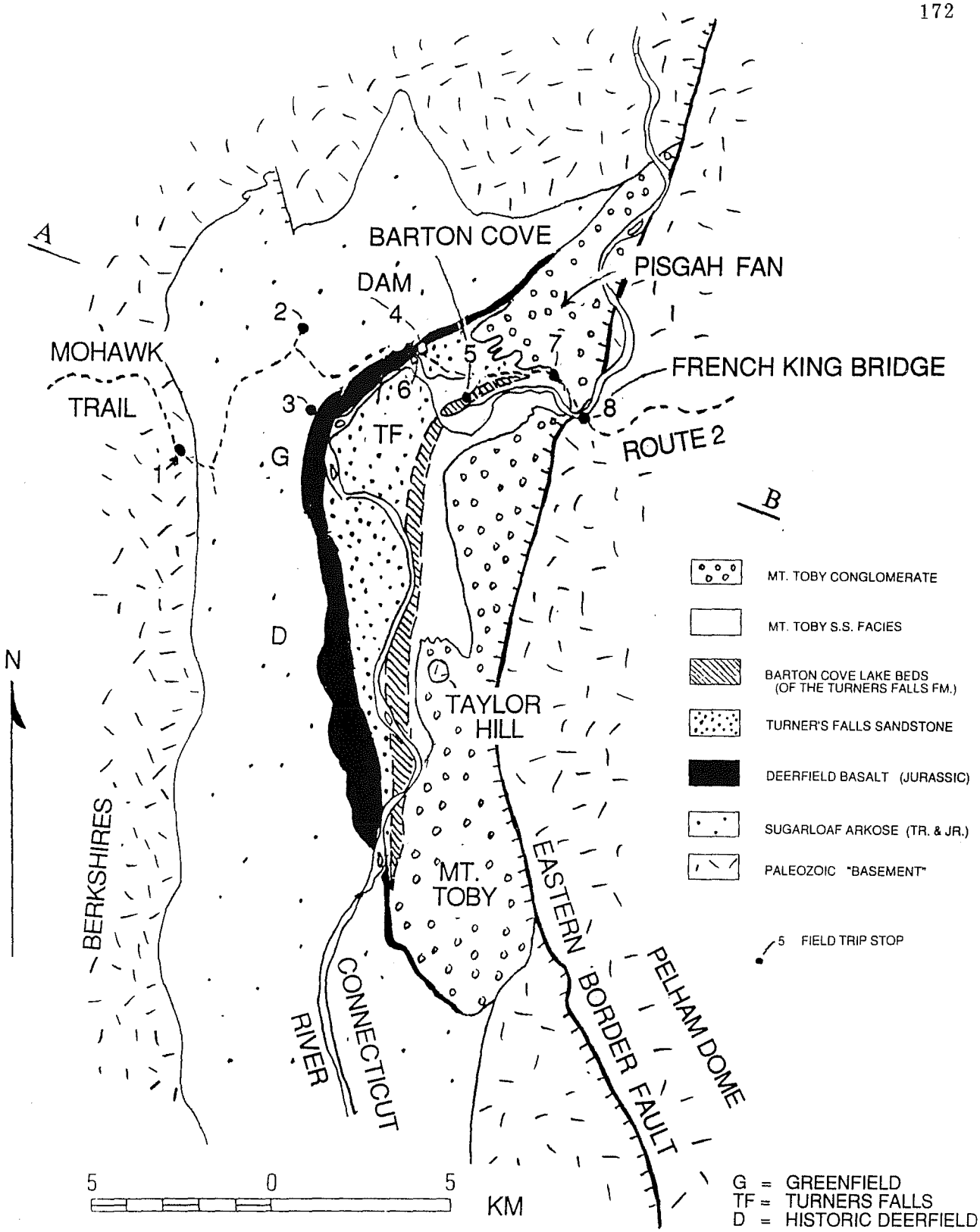


Figure 1. Index map to the Mesozoic Deerfield basin and route of the field trip.

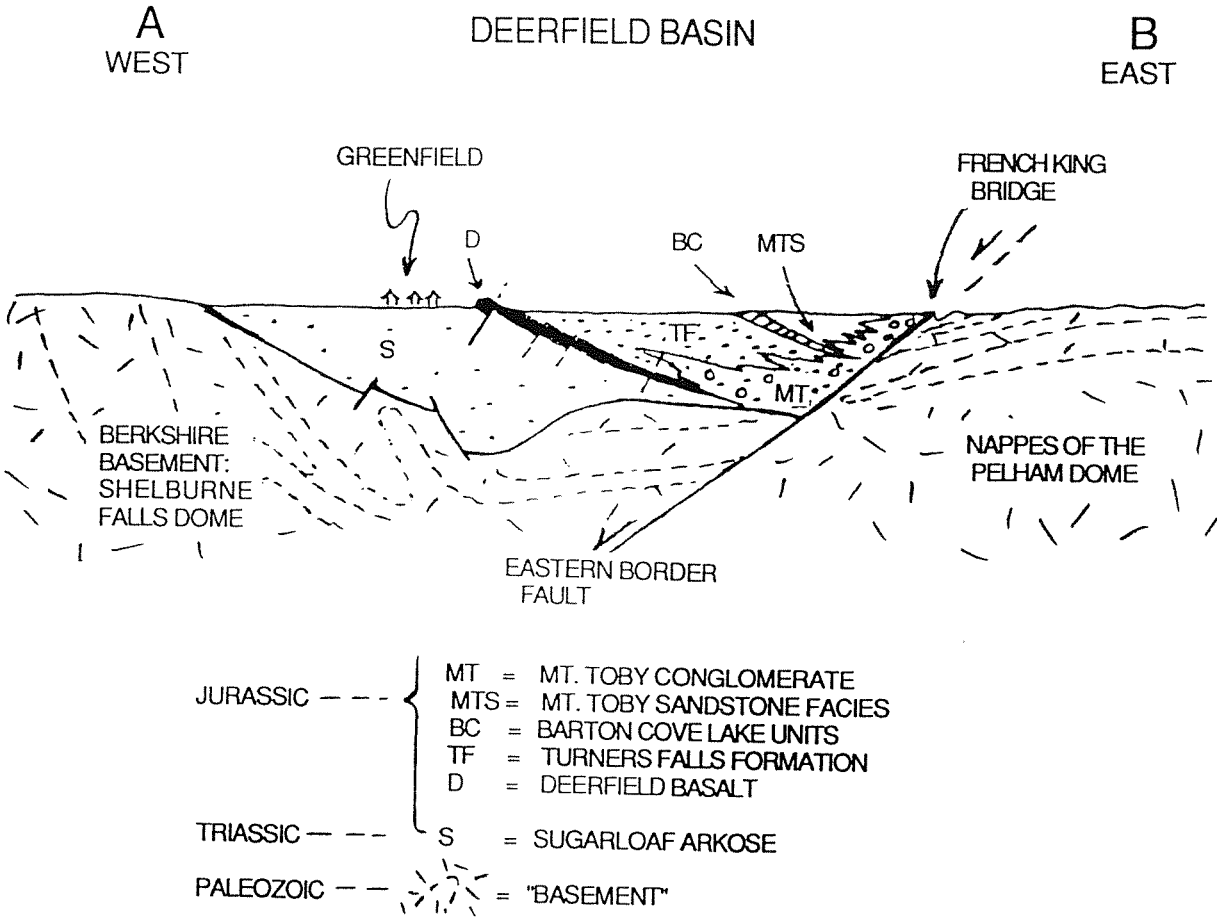


Figure 2. Geologic cross section A-B across the Deerfield basin.

FIELD TRIP STOPS

Stop 1	Route 2 overlook of Deerfield basin from the Berkshire foothills.
Stop 2	Fluvial redbeds of the Sugarloaf Arkose on Country Club Road (Swamp Road) in Greenfield.
Stop 3	North Greenfield Super Stop and Shop cut in the top of the Sugarloaf Arkose.
Stop 4	Deerfield Basalt roadcut on Route 2 across from the overlook of Turners Falls dam.
Stop 5	Deformed lake beds in the Turners Falls Formation at Barton Cove. Pack your own food for a picnic-style lunch at the Barton Cove Nature Area maintained by Northeast Utilites. Toilet facilities are available.
Stop 6 A,B,C,D	Turners Falls Formation and faulted Deerfield Basalt along the floor of Connecticut River below the Turners Falls dam.
Stop 7	Mount Pisgah alluvial-fan conglomerate in the Mount Toby Conglomerate, along Route 2 east of and on strike with Barton Cove lake beds (rolling stop).
Stop 8	Border fault and eastern "basement" beneath east end of French King Bridge.

ROAD LOG

The trip will assemble at 8:00 A.M. on Saturday, October 10, 1992, in the parking lot at the north end of the University of Massachusetts football stadium in Amherst. It will proceed to Interstate 91, then go north to Exit 26 (Routes 2 and 2A) in Greenfield, turning west on Route 2 (a 270 degree turn on the cloverleaf). The mileage log begins at the cloverleaf.

Miles

0.0 Cloverleaf of Interstate 91 and Route 2 in Greenfield, MA., head west on Route 2.

0.5 Topographic break marks the west edge of the Deerfield basin, start the Berkshire climb.

0.9 Turn right into parking lot at observation tower of Long View Gift Shop. Assemble on observation platform.

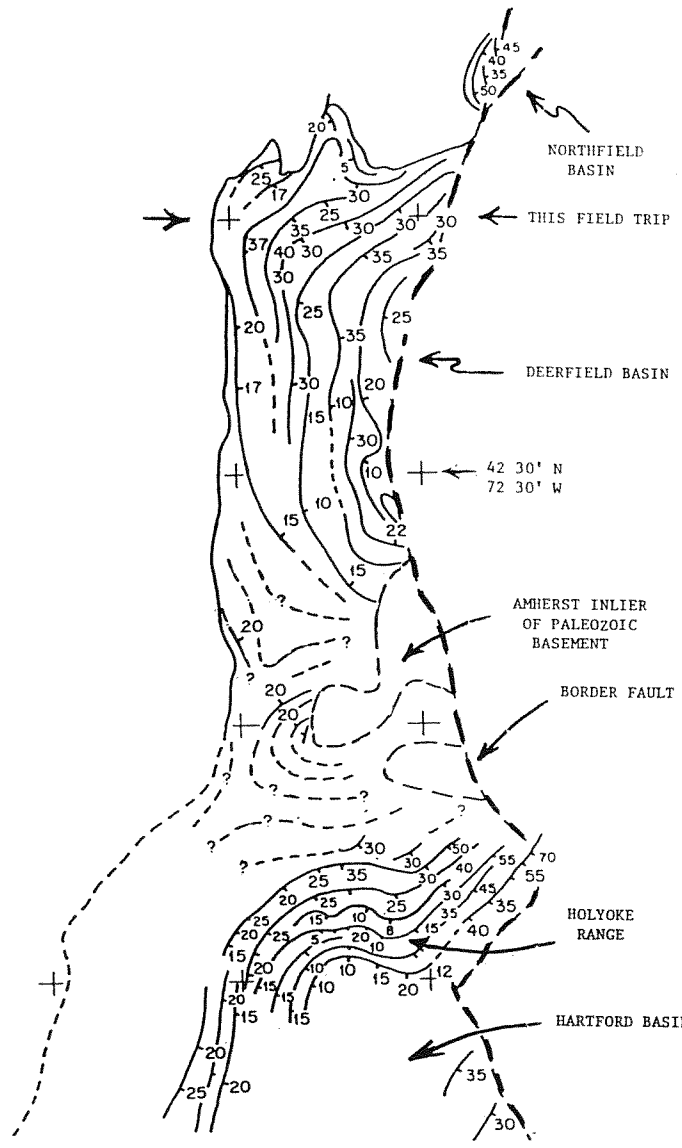


Figure 3. Generalized strike and dip orientations of the Deerfield basin. (Data from Wise, in press, 1992)

STOP 1. OVERLOOK OF DEERFIELD BASIN.

Topography . Our view is eastward across the basin and the town of Greenfield (Figures 1, 2, 3). The most distant hills are the Bronson Hill line of Devonian mantled gneiss domes just beyond the border fault. On a clear day, Mount Monadnock is visible on the most distant skyline at N60E. In the foreground, just beyond Greenfield is a ridge supported by the Deerfield Basalt that dips away from us towards the east at about 25 degrees. Trace this ridge southward (to the right) as it increases in height to become the Pocumtuck Range, whose ridge-like crest is held up by the particularly resistant diagenetic facies of the Sugarloaf Arkose cemented by interstitial crystals of albite (Taylor, 1991; Taylor and Hubert, 1991). The Deerfield Basalt forms the east-facing dip slope. In contrast, the lowlands before us are eroded into less resistent strata of the New Haven Akose. These sandstones are in the albite diagenetic zone beneath Greenfield and the area to the north (Taylor, 1991). To the south of Greenfield, the sandstones are in the illite diagenetic zone, and particularly susceptible to weathering. In the illite zone, there has been dissolution of plagioclase grains and albite cement accompanied by precipitation of interstitial illite clay (Taylor, 1991). On the Skyline behind the Pocumtuck Range is Mount Toby, held up by the resistant, early Jurassic Mount Toby Conglomerate. Just behind and to the left of Mount Toby is a gap in the skyline formed by the erosional removal of materials weakened by the border fault.

Glacial Lake Hitchcock. This former lake is not a focus of this field trip but its effects extended across much of the topographic basin. A dam of glacial drift in Rocky Hill, Connecticut, south of Hartford, ponded a large, elongate glacial lake from about 15,600 to 12,400 years ago. This lake occupied the valley floor below us to elevations of about 320 ft above sea level, leaving a record of lake-bottom varved clays and extensive deltaic deposits of sand and gravel. Stop 5 at Barton Cove is the site of a former waterfall with plunge pools of the Connecticut River, created as the river cut down through a Hitchcock delta to encounter a resistant bedrock ridge.

Outcrops on the west side of Route 2. These are low-grade metamorphic rocks, formerly marine sedimentary rocks of Devonian age, in the Connecticut Valley-Gaspe Synclinorium, that comprise the eastern half of the Berkshires. An extension of this metamorphic "basement" formed the floor for deposition of the Upper Triassic-Lower Jurassic Sugarloaf Arkose. The rocks in the road cut are phyllites, phyllitic graywackes, and metamorphosed carbonates of the Devonian Gile Mountain Formation. Many examples of quartz veins, multiple periods of folding, and spaced cleavages are visible, deformations generally ascribed to Acadian disturbances.

Regional setting. The Berkshires are now generally interpreted as the product of complex, major thrusting to the west in both Taconian and Acadian time. Cross sections by Stanley and Ratcliffe (1983) and seismic work in Vermont (Ando et al., 1984) suggest that the master thrusts beneath the Berkshire-Green Mountain uplifts may extend eastward to the vicinity of the Connecticut Valley. Farther east, the Bronson Hill structures may be another mass which was thrust and folded westward across the eastern edge of the displaced Berkshire-Green Mountain masses. Between the two zones of higher grade metamorphic rocks, a zone of low-grade metamorphism marks the present Connecticut River Valley along the VT/ NH border. This zone of weakness disappears beneath the basal unconformity of the Deerfield basin to reemerge at the southern end of the Hartford basin. This geometry makes a strong case for the location of the extensional Deerfield and Hartford basins along a tectonically weak zone. A geologic puzzle yet to be resolved is that all the major deep structures in north-central Massachusetts and the VT/NH border region seem to dip to the east whereas the shallow eastern border faults of the rift basins dip moderately to shallowly to the west.

0.9 Exit parking lot, turning left (east) back down hill on Route 2.
1.9 Turn left at light on Colrain Road opposite Friendly's Restaurant.
2.6 Pass entrance to Greenfield Community College. See Richard Little for information on Greenfield as the world center for armored mudballs.
3.7 Right turn on Nashs Mill Road.
5.0 Pass Captain William Turner Memorial sign. He was dragged from his horse and killed on the river bank just east of here, along with 40 his men, after massacaring 300 (?)

Indians in a dawn raid at Turners Falls on May 19, 1676.
5.3 Stop sign. Turn right on Lyden Street.
5.5 Left onto Silver Street.
6.0 Left onto Country Club Road.
6.5 Park on east side of Country Club Road (Swamp Road) just before the first Interstate 91 overpass bridges.

STOP 2. FLUVIAL SUGARLOAF ARKOSE.

The roadcuts are a typical outcrop of fluvial reddish-brown strata of the Sugarloaf Arkose of Late Triassic age. This stop is in the albite diagenetic zone with extensive albite overgrowths on the detrital sodic plagioclase grains (Taylor, 1991). These strata are mostly channel sandstone with minor conglomerate interbedded with thin layers of overbank mudstone. The sandstones are dominated by horizontal lamination of the upper flow regime with rare cross-beds that formed on dunes of the lower flow regime. A number of small to moderate-size channel scours plunge steeply down dip into the outcrop. Near the fire plug, the underside of one large scour channel exposes groove and scour marks, including possible flutes that suggest paleoflow towards the southeast. Boulders of granite pegmatite up to 10 cm in diameter occur in the sandstones. The high ratio of sandstone to mudstone, coarse grain size of the clastics, dominance of horizontally-laminated sandstone, and presence of small channels combine to suggest that this is the record of a braided river.

A map of paleocurrent azimuths for the entire Sugarloaf Arkose suggests that braided rivers flowed from highlands located to the north and northeast, crossing the basin towards the south and southwest (Figure 13 in Stevens and Hubert, 1980). We think it likely that the fluvial strata of the Sugarloaf Arkose accumulated during inital sagging of the basin in late Triassic time, before the major eastern border fault was subsequently established in Early Jurssic time.

6.5 Leave Stop 2. Make a U-TURN (with caution) using parking pull-off just beyond the second bridge of the interstate overpass.
7.4 Stop sign. Left on Silver Street.
8.0 Straight through the traffic light of Routes 5 and 10.
8.5 Traffic Light of Route 2A. Left turn.
8.9 Super Stop & Shop. Pull into parking lot on the right. Outcrops are in cuts behind the building.

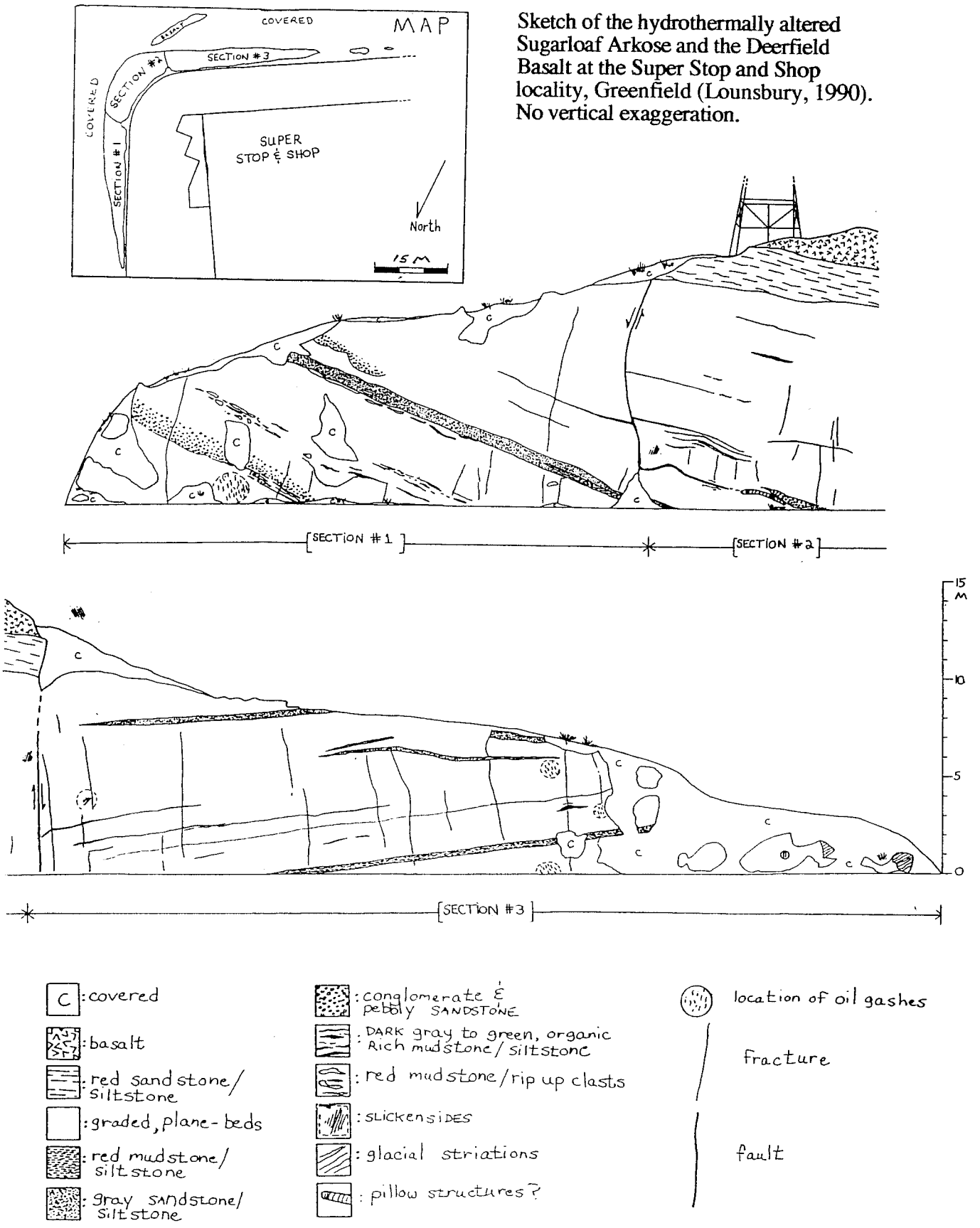
STOP 3. TOP OF SUGARLOAF ARKOSE AT REAR OF GREENFIELD SUPER STOP & SHOP.

These exposures of the top 25 m of the Sugarloaf Arkose extend upwards to the contact with the Deerfield Basalt of Early Jurassic age at the top of the outcrop (Figure 4). On the state geologic map of Massachusetts, these uppermost beds of the Sugarloaf Arkose are interpreted as being of Early Jurassic age because a palynoflora of this age occurs in the lacustrine sequence along Route 2 in Greenfield (Cornet, 1977; Figure 9 in Stevens and Hubert, 1983). The exact location of the Triassic-Jurassic boundary is unknown, but is assumed to be somewhere in the fluvial redbeds of the upper part of the Sugarload Akose.

Sandstones and minor conglomerates dominated by horizontal lamination of the upper flow regime comprise most of the strata at this stop. The prevalence of horizontal lamination, and presence of some graded bedding, in these coarse clastics suggest that they were deposited mostly on a distal alluvial fan-sandflat system (Taylor, 1991). Beds of lacustrine gray mudstone, 10-20-cm in thickness, are interbedded with these sandstones, demonstrating that lakes lay off the toes of the fans. Locally, the lacustrine beds show soft sediment deformation and suggestions of gravity flow to the SSE. The fan sandstones include some layers of igneous and metamorphic pebbles and rip-up clasts of the lacustrine gray mudstone.

The 1.5-m sequence of red siltstone just beneath the pillowed basalt is interpreted as a playa deposit. Cross-bedded sandstones exposed at the southwest end of the outcrop are probably fluvial in origin. The average paleoflow direction based on 6 cross-bed sets is southeast with an azimuth of 166 degrees (Loundsbury, 1990).

Taylor (1991) has described the petrography of these units in considerable detail. Pervasive cementation of the sands by albite overgrowths on detrital sodic plagioclase grains started at moderate burial depths as



Sketch of the hydrothermally altered Sugarloaf Arkose and the Deerfield Basalt at the Super Stop and Shop locality, Greenfield (Lounsbury, 1990). No vertical exaggeration.

Figure 4. Details of Stop 3 (unpublished sketches by Lounsbury, 1990).

indicated by minus-cement porosities of about 20 volume percent. Topographic-driven groundwater flow from surrounding highlands carried Na, Si, and Al derived by weathering of oligoclase-rich gneisses and schists through the partly compacted, permeable sands.

Subsequently, the sandstones suffered bleaching by hot, rising fluids whose temperatures were elevated above the regional geothermal gradient present at this paleo-depth of the basin. Several normal faults, including the prominent one in the central part of the outcrop, cut the arkose and the basalt. Some of them are mineralized with a quartz druze coated with brown iron-oxide stains. These faults commonly have less than a meter of displacement and strike NE, dipping steeply NW. The rising, silica-bearing solutions had temperatures of 150-200 degrees C based on the presence of type 2b authigenic chlorite in sandstones associated with alteration of mudstone intraclasts about 20 m from the top of the Sugarloaf Arkose ((Taylor, 1991). The chlorite may have been the result of a local thermal event and not an indicator of basin-wide temperatures (Taylor, 1991).

Hydrocarbons generated from the lacustrine gray mudstone migrated into extension fractures now stained with bitumen. Good examples are present along the NW corner of the cut. The problems of hydrocarbon genration and migration are discussed at Stop 6D.

8.9 Leave parking lot, turning right (north) on Route 2A East.
9.9 Merge with Route 2 east. Bear right.
10.1 Pass Factory Hollow historical marker on the right. The open area ahead and to the left was the industrial area of Greenfield in the days of water power. It was the site of mills dating from 1784 with the major mills built in 1830 as a number of four story granite structures.The mills provided woolens for the Union army in the Civil War but ceased operations in 1872 when Turners Falls industrialists bought the water of Falls Creek, which powered the mills. The industrialists piped the water across the Connecticut River for their factories, leaving the Greenfield industrial area high and dry. The path of the water line blasted out of bedrock of the Connecticut River floor is visible at low flow. The mills burned in 1933 and only a bell tower remains.
10.5 Stop 4: parking lot for Turners Falls dam overlook.

WATCH TRAFFIC WHILE CROSSING ROAD.

STOP 4. DEERFIELD BASALT IN ROUTE 2 ROAD CUT.

This exposure includes almost all of the 55-m Deerfield Basalt. The lower part is not well exposed here but the base can be located in the woods just to the left of the cut. Time constraints preclude stopping at the exposures of the base and lower pillowed zone in the Mackin sand pit 0.5 km to the west, as described by Wise and Belt (1991).

The basalt comprises two flows (Figure 5). The lower flow is about 17 m in thickness with a few poor pillows exposed low in the flow at the left of the crop. The basalt of the lower flow passes upward through a zone of normal faults with about 5 m displacement into the oxidized, vesicular top.

The thicker upper flow consists of massive basalt at the base with increasingly coarser crystal size upwards until finer grain sizes are reached in the chilled, vesicular, oxidized top of the flow. Near the middle of the flow are several sill-like zones, 5-10 cm in thickness, with very coarse grained diabasic texture and local oxidation. These zones are interpreted as the result of trapped volatiles, possibly derived from the overridden lacustrine beds (Figure 9 in Stevens and Hubert, 1983). We infer that enough pressure developed to exceed lithostatic pressure within the cooling mass.

Large columnar joints distinguish the lower half of the upper flow, with many joints showing paleo-horizontal arrest marks normal to the column axes. These marks, 1-2 cm steps on the joint surfaces, are evidently formed by rapid propagation of the main joints, followed by cessation of propagation and decay of the local stress field. During the time needed for further cooling and rebuilding of the local, tensile stress, far field stresses slowly reoriented the fracture before initiation of the next cycle of joint propagation. The presence of these columns only in the lower half of the flow suggests that their propogation direction was upward from the base. Because the columns cut the diabasic sills, their formation post-dates sill formation, an interpretation supported by lack of sill material or selective alteration along the columnar joints.

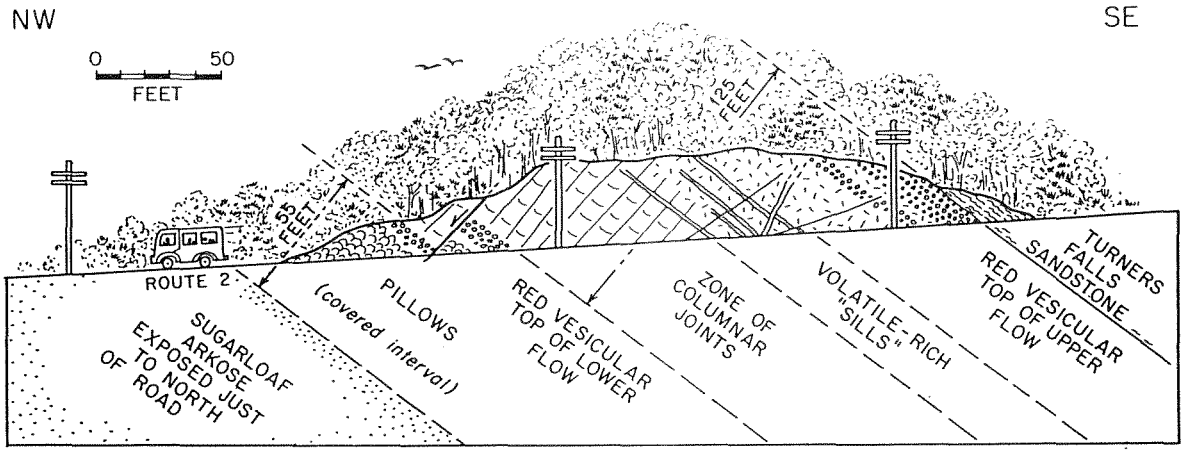


Figure 5. Detail of Stop 4. Route 2 road cut in Deerfield Basalt opposite parking lot for Turners Falls Dam overlook, Turners Falls, MA.

Many tectonic joints also are present with N45E strike, some with associated calcite and/or quartz veins (Figure 6). Some of these veins show pull-apart structures indicative of down to the NW motion during precipitation of vein minerals. The joints and veins are interpreted as part of general NW-SE extension associated with later wholesale tilting of the basin.

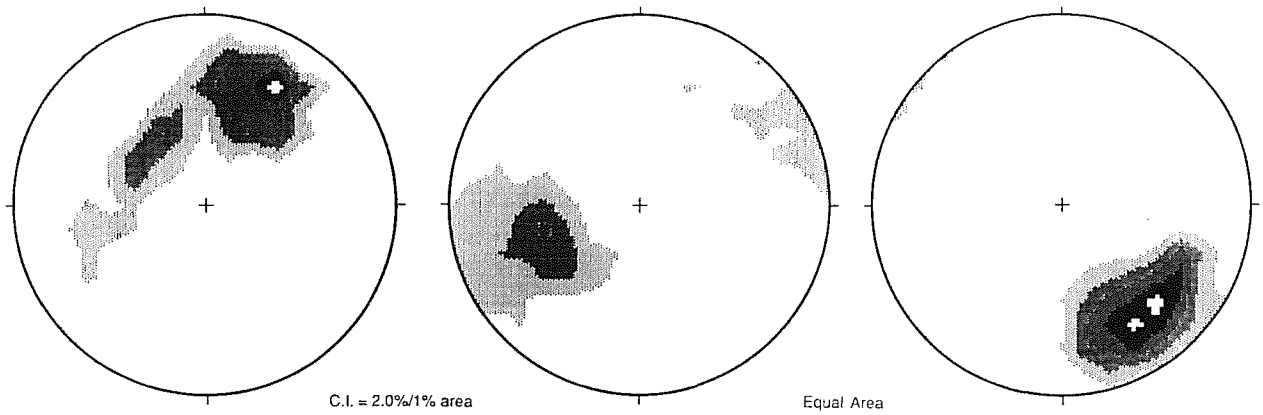
Several families of minor slickenlined fault surfaces are present, mostly as reactivations of pre-existing joint surfaces. The steps of arrest lines can be used to provide evidence of transpressional and transtensional relationships and the changing stress history. The sense of fault displacements, commonly only a few centimeters, can be determined using the associated differences in mineralization and local polish on opposite sides of asperities. Crosscutting or overprinting relations of the slickenlines are not common but where present suggest that strike-slip motion followed major dip-slip motions. A) Oldest is a series of NE striking dip-slip normal faults, the largest of which is near the western edge of the crop with several meters of displacement. Dips tend to be sub-parallel to the columnar joints reflecting a strength anisotropy during the tilting. Mineralization is largely calcite with minor quartz. B) Strike-slip and oblique-slip slickenlined surfaces are comon. These may have calcite and quartz present but polished surfaces are also comon. Both dexteral and sinistral senses of displacement are present. Stress tensor analysis indicates a NE-SW compression and NW-SE extension for this system (Figure 6). C) A second set of younger, shallow-dipping faults with reverse motion is marked by polished surfaces and hematitic staining. Motions are appropriate to the NE-SW compression of set B.

Throughout the field trip, this same story will appear of NW-SE extension followed by N-S or NE-SW compression. We interpret this pattern as due to the change from the early tectonic rifting environment to the drift environment as first the Labrador Sea and then the main North Atlantic opened and ridge push became more effective.

O'toole (1981) found that the bulk chemistry of the Deerfield Basalt at this stop has many similarities with the Holyoke basalt of the Hartford basin, but that the Deerfield Basalt is much higher in sodium and lower in calcium and strontium. The average composition of the plagioclase is sodic andesine. She concluded that the Deerfield Basalt was probably a tholeiitic magma that was splitized by reaction with Na-rich groundwaters moving through the basin fill (O'toole,1981).

Figure 6. Data from Stop 4.

STRESS TENSOR ORIENTATIONS CALCULATED FROM 29 MINOR FAULT MOTIONS FROM THE DEERFIELD BASALT AT THE TURNERS FALLS DAM OVERLOOK



SIGMA 1 SIGMA 2 SIGMA 3

USE CARE IN RETURNING ACROSS TRAFFIC TO THE PARKING LOT.

10.5 Leave parking lot, turning to the right, continue east on Route 2 through light.
11.2 Pass broad flats of boat ramp area on the right. This is part of a Connecticut River terrace formed as a temporary stillstand in the downcutting through the former Lake Hitchcock delta which covered the area. The top of the old delta is visible as the flat skyline on the opposite side of the river.

This site is the location of the 1676 dawn battle (massacare?) of the sleeping Pocumtuck Indians by a group of Deerfield settlers led by Captain Wiliam Turner. The marker by the traffic light notes that the entire encampment of 300 Indians were killed. Local historical documents say only one settler was lost in the battle but a rumor of the impending arrival of King Phillip and 1000 of his braves sent the settlers into a disorganized scramble toward Deerfield. The books say that Indians from nearby villages attacked the retreating settlers but suspicion is that some of the 300 "dead" Indians counterattacked the rag-tag band. Captain Turner was dragged from his horse and killed while crossing the Green River. In all, 40 settlers were killed in the retreat before a Lieutenant Samuel Holyoke got some defensive order into the mob (?) and supervised a withdrawl into the fortifications of what is now Historic Deerfield. In honor (?) of this event, the falls and the town later founded across the river were named after Captain Turner.

11.5 Turn right at entrance to Barton Cove Preserve.
11.9 Drive to the innermost parking area. The preserve provides rest rooms, walking trails, camp sites, boat and canoe docks, fishing, scenic views, and picnic tables.

STOP 5. BARTON COVE KINK FOLDS AND LACUSTRINE BEDS IN THE TURNERS FALLS FORMATION.

The Barton Cove Peninsula, projecting into the Connecticut River, is now a public access nature area maintained by Northeast Utilities. In the 19th Century, it was the site of a dinosaur track quarry that provided numerous slabs for Hitchcock's collection at Amherst College. From the parking area proceed on foot through the barrier up to the top of the hill. Turn right along an access road for 50 m and then turn right onto a series of wooden steps leading down over the cliff.

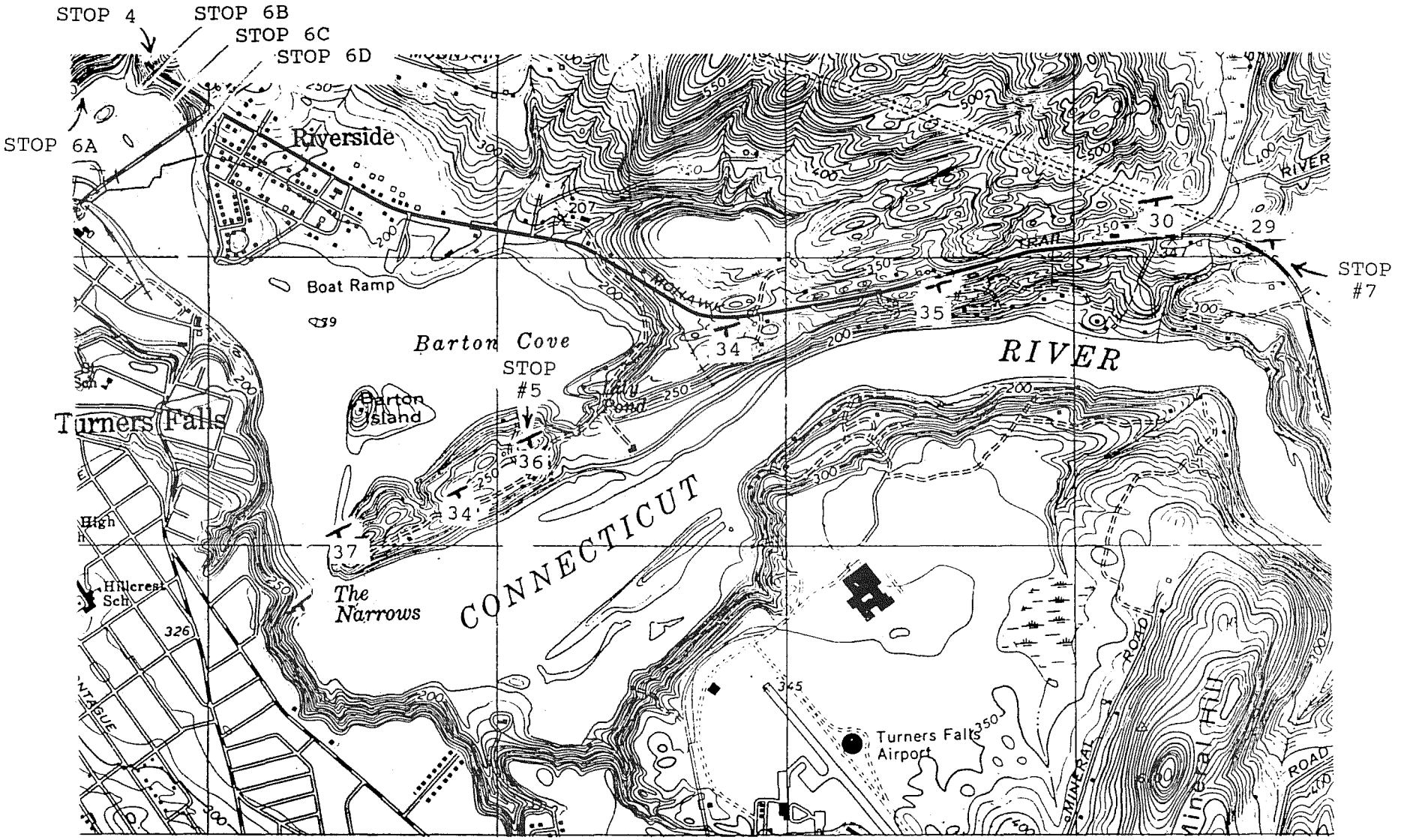


Figure 7. Barton Cove Area showing a number of stop locations. Note: Stops 5 and 7 are essentially on strike and show facies changes northeastward from lake bottom units to alluvial fan conglomerates. The resistant lake beds of Barton Peninsula formed a ridge in pre-Lake Hitchcock time. A Hitchcock-age delta covered the ridge to the 320 foot level of the Turners Falls Airport. Subsequent downcutting through that delta by the Connecticut River produced a waterfall across the ridge and the associated plunge pools at Stop 5, the "Lily Pond.". Subsequently the river has shifted its channel to "The Narrows."

The geomorphic development of this area centers on the resistant lacustrine beds that form the backbone of the peninsula (Figure 7). In pre and late-glacial times, these beds were eroded to a ridge cut by a stream valley about1 km to the SW of where we are standing. Whether this valley was the ancestral Connecticut or one of its tribituaries is unknown. During Lake Hitchcock time, about 15,600 to 12,400 years ago, the massive Montague delta of sand and gravel was built across the area at glacial lake level, about 320 ft present elevation. The depositing agent was probably the ancestor of the present Connecticut River that carried glacial debris from the lowland separating VT and NH. This delta buried the topography of the area, including the peninsular ridge and the bedrock exposures in the dam area at Turners Falls. With the draining of Lake Hitchcock about12,400 years ago, the Connecticut River cut rapidly through the deltaic deposits to expose the former bedrock topography. With a larger discharge than the Connecticut River has today, the water cascaded over the Barton Cove ridge and the bedrock at Turners Falls, creating spectacular plunge pools, now flooded as the small lagoon of the preserve. Temporary stillstands of erosion at the dam area produced the flats on which the 1676 massacare occured. With the passing of time, seepage and erosion of weaker deltaic deposits in the former bedrock valley through the ridge caused the main channel of the Connecticut River to shift and reoccupy the old valley. The result was abandonment of the plunge pools and deepening of the present narrows at the end of the peninsula. Prior to the recent raising of water level by additions to the dam, these pools were isolated small lakes known as the "lilly ponds."

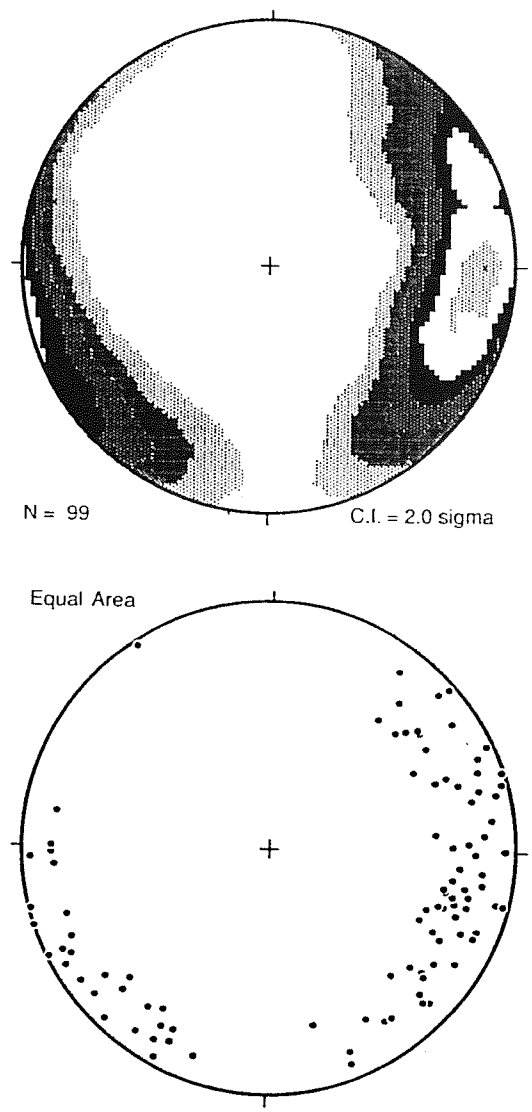


Figure 8. Axes of kink folds in the Barton Cove Area, in and around Stop 5. Data from Handy (1976). Kamb contours with c.i. = 2.0 sigma.

The Barton Cove lacustrine beds are a distinctive stratigraphic unit of gray and black dolomitic mudstone at the top of the Turners Falls Formation (Handy, 1976). The mudstones commonly comprise alternating layers up to a few millimeters in thickness dominated by 1) illite, chlorite, and vermiculite and 2) dolomite and ferroan dolomite (Handy, 1976). Some of the latter layers are rich in albite. The lake evidently was alkaline with substantial dissolved Na, Ca, Mg, and bicarbonate ions. Handy speculated that the layering may be annual in origin, with the carbonate layers deposited during seasonal dryness.

Handy mapped this lacustrine unit southward in three discontinuous segments for about 14 km to the vicinity of Chard Pond west of Mount Toby (Figure 1). If the segments are continuous through the covered intervals, as Handy believed, then a sizeable lake existed in front of and between alluvial fans. At places along its length, thin beds of horizontally-laminated or graded sandstone are interbedded in the gray mudstone. The graded beds record turbidity currents that coursed down the sides of the lake. Alluvial fans from the north prograded into the lake as will be demonstrated at stop 7. Handy interpreted the redbeds above and below the lacustrine strata as playa and fluvial deposits.

The lake beds are disrupted into large breccia masses and kink folds in the Barton Cove exposures. All gradations from overturned folds to complete brecciation are present. The fold geometry tends toward kinking and shortening of bed lengths. Axial planes dip mostly toward the south whereas fold axes trend generally E-W (Figure 8).

In the past, discussions have centered on whether the deformations of these lacustrine strata represent gravity slide phenomena or whether they are fundamentally tectonic in nature. The issue is not completely resolved as yet, but the following observations are pertinent. The kink folds are clearly compressional rather than extensional. Also there is abundant evidence in the dam area at Turners Falls of thrust faults, spaced cleavages, and conjugate strike-slip faults that indicate a late-stage regional N-S to NE-SW compression. These observations have caused most present opinions about the deformed lacustrine beds at Barton Cove to shift towards the tectonic origin. Under this interpretation, the lake beds formed strength discontinuities in the stratigraphic column and localized the compressional deformation at this horizon.

There maybe a gravity-driven story yet to be unravelled. At Taylor's Hill, 8 km to the south, a 5-acre tract of basement rocks is present just above this same lake-bed horizon. A similar mass also occurs at Whitmore's Ferry along the Connecticut River. These masses are rootless and probably were emplaced into the Jurassic basin by some landsliding process.

RETURN TO PARKING LOT FOR LUNCH AT TABLES DOWN ON THE NEARBY POINT. REST ROOMS ARE AVAILABLE.

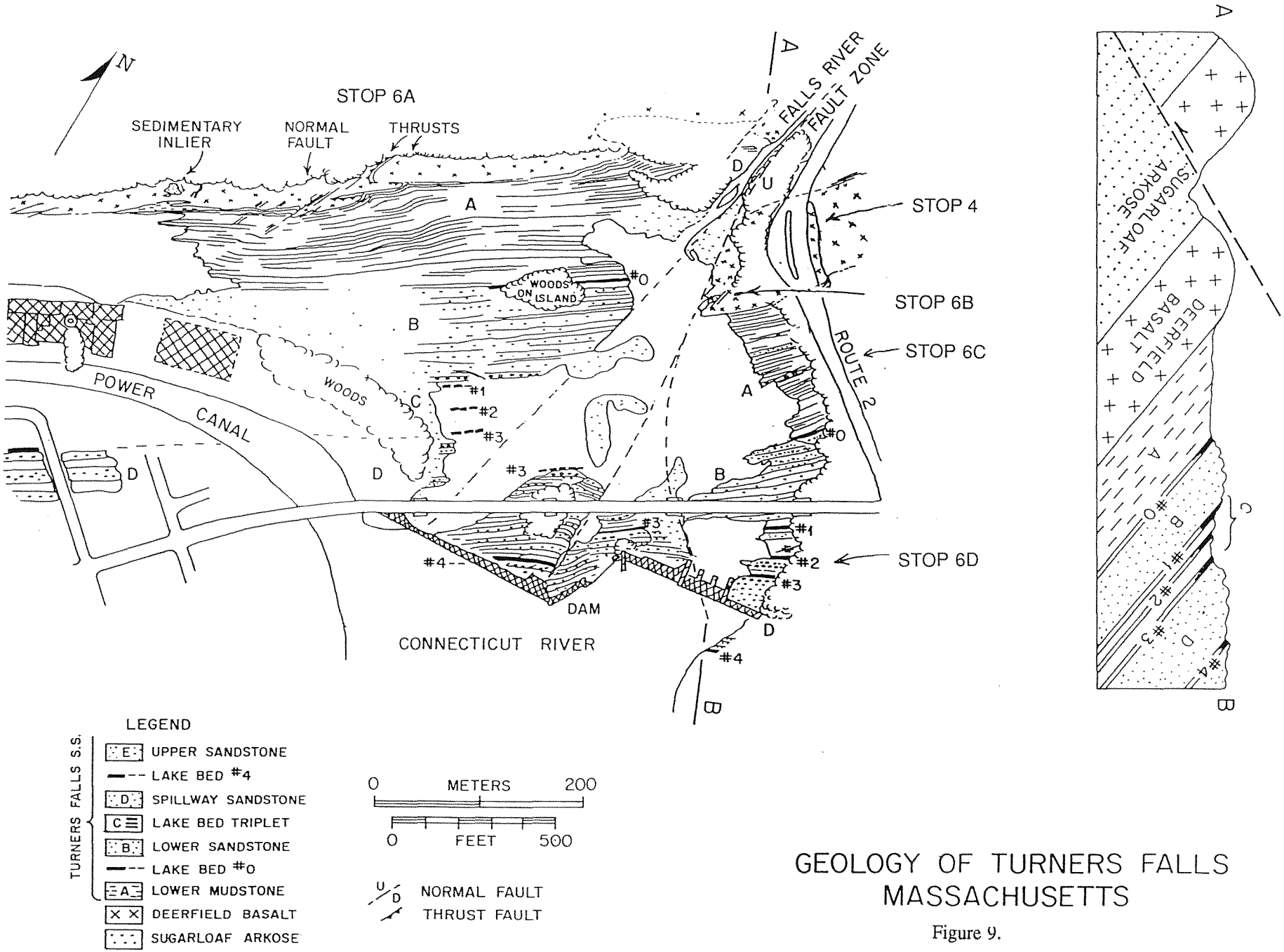
11.9 Leave Barton Cove and return to overview parking lot at Turners Falls dam.
13.3 Overlook parking area. Stop 6.

Parts of Stop 6 are accessable only at low water. If there is rain or the rocks are wet, <u>use judgement and great caution on the riverbed</u>. Descend from the parking lot to the floor of the river by following a steep path down the crest of the basalt ridge from the parking lot.

STOP 6A. NORMAL FAULT AND THRUSTS CUTTING TOP OF BASALT.

Proceed downstream for 300 m, crossing the mouth of Falls River enroute. Depending on the flow level of Falls River and the level of personal aversion to wet feet, an alternative route is to walk west across the Route 2 bridge; then follow a poor path along the cliffs of Falls River to the shore of the Connecticut River. Stop 6A is located on the river bank 200 m downstream, just beyond a cliff-like projection of the basalt (Figures 9, 10).

<u>Normal fault</u>. The outcrop-size "normal" fault at Stop 6A is approximately a 1:25 scale version of a much larger fault that trends SSE from lower Falls River to the dam (Figures 9, 11). The larger fault produces an apparent map offset of the top of the basalt of about 200 m. The fault also splays upward into resistant sandstones to form the separate blocks that underlie the dam. An intermediate scale version of both of these structures is the Canada Hill Fault Zone, beneath another bridge 500 m downstream; this



GEOLOGY OF TURNERS FALLS MASSACHUSETTS

Figure 9.

COMPOSITE STRATIGRAPHIC COLUMN
TURNERS FALLS DAM, MASS.

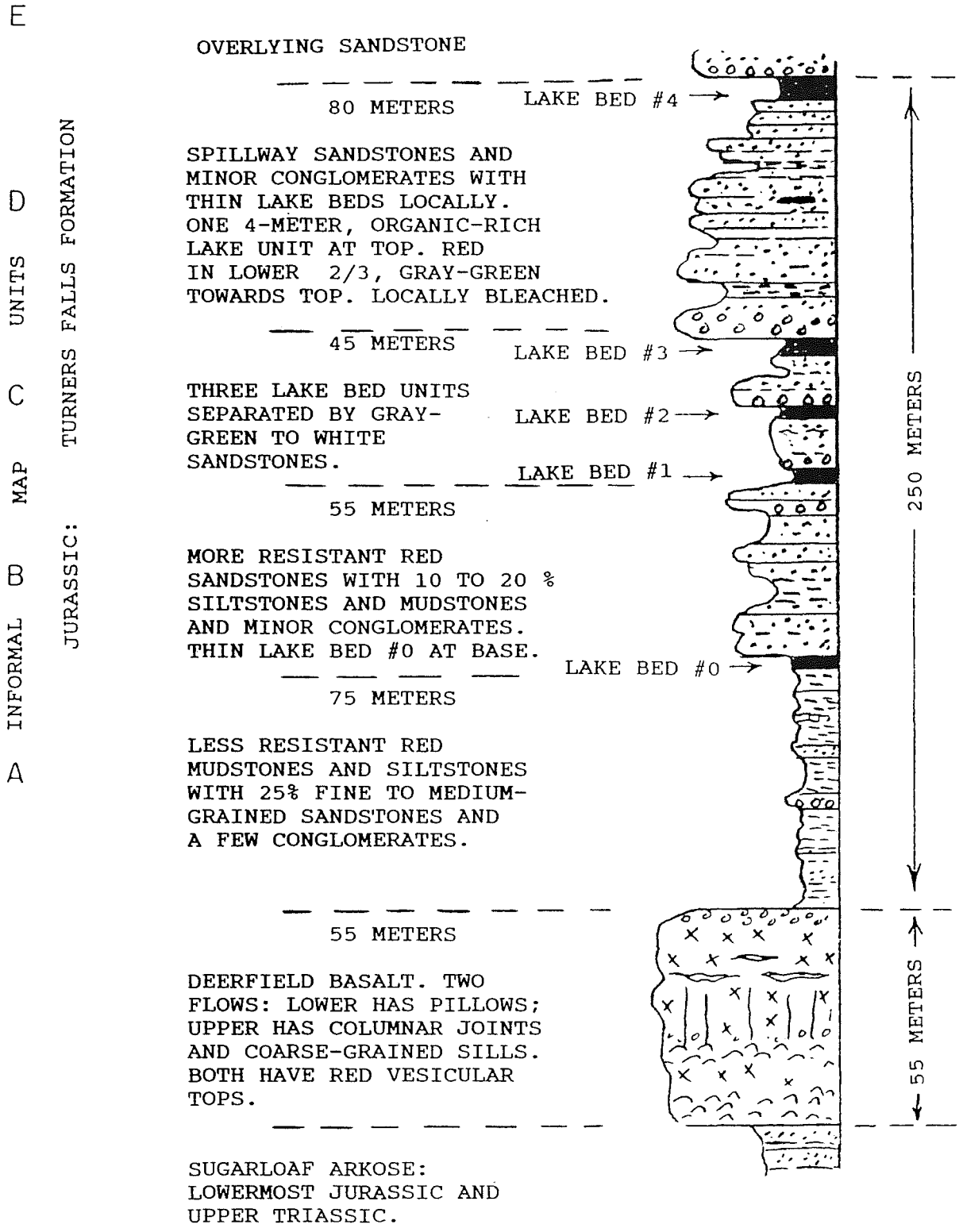


Figure 10. (A, B, C, and D are map units indicated on Figure 9).

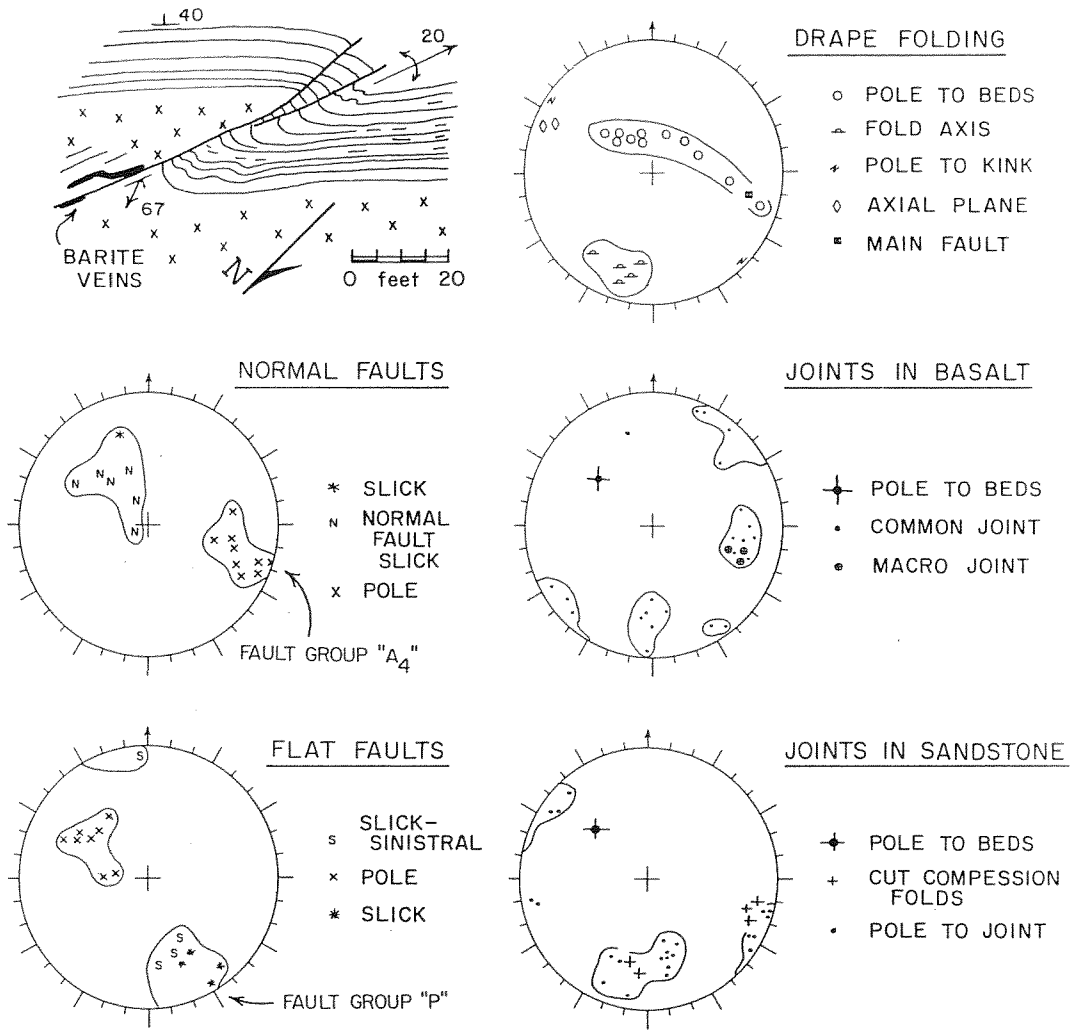


Figure 11. Normal faulting, drape folding, and jointing at Stop 6A.

fault and associated folds are well exposed at low water stages. Sketches and structural details are given by Wise,1988.

The "normal" fault in the basalt at Stop 6A is oriented N15E, 65W, approximately a paleo-vertical orientation. The displacement is about 3 m, down on the west, dying upward through the overlying strata as a series of minor splay faults to become only a gentle curve after about 25 m. The fault within the basalt shows slickenlines with a paleo-vertical orientation. Local mineralization includes quartz, calcite, barite, and minor sulfides. Please refrain from mining the barite for lab collections.

Tight folding, kinking, and crowding of the beds immediately above the basalt are restricted largely to the synclinal area of the downthrown block. Fold hinges plunge S15-25W at 20 degrees (Figure 11). In contrast, there is little folding of sedimentary beds on the upthrown side. The hanging wall of basalt is marked by a prominent 3-m-wide zone of jointing parallel to the fault. This fault/fold geometry is not typical of a normal fault where one might expect stretching of beds being draped across a paleo-vertical fault. Instead, the geometry resembles some of the forced folds of the Middle Rockies where highly confined, but weak, Paleozoic strata were forced to conform to the geometry of a strong basement lip during regional shortening. Here at stop 6A, the localization of jointing parallel to the fault on its upthrown side

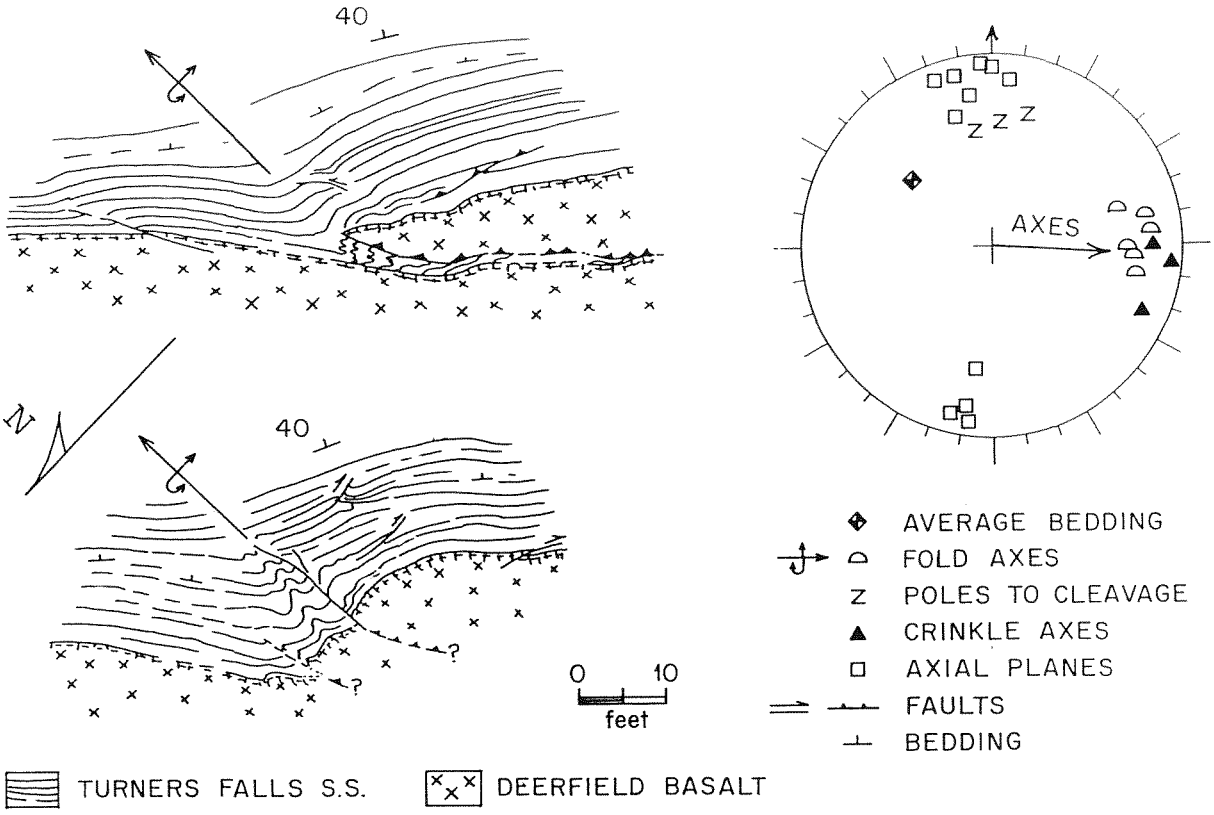


Figure 12. Compressional folding and thrusting at Stop 6A, 0.8 km downstream from Turners Falls Dam.

is interpreted as a local extension feature of the brittle basalt as the rock started to flex in the early stages of fault development. The details cited above speak against a simple extensile normal fault.

Thrust faults. Return northeast back up river about 50 m along the basalt-redbed contact where a second thrust fault cuts the top of the basalt. Structural details of this fault and other thrust faults and associated compresional folds are given in Wise, 1988.

The thrusts caused detatchment and northward displacement of the top few meters of the basalt, obliquely upward across the present dip slope (Figure 12). These detatchment may have been aided by pre-existing slab-like sheeting or jointing of the top of the basalt, a feature evident in the walk down river to this location. The basalt sheet in each thrust traps an average 0.5-m-thick septum of redbeds beneath it and drapes the overlying redbeds across its lip. In the outcrops just downstream and 5 m above the lower thrust, a septum of redbeds is also visible. There is no indication of metamorphism of this septum and local slickenlines occur on some of its contacts. We interprete this thrust as a continuation of the pattern of detatchment and thrusting of the uppermost part of the basalt as seen at water level. The thrust movements shortened and crumpled the overlying beds. These folds have axes that plunge almost due east and axial planes and/or associated spaced cleavages that strike east-west, dipping steeply to the south. The shortening and gentle folding of the overlying redbeds is traceable at lowest stage water levels about 100 m eastward along plunge or about 25 m upward through the stratigraphy.

Timing. Direct evidence of the relative ages of "normal" faulting versus thrust faulting is difficult to find at this outcrop. In the basalt cliffs between these two exposures there is a small face that shows a set of closely spaced extension fractures associated with the normal fault and overprinted by slickenlines typical of the thrust-fault displacements. This supports the general pattern seen throughout these riverbed exposures of ESE-WNW extension followed by N-S or NE-SW compression.

STOP 6B. MINOR FAULT BLOCK IN THE FALLS RIVER FAULT ZONE.

Return to the river's edge in the basalt just below the Route 2 parking lot. The basalt ridge is the SW end of a long outcrop zone of Deerfield Basalt. The basalt ends abruptly just beyond the water's edge, to reappear 200 m to the north along Falls River, from were it continues SW through the outcrops at Stop 6A. The wooded island in the river immediately on strike from the present Stop 6B shows poorly exposed strata of lake 0 of the Turners Falls Formation. Just as the small fault at Stop 6A splayed upward (south) into several fault blocks, the Falls River fault zone splays southward under the main bridge to break the massive spillway sandstone into several large blocks that form the abutments for the dam (Figure 9). Stop 6B illustrates some of the minor structures associated with the Falls River fault zone.

Many of the fault structures at this stop are parasitic, that is they make use of pre-existing joint surfaces for their motion. The descending path from the parking lot follows a prominent joint set oriented N10E, 40W. This set is similar to one at Stop 6A along the upthrown side of the faulted lip of basalt. When viewed from the bridge above the dam, this joint zone of perhaps 15 m width in the basalt can be seen as a concentrated set of parallel planes. The set is also well developed in the overlying sedimentary units where the joints shallow in listric fashion toward the main fault zone. This local family of joints is interpreted as bending and extensional structures produced by the initial stages of very minor bending along the future fault zone. Once the main fracture occurred, these joint surfaces were available to serve as planes of minor fault adjustment and as hosts for carbonate veins.

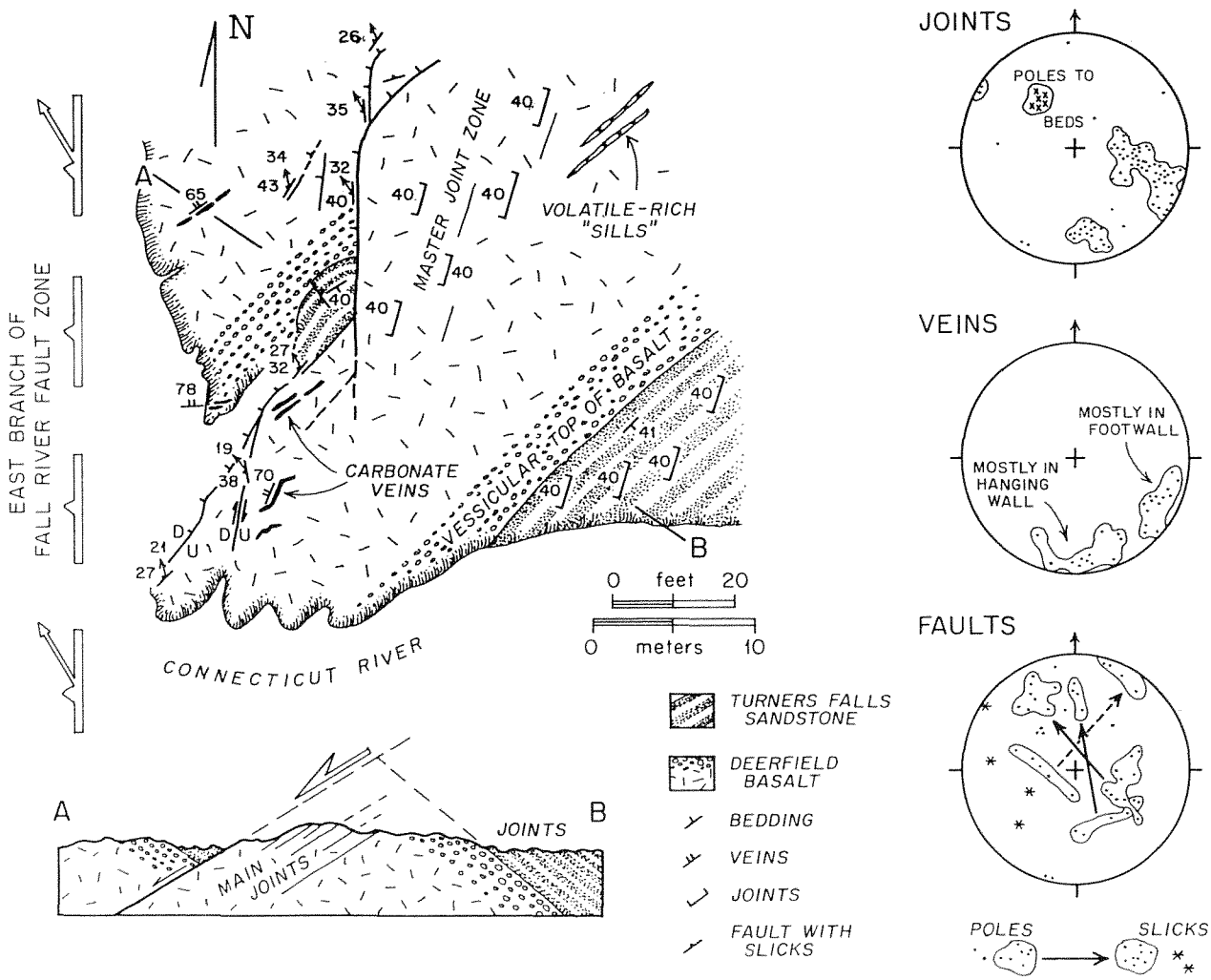


Figure 13. Details of minor fault motions of the Falls River Fault Zone, Stop 6B. The location is on the edge of the Connecticut River, 100 m. SW of the Route 2 dam overlook parking area.

At river level, the upper contact of the vesicular basalt with the overlying redbeds is obvious, but 20 m downstream this contact is a small depression that contains an in-faulted block of the overlying redbeds (Figure 13). The beds in the block dip SE, are unmetamorphosed, and at their base appear to be in conformable contact with the vesicular top of the Deerfield Basalt. However, along strike to the NE and downdip in the small block, they are truncated by faulted and slickensided faces of basalt. Throughout the area there are many small fault planes, mostly striking N to NE with westerly dips; the associated slickenlines plunging at shallow to moderate angles to the NW. A few very shallow-dipping faults have slickenlines to the NE. Individual faults show abrupt changes in strike as their planes merge with or truncate against other faults with strike ranging from N to NE. In spite of the variability in strike, the lines of intersection of the dipping fault planes have a tightly constrained common orientation parallel to the slickenlines (about 30-degree plunge to the NW), a requirement for geometric compatability.

This fault complex is interpreted as the result of interplay between slightly older NE-striking extensional joints and veins and the N to NNE- striking fault zone and its associated initial joint set. Once a brittle lip started to develop on the upthrown side, scollop-like masses were dropped and dragged along the fault zone. The facts that the average strike and dip in the small block are essentially the same as those in the overlying redbeds indicate that drag was accomplished with relatively little rotation of individual masses. Stress fields may have rotated as indicated by the contrast between 1) the NE strike of most veins in the footwall and 2) the EW strike of veins in the hanging wall.

STOP 6C. PLAYA REDBEDS OF THE LOWER TURNERS FALLS FORMATION.

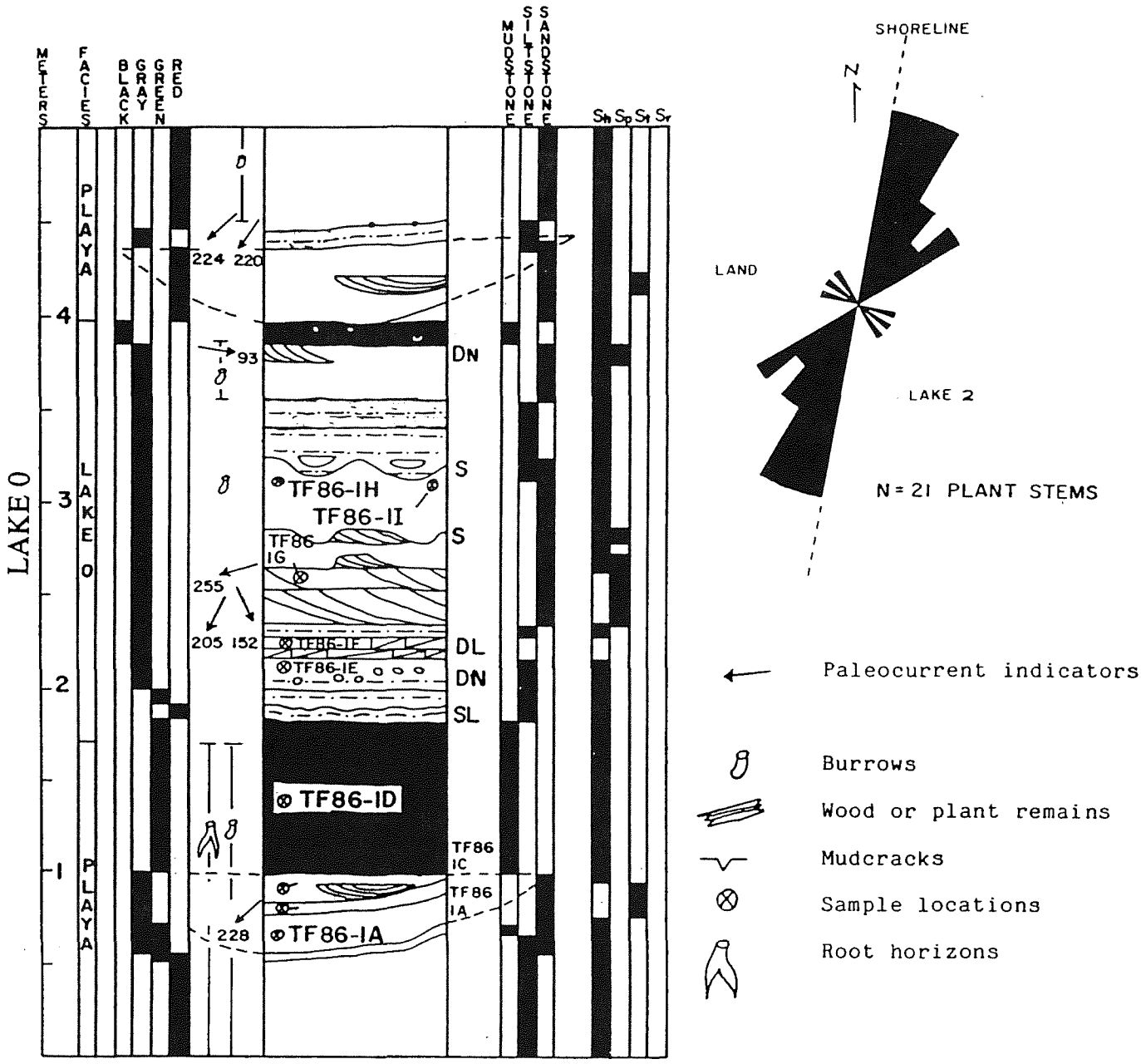
The sequence of redbed sandstones and mudstones that lie on the Deerfield basalt are interpreted as the record of a playa that succeeded the lava flows of the Deerfield Basalt. The initial playa muds and sands filled fissures and irregularly-shaped openings in the upper surface of the lava, visible as we cross the contact between the basalt and Turners Falls Formation. These are "neptunian dikes" injected from above.

Here at Stop 6C, the evidence for accumulation of the redbeds in the topographically closed basin of a playa includes the following. 1) The bedding planes are smooth and level without plant root traces. 2) Graded beds that record flood events, thicker examples of which proceed from plane-bedded sandstone with chips of red mudstone -> climbing ripple cross-lamination of the stoss erosional type -> laminated mudstone -> mudcracks. 3) Superabundant burrows, perhaps made by insects. 4) Superabundant mudcracks. 5) Abundant layers of ripple marks. 6) Occasional raindrop impressions. 7) Dinosaur and other reptile tracks were once common, but have been mostly removed. 8) Groove marks made by tools (plant fragments and pebbles ?) were dragged over the muds by flood waters. Over time, the playa surface was built up by floods that varied from major events to small-scale mud-laden ponding of runoff. These flood waters coursed to the southwest across the playa surface from an alluvial fan located to the northeast. The nearly flat surface would be covered by a playa-lake for a few hours to weeks after each flood event, to be followed by drying and desiccation of the muds.

In the middle of the these playa redbeds is a 2-m-thick fluvial channel body of pebbly arkose with some clasts up to 4 cm in size. The sandstones are in part cross-bedded, and there is minor channelling into the underlying mudstone. A river flowed a substantial distance across the playa surface, fed by rivers that flowed down from highlands to the N and NE.

The sandstones at Stops 6 C and D are variable in composition, averaging a lithic arkose (Meriney, 1988; Taylor, 1991). They all have an albite diagenetic pattern (Taylor, 1991).

From this channel northward along the shore of the Connecticut River to lake sequence 0, faint, spaced cleavages are visible as subtle parallel lines on bedding. Also visible are many depositional grain lineations in horizontally laminated sandstones. A change in strike of about 10 degrees occurs in this area, but it is uncertain whether this is the result of later faulting or of rapid subsidence during sedimentation. The former seems more likely. About 6-8 m below lake 0, a zone of nested scours (?) occurs at the base of a sandstone which has fragments of the underlying mudstone imbricated by flow to the SW.



Measured section of lacustrine bed 0 in the Turners Falls Formation at the Turners Falls dam (Meriney, 1988). Symbols are DN = dolomite nodules, DL = dolomite laminate, S = scour surface, SL = folded layer, CN = carbonate nodules, and CR = rip-up clasts. The primary sedimentary structures in the right-hand four columns (from left to right) are Sh = horizontally-laminated sandstone, Sp = planar cross-bedded sandstone, St = trough cross-bedded sandstone, and Sr = ripple cross-laminated sandstone. Sample numbers refer to samples analysed by Meriney (1988). The equal-area plot shows the orientation of 21 plants stems, inferred to be parallel to the shoreline of lake 2 (Meriney, 1988).

Figure 14. Data from Stop 6C.

STOP 6D. LACUSTRINE STRATA AT THE TURNERS FALLS DAM.

Walking up section, we cross lakes 0, 1, 2, and 3 interbedded with playa redbeds and minor fluvial strata (see the mapping units on Figures 9, 10). The lacustrine gray/black mudstones are Van Houten cycles that record the initial development, deepening and expansion, and contraction of perennial lakes. At this time, the basin was a topographically closed depression, and the lakes existed for long periods of time, evidently some thousands of years, when precipitation exceeded evapotranspiration. Olsen et al.(1989) used power spectrum analysis of their depth rank curve for the Turners Falls Formation along the river bed below the dam to demonstrate that the fundamental control of the long-term wet-dry alternations was the 21,000-year precession cycle of the earth. They also recognized Milankovich-type climate-forcing cycles at 39,000 and 116,000 years.

Figures 14 and 15 show detailed measured sections of lakes 0, 2, and 3 (Meriney, 1988). Lake 2 is mostly covered and can be seen only a lowest river levels. All four lakes held alkaline water with substantial dissolved Ca, Mg, and bicabonate ions as evidenced by dolomite laminae and nodules in the mudstones and albite and dolomite cements in the sandstones.

Lake 0. A transition along strike of the same beds from red near the water to gray and black onshore is interpreted as the result of removal of the original limonite stains on the detrital particles as ferrous-iron organic complexes in the bottom waters of the lake. Poor tracks of Grallator (Anchisauripus) occur in the gray siltstones (Olsen et al., 1989). In the lacustrine sequence is a sandstone body inferred to have been constructed as a fan-delta into the lake. Progradation of the delta into the lake is shown by the observation that the sandstone layers at 2.7 m in the measured section overlap the underlying lakebeds to the south and southwest. This fan-delta contains a cross-bed set of sandstone of the lower flow regime with unusually fine preservation of the crests of the bedform.

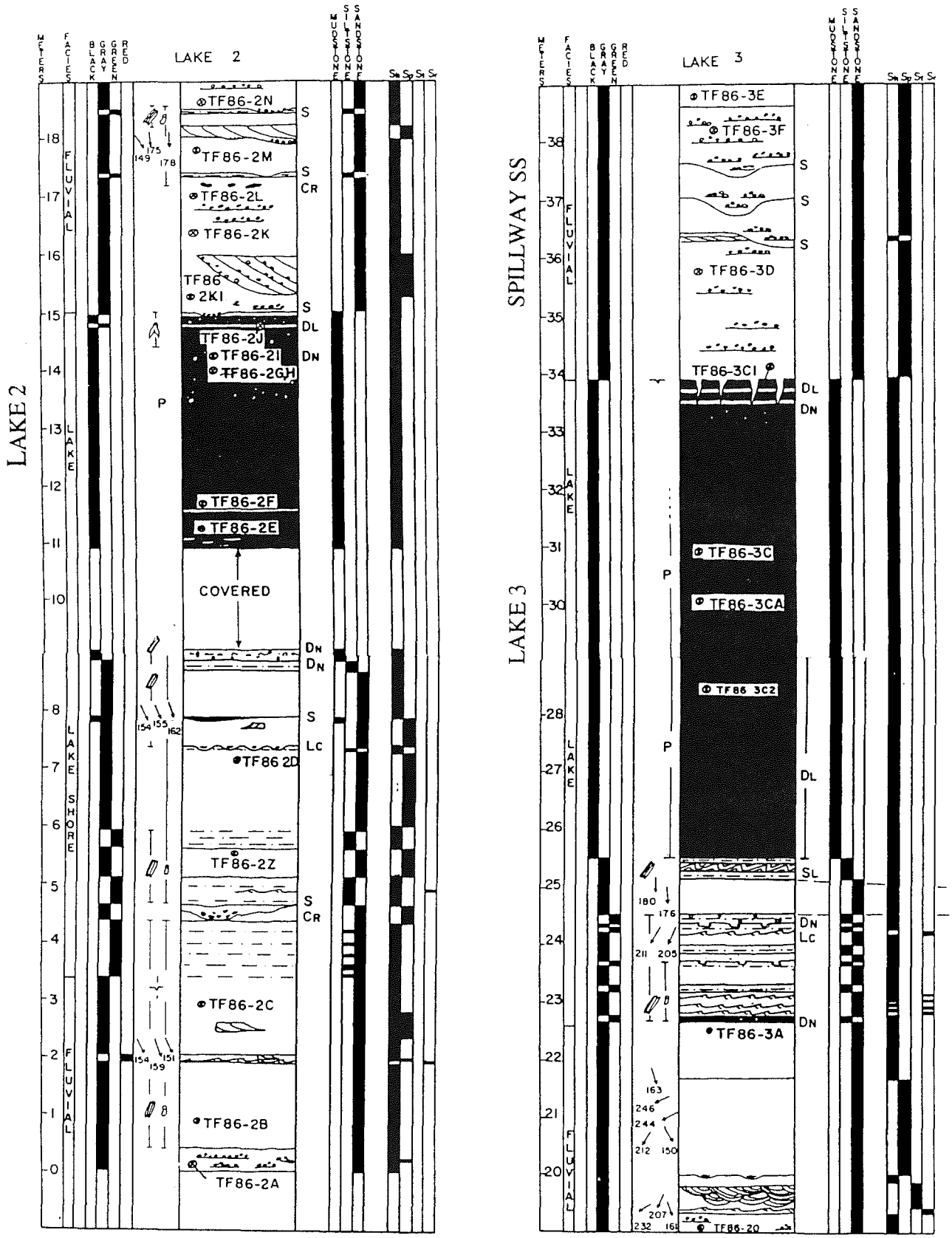
The lake beds show tectonic and/or gravity disruption with many local bedding-plane faults and minor folds. The base of the sandstone at river level is cut by a 1-m displacement, sinistral strike-slip fault. The fault, typical of the family of late, NE-SW compressional structures, has well developed polish, grooving, chlorite mineralization, splays, and local breccia. Lithification was complete enough at the time of motion to truncate quatrtz pebbles in the sandstone. Units just above this horizon have minor folds, possibly of tectonic origin, with associated incipient cleavage that produced minor lineations on bedding planes.

Approaching the bridge, the sandstones become more fluvial with some conglomerate-filled channels. Locally bedding surfaces show parting-step lineations and ripple crests as well as tectonic lineations formed by faint, spaced cleavages (NNE-SSW compression). In general, sediment transport directions for the playa, lacustrine and fluvial strata at Stops 6C and D were to the southwest, south, and southeast (Figures 14, 15).

Lake 1. Located just upstream of the bridge, this unit is mostly covered by float blocks and unavailable for examination except at extreme low water conditions.

Lakes 2 and 3. Six black mudstone samples from lakes 2 and 3 contain 0.7-1.7 wt % total organic carbon (Meriney, 1988; Taylor, 1991). Vitrinite reflectance values are 1.46-1.73 %, above the approximately maximum 1.3 % value for liquid hdrocarbon generation. Furthermore, the kerogen is gas prone, comprising woody and other lignin-rich material of terrestrial origin. These black mudstones are thermally in the gas and condesate zone or possibly above (Taylor, 1991). Temperatures were at least 175 degrees C long enough to thermally crack most of the kerogen (Taylor, 1991).

The lacustrine beds contain 1-cm thick, sill-like, fibrous calcite veins. These veins indicate that fluid overpressure in the compacting lake beds was equal to or very slightly in excess of the weight of the overlying load, lifting the overlying strata slowly to permit the calcite fibers to grow. Bedding plane motion to the SE, probably under the influence of gravity, was associated with this overpressure as indicated by 1) the slightly oblique SE tilt of the fibers with respect to the plane of the veins and 2) SE-directed bedding plane faults and slickenlines are associated with the veins. The best examples of these features are in the strata of lake 4 at the opposite end of the dam, a location not included in this trip because of time and difficulty of access but described by Wise, 1988 (his Figure 20). The overpressure / gravity slide



Measured section of lacustrine beds 2 and 3 in the Turners Falls Formation at the Turners Falls dam (Meriney, 1988). Symbols are the same as in the legend for lacustrine bed 0.

Figure 15. Data from Stop 6D.

systems in lacustrine strata 3 and 4 are ideal examples of the Rubey and Hubbert fault mechanics system, the famous "beer can experiment."

No bitumen occurs in these fibrous calcite veins, but younger calcite/quartz veins with a N30E strike do have bitumen imbedded in them and many younger druzy quartz veins of the same orientation include small masses of bitumen. Because hydrocarbon generation from these organic-rich lacustrine muds required at least 5-10 milion years, considerable time elapsed between lacustrine deposition and formation of at least some of the N30E veins. In the vicinity of the dam, these veins are closely associated with the major normal fault system. Wise (in press, 1992) on the basis of dip-domain analysis of the Hartford and Deerfield basins argues that the early faults in both basins were an irregular and disconnected system. These faults only became integrated into a single border fault late in the history of each basin in association with the major regional tilting of the entire basin and its contents. The generation of hydrocarbons and their migration into veins in the Deerfield basin help constrain the timing of the veins as at least 5-10 million years after deposition. Many NE-SW compressional strike-slip faults cut the calcite veins with bitumen in the vicinity of lakes 2 and 3, making it appear that the bulk of hydrocarbon migration had been completed before this time.

The black mudstone of lake 3 contains the "Turners Falls fishbed" where fish fragments and complete skeletons were preserved in the anoxic bottom waters of the lake. Until recently the holostean Semionotus was the only genus recovered from the bed, but McDonald (1991) has reported the presence of the advanced chondrostean Redfieldius and the coelacanth Diplurus. Semionotus also occurs in the strata of lakes 0 and 2 (Olsen et al., 1989). Reptile footprints occur in some of the gray sandstones and siltstones of lakes 1, 2, and 3 (Olsen, et al., 1989).

The top of lake 3 contains mudcracks 40 cm deep in the black mudstone (Figure 15). This lake abruptly drained, exposing the black muds on the lake floor which then desiccated to deep mudcracks. As the muds dried, dolomite precipitated as concretions that line the mudcracks and as septarian nodules. Subsequently, a river flowed over the dry lakebed, filling the mudcracks with pebbly sand and depositing a thick body of cross-bedded conglomeratic sandstone.

The effect of dewatering of the black muds is seen in the gray colors of the fluvial sandstones that overlie lakes 2 and 3 (Figure 15). These fluvial sands presumably were deposited with the ordinary limonite surface stains on the grains that with time dehydrate to hematite as has happened to produce the typical pale red colors of many other fluvial sandstones in the Turners Falls Formation. Here, in place of this usual sequence, the limonite stains were removed as ferrous-iron organic complexes as the water flush from the compacting underlying black muds passed through the fluvial sands, leaving them without this method of hematite production.

13.3 Return to parking lot. Proceed east on Route 2.
13.5 Through light and then immediate right turn onto Bridge Street. Follow water's edge.
13.7 Abutment of 1878 suspension bridge across Connecticut River. On far side is a small park with large conglomerate blocks containing armored mud balls. You might wish to visit these at some other time. Continue along this road and through old massacare graounds, without taking any turnoffs.
14.0 Stop sign for Route 2. Turn right (east).
14.6 Again, pass entrance to Barton Cove.
15.1 Road cuts parallel to strike of the Barton Cove units, here redbeds.
16.1 Start of outcrops for Stop 7, probably a "rolling stop." Best outcrops are on the north side of the road. Parking is prohibited on shoulder of road. Proceed 100 m and park in the lot of Chases Country Store and A J Cycle Shop.

STOP 7. MOUNT PISGAH ALLUVIAL-FAN CONGLOMERATE (BARTON COVE CORRELATIVE) ON ROUTE 2.

These conglomeratic outcrops are essentially on strike (Figure 7) with the lake units at Barton Cove, 2 km to the west. Mount Pisgah is a hill just to the north, designated by Handy (1976) as the type locality of one of three fans of the Mount Toby Conglomerate. These fans were mapped using paleocurrent dispersal

patterns and decrease in size of the gravel in the downfan directions (Wessel, 1969; Handy, 1976). The Mount Pisgah fan also is distinquished from the Taylor Hill and Mount Toby fans to the south by more brown tournmaline and zoned tourmaline and less clear garnet in the accessory heavy-mineral assemblages (Handy, 1976).

The Mount Pisgah fan built out and interfingered with the Barton Cove lacustrine strata. A few of these lacustrine gray mudstones are interbedded with alluvial-fan coarse clastics on the northern side of the highway at the western end of the outcrop. The fan origin of the conglomerate is indicated by 1) boulders up to 20 cm in diameter, 2) pebble imbrication that dips up fan, 3) dominance of upper flow regime plane beds in the pebbly sandstone, and 4) 10-15% sandstone interbeds with a near absence of mudstone.

The Mount Pisgah fan built out from the N or NNE. The source area for the fan contained many phyllites and low-grade metamorphic rocks similar to those exposed immediately N and NNE of the basin. These lithologies stand in sharp contrast to the high-grade metamorphic units typical of the Pelham Dome just to the east of the border fault. In early Jurassic time, erosion levels in this eastern area would have been structurally much higher than the present level of erosion of the dome and thus could have exposed low-grade metamorphic rocks. However, the pattern of inverted metamorphism (higher grade over lower grade metamorphic rocks) in the nappes of the dome suggests that early Jurassic erosion would would have yielded clasts of higher grade rather than lower grade metamorphic rocks. Thus the sources of the low-grade metamorphic clasts in the conglomerates were more northerly than easterly.

16.8 Continue east on Route 2 across French King Bridge.
17.4 Turn right onto Dorsey Road.
17.8 Park on the right by the first outcrops of Stop 8.

STOP 8. MESOZOIC BORDER FAULT UNDER FRENCH KING BRIDGE.

The French King Bridge of Route 2 spans a narrow gorge of the Connecticut River cut into the Mesozoic Border Fault zone (Figures 16, 17). The western abutment of the bridge is entirely in conglomerate of the Mount Pisgah fan facies of the Lower Jurassic Mount Toby Conglomerate whereas the eastern abutment stands entirely on Paleozoic metamorphic rocks. Glacial and river erosion was concentrated along the line of outcrop of the Ordovician Partridge sulfidic schist, a naturally weak unit further degraded by the fault zone.

The name French King Bridge is derived from French King Rock, a large flat slump block, or erratic of conglomerate, located in the middle of the river 300 m upstream from the bridge. Before construction of the Turners Falls dam , it stood much higher above water level but is still a pleaant picnic spot at low river stages. The name, according to Emerson, (1898, p. 296) derives from the time when ".....a bateaux of the French and Indians during the French wars, were stopped here by the rapids, and one adventurous Frenchman pressed on to this rock and broke a bottle of wine over it, claiming the country in the name of the French King." Subsequently, Mike Rhodes of the UMASS geology group "liberated" the rock by landing on it by kayak to pour (a very small amount of) English ale on the rock and claim it in the name of her majesty, Queen Elizabeth II. Currently negotiations are in progress for the entire Massachusetts Legislature to meet on the rock to take formal possession for the Commonwealth, using Sam Adams beer.

Stop 8 comprises the best known portion of the the border fault in Massachusetts because of excellent outcrops, detailed work by Peter Robinson for the Massachusetts state geologic map, a thesis by Stopen (1988), and extensive drilling by Northeast Utilities for a proposed nuclear power plant. Some of the data for the Northeast Utilities' drill cores are shown in cross section C-C', located about 1 km north of the bridge (Figures 16,17). Note that the fault dips westward at 30-45 degrees and that the fault-related disturbed zone has a thickness of about 200 m. This zone has a complex structure, including mylonitic rocks, possibly produced by Paleozoic thrusting along a southward continuation of the Ammonusuc Fault Zone of New Hampshire and/or as a deeper, higher grade part of the early Mesozoic extensile movements. Stopen's work on the S and C fabric of the mylonites shows entirely normal motion, suggestive of the early, deeper phases of Mesozoic extension. The mylonites are thoroughly rebroken; vein quartz has been injected, broken, silicified, and rebroken within the zone.

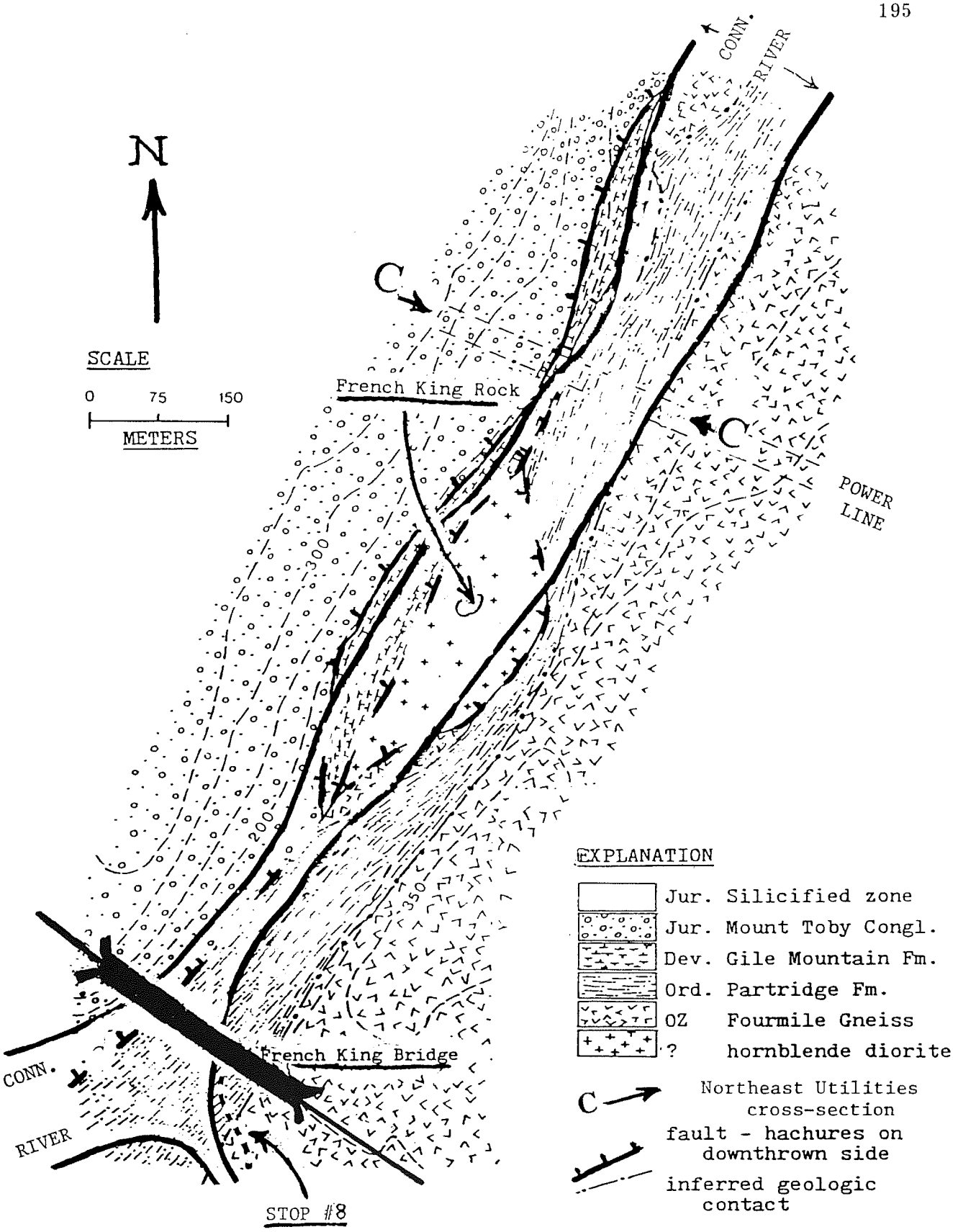
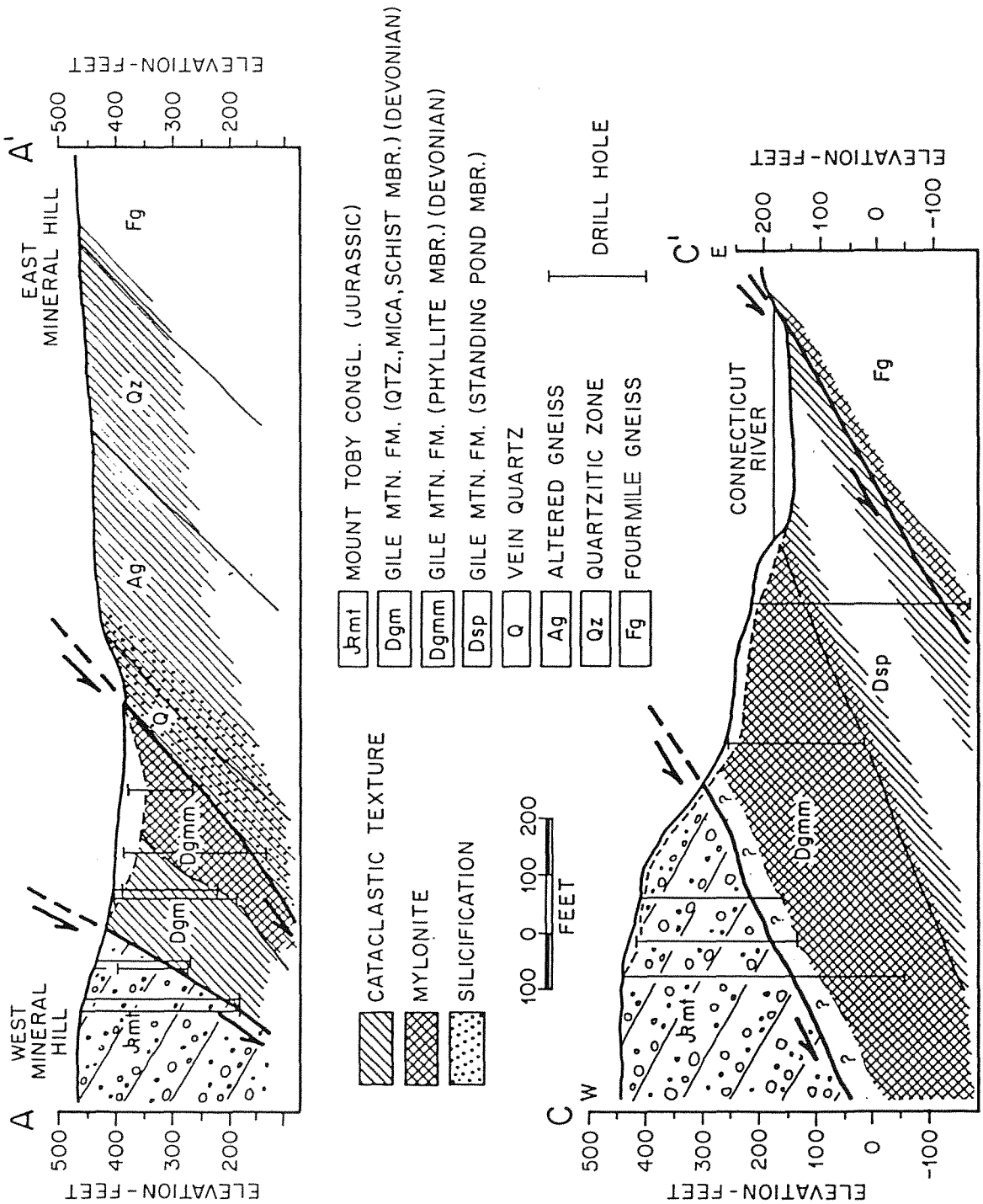


Figure 16. Geologic map of the French King Bridge Area, Stop 8.
(Extensively modified from Robinson, unpublished, and Stopen, 1988).



Figrue 17. Eastern Border Fault near French King Bridge. Section C-C' is along the power line 700 meters upstream from the bridge. Section A-A' is at Mineral Hill, about 2 km S of this location. Sections modified from drill data of Northeast Utilities Public Safety Analysis Report for possible Montague nuclear power plant, 1975.

The Northeast Utilities' documents argue that all these structures were Paleozoic in age and that the contact between the Paleozoic and Mesozoic rocks is an unconformity even though the Mesozoic rocks intersect the map trace of the inferred unconformity at a high angle. On the basis of an unconformity being present, they argue that the border fault has been inactive since the Paleozoic and that the then proposed (1975) nuclear reactor need not be engineered for a higher level of seismic risk. Their interpretation has fallen far short of widespread acceptance by the geologic community!

The unpublished geologic maps of the area by Robinson (1985) and the Stopen (1988) show a 150-m wide band of Ordovician Partridge Schist largely beneath the Connecticut River. At its eastern contact, the schist rests on the west-dipping Fourmile Gneiss of the Pelham Dome. A number of horse slices (a herd ?) of younger Paleozoic rocks lie along the western shore, just upstream of the bridge. In this area, a slice of Devonian Gile Mountain Formation is immediately adjacent to the Mesozoic rocks and is separated locally from the Partridge Schist by a zone of Mesozoic silicification or by a slice of hornblende diorite of uncertain age.

The outcrops we will visit are along Dorsey Road - Meadow Road which circles to pass beneath the eastern abutment of the bridge. Most of the roadcut SE of the bridge shows the well foliated, amphibole-rich Fourmile Gneiss, possibly the roots of an early Paleozoic island arc. These rocks are part of the late Precambrian to early Paloezoic basement in the line of Bronson Hill mantled gneiss domes of Devonian age. The western dips in the gneiss are part of the western flank of the Pelham Dome. Similar western dips along the line of domes provided the strength anisotropy that controlled much of the local strike and dip of the border fault zone in Massachusetts and Connecticut, a clear case of tectonic heredity.

Immediately under the bridge are outcrops of the Partridge Schist, a widespread marker horizon in New England. Organic-rich muds for the schist formed under pervasive euxinic conditions just as the Taconian Orogeny was getting underway. Farther south, the unit becomes the Brimfield Schist and farther west it correlates with the Walloomsac black phyllites.

As might be expected in close proximity to a major fault zone, many of the joint surfaces in the roadcuts are slickenlined, polished, or have mullion structures on them. For those with more time available, good exposures of the silicified rocks occur on East Mineral Hill, just east of a road of the same name 1 km south of this stop (see cross section A-A' on Figure 17).

17.9 Continue north beneath the bridge.
18.2 French King Rock visible in the river at low water stages. Note the upper part of the hill across the river is Mesozoic conglomerate whereas the lower parts of the hills are Paleozoic rocks (Figures 16 and 17).
18.4 Pass under power line and across section C-C' of Figure 17.
19.4 Minor cross-roads at intake area for Northfield Mountain Pump Storrage Facility. Parking area for picnics at left.Turn right under railroad bridge.
19.5 Stop sign, intersection with route 63. Entrance to Northfield Mountain Visitors Center is ahead and to the right.Turn right onto route 63. END OF TRIP LOG.

Route 63 can be followed south 2 miles to Route 2 where you can go E and W or you can follow Route 63 south 18 miles through Millers Falls into North Amherst. Proceed straight through the light in North Amherst for one mile to the UMASS campus.

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HYDROGEOLOGY AND WATER RESOURCES OF THE CONNECTICUT VALLEY AND WESTERN QUABBIN RESERVOIR WATERSHED

by

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INTRODUCTION

The Connecticut Valley area of western Massachusetts contains an extraordinarily wide variety of water resources, from buried river channels and deltas to the catchment of the Quabbin Reservoir. Most of the local communities rely upon groundwater in the surficial deposits within the Connecticut River Valley for the bulk of their water supply, and this has often caused some problems with the agricultural use of much of the flood-plain area. Glacial Lake Hitchcock left behind a legacy of deltas, kame terraces and buried gravel which serve as major water sources. Finally, the eastern margin of the Valley is in the catchment of the Quabbin Reservoir, which supplies water for the 2.5 million residents of the Boston metropolitan area. We will look at these various characteristics during this excursion (Fig. 1).

CONNECTICUT VALLEY OVERVIEW

The Connecticut Valley Lowland region lies within a down-faulted basin of Triassic/Jurassic age bordered on the east and west by older crystalline rocks. The Triassic/Jurassic rocks in the Lowland consist chiefly of arkosic sandstone and other sedimentary rocks interlayered with basaltic lava flows. In Whately, the principal Triassic unit is the Sugarloaf Arkose, named for the mountain in South Deerfield (Fig. 1). Throughout the Lowland Region, pre-glacial streams incised deep channels into the relatively soft sedimentary rocks. In some cases these channels correspond closely to present stream courses; in other cases they relate to a pre-glacial drainage system totally different from that which exists at present.

During the most recent Pleistocene glaciation, southward drainage was dammed by natural deposits in Rocky Hill, Connecticut, and virtually the entire Connecticut Valley Lowland region became a glacial lake. This lake, known as glacial Lake Hitchcock, stood at an elevation that now corresponds to approximately 300 feet above sea level in Whately. Its strand line is tilted upward to the north due to isostatic crustal rebound. Large quantities of sand, silt, and clay accumulated in the lake during deglaciation, which accounts for the very flat surface topography of the Lowland.

The following sequence of events led to the development of separate aquifers and interaquifer confining layers. During glaciation a layer of till was deposited directly over the bedrock; this layer was probably thicker in the incised bedrock channels and thinner over the higher portions of the bedrock surface. During deglaciation large quantities of meltwater were discharged from the wasting ice masses. Some of the meltwater streams occupied the pre-glacial channels and cut rapidly through the previously deposited till. In some cases, till was completely eroded and the pre-glacial channels became filled with glaciofluvial sand and gravel. This sand and gravel comprises the "lower aquifer." Following the damming of drainage in Connecticut, the till and glaciofluvial deposits occupying the channels were blanketed with silty fine sand which graded upward into varved clay. After several thousand years of fine sediment deposition, the dam was breached and subareal (as opposed to subaqueous) deposition processes again predominated. Sand deposited into the rapidly shallowing lake by anastomosing streams left a relatively thin layer of sand on top of the old lake bottom. These sand deposits form the "upper aquifer."

From the vantage point of Mount Sugarloaf (**STOP 1**) we can see the agricultural land in Whately, where pesticide application to sandy soils resulted in ground water contamination. In Whately, for several miles south and southwest of this stop, private wells in the shallow upper aquifer were contaminated. In South Deerfield, almost immediately southeast of this stop, a public well field in the upper aquifer was contaminated. For Whately, a replacement water supply was sought in the

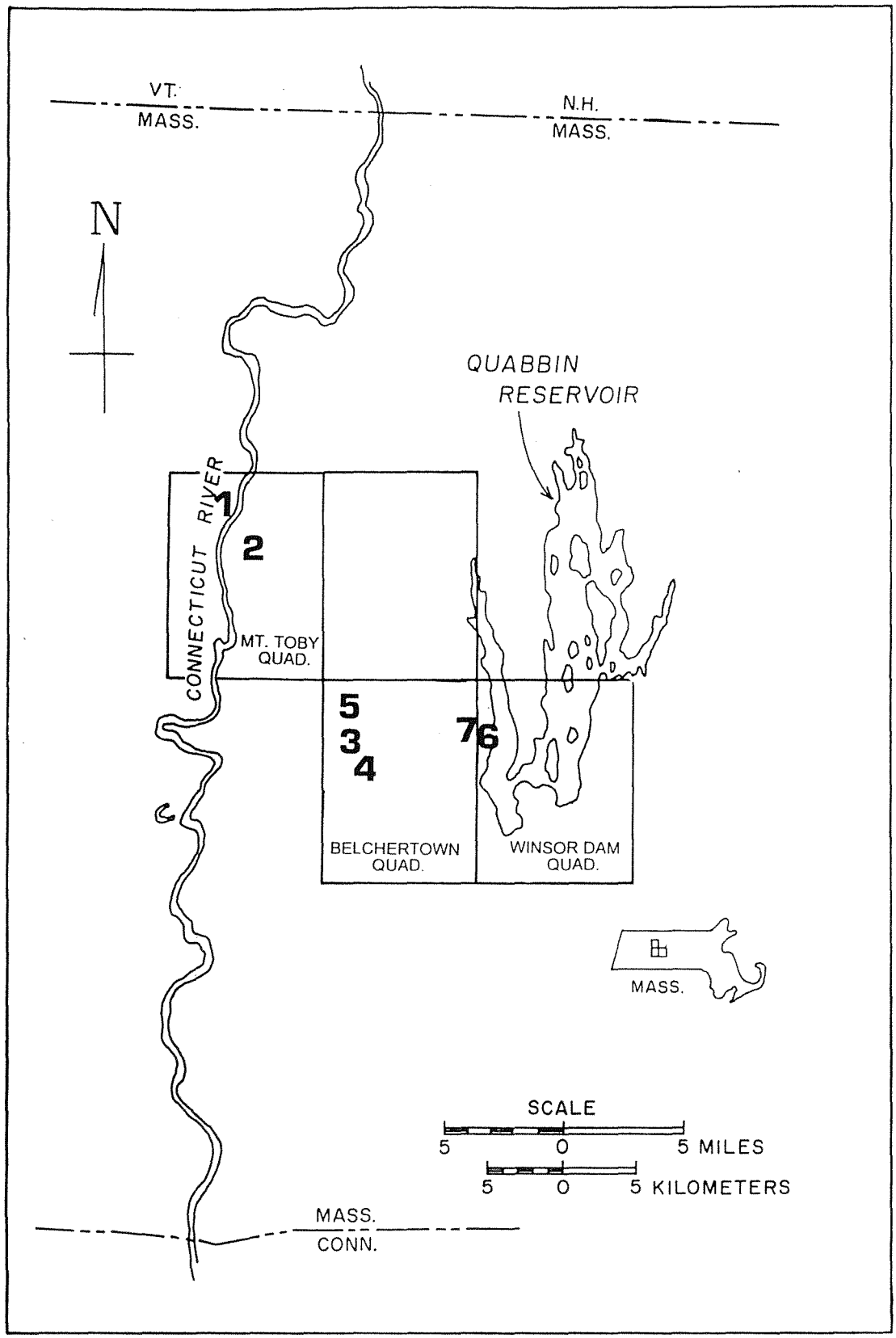

Figure 1. Location map showing field trip stops within USGS 7.5' quadrangles: 1. Mount Sugarloaf; 2. Sunderland Delta and salmon hatchery; 3. Lawrence Swamp overlook; 4. Amherst Town Well #6; 5. Old and new Amherst landfills; 6. Mouth of Cadwell Creek at Quabbin Reservoir; 7. Monitoring site in Cadwell Creek watershed.

deep, lower aquifer, which is protected from surface contamination by a thick aquitard consisting of clay (Heeley and others, 1986).

Analysis of existing data led to the conclusion that a pre-glacial channel probably was cut into the Sugarloaf Arkose in close proximity to the present course of the Mill River, a minor tributary of the Connecticut River. The Deerfield River, a major tributary of the Connecticut, may have used this channel prior to the formation of Lake Hitchcock, but we refer to it as the pre-glacial Mill River channel due to its close proximity in Whately to the present stream of that name.

Seismic refraction lines were positioned along and across the presumed valley trend in order to test the hypothesis of a buried Mill River channel. A municipal water supply well in the town of Hatfield to the south derives water from an aquifer believed to occupy a continuation of the same buried channel that passes through Whately. Resistivity soundings were used as a means of establishing the geoelectric "signature" of the aquifer and testing the hypothesis that it extends northward into Whately.

The geophysical survey program conducted in Whately consisted of surface and borehole techniques designed to accomplish the following objectives: (1) Define the trend of the pre- glacial Mill River bedrock channel and locate the deepest areas; (2) Evaluate channel sediments at potential test well sites; (3) Identify potential high yield zones in test borings prior to setting well screens.

The layout of seismic refraction profiles, electrical resistivity soundings around the buried valley aquifer system (lower aquifer) is shown in Figure 2. The Mt. Sugarloaf stop is located about 6000 feet east of the northeast corner of each figure.

A water supply production well was drilled to a depth of 217 feet at the location of W-7 (Fig. 2). This well now provides water for Whately through a distribution system covering the eastern part of the town. Recharge to this aquifer may be supplied by upward leakage from the underlying arkosic bedrock as well as from southward migration down the buried valley aquifer, which extends northward to a glacial lake delta in Deerfield, similar to the Sunderland Delta which is the subject of the next stop.

HYDROGEOLOGY OF SUNDERLAND DELTA

The Sunderland Delta (**STOPS 2A and 2B**) is an example of the occurrence of ground water in one of the deltas formed in Lake Hitchcock. These deltas (including the Sunderland, Montague, and Chicopee) influence ground-water occurrence and movement for two reasons: first, the highly permeable soils (Hinkley and associated soils) on the delta result in high amounts and rates of recharge to the underlying aquifers; and secondly, the coarse sands, very coarse sands, and gravels in the deeper parts of the deltas form productive aquifers. The aquifers may occur in buried valleys (as the case of the Montague Delta) or occur in deeper coarse sediments deposited as forset beds and/or pre-lake fluvial sand and gravel (as the case of the Sunderland Delta). The Sunderland town well is screened in the latter deposits (Fig. 3). As shown in the figure, the delta is bounded by negative boundaries from the bedrock valley walls and the interfingering of delta sediments with poorly permeable lake clays, and by a positive boundary from Long Plain Brook which has a constant and steady flow of the brook most of the year. The lake clays form a down-gradient barrier or "subsurface dam" forcing water to the surface as the springs, ponds, and streams of the extensive hatchery discharge area. All the surface and ground-water discharge passes through a culvert in the eastern part of the hatchery area and has been monitored continuously for more than 35 years. A flow net was constructed for the southern area of Sunderland Delta (Fig. 3) and with the discharge data transmissivity of approximately 14,000 gpd/ft was calculated for the area--a value that is close to that calculated from the Sunderland Well pumping information. The total discharge from the aquifer at the culvert has also allowed calculation of the quality stressing from a proposed housing development of 60 units in Puffer Quarry (Fig. 3). The long-term drought and seasonal flow as monitored from 1960-65 (the most severe drought conditions for at least 150 years) was approximately 450 gpm (gallons per minute). Using this conservative value, calculations were made of contamination build-up from septic tanks, road salts, and lawn applications. This contaminant build-up was negligible; however, sediment from construction activities could cause serious hatchery problems unless it was properly disposed and contained in sediment catching depressions and traps. Other aspects of the hydrogeology of the Sunderland Delta are discussed by Motts (1979) and McElroy (1987).

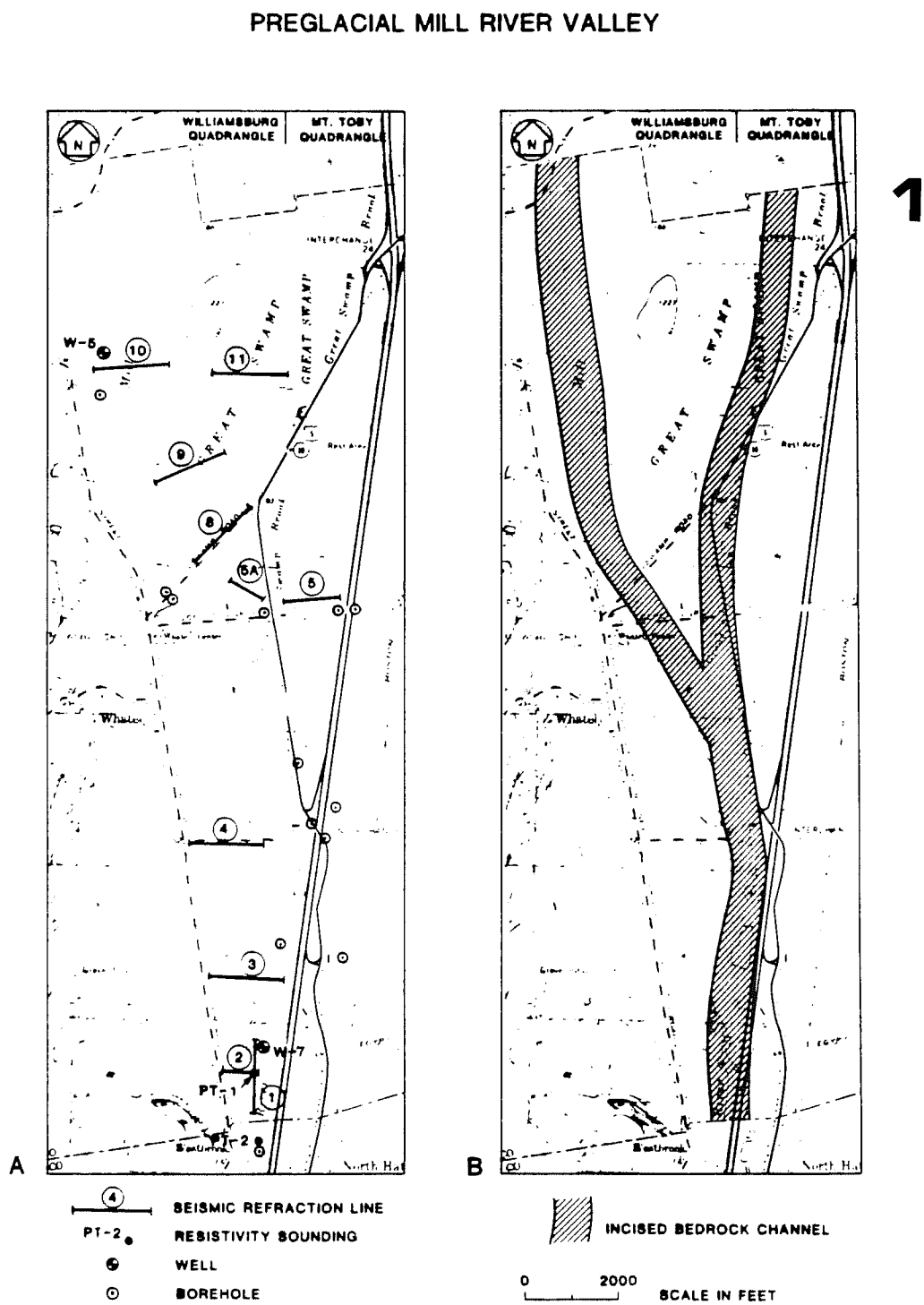

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Figure 2. Preglacial Mill River valley (from Heeley and others, 1986). A. Location of borehole and geophysical data. B. Location of incised bedrock channel. Location of **STOP 1** is just to the east of the upper part of the figures, relative position indicated next to fig 2B.

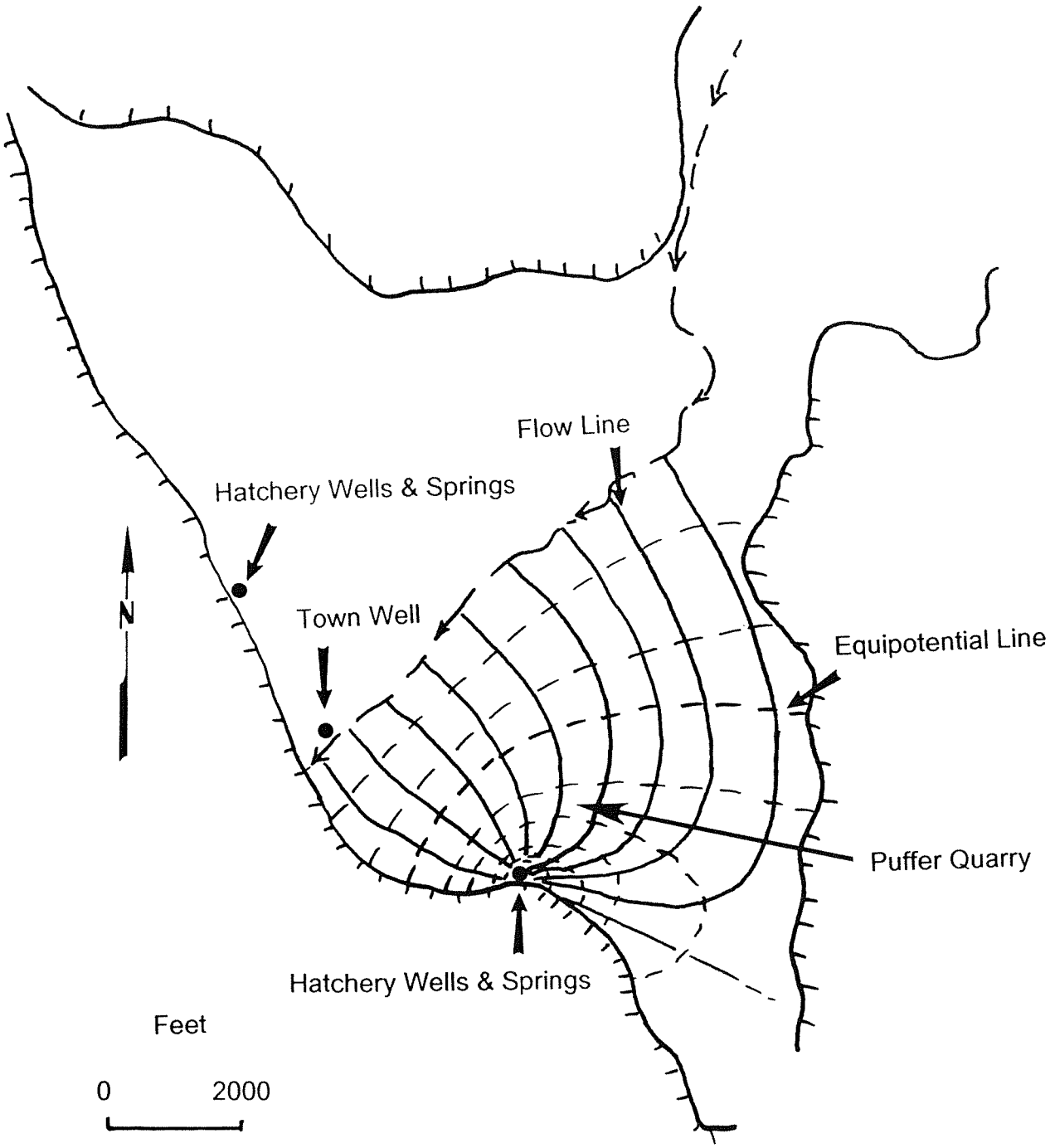

Figure 3. Outline of Sunderland Delta (**STOP 2**) showing discharge areas, Puffer Quarry, flow net, and positive and negative boundaries. Positive boundary is a stream (Long Plain Brook) and negative boundaries are bedrock walls and interfingering of delta sediments with lake clays shown by slanted lines.

THE LAWRENCE SWAMP AQUIFER

Lawrence Swamp is an important environmental and recreational asset because it is one of the largest, most distinctive and diversified wetlands in western Massachusetts. It is an important economic resource because at present the deeper stratified drift aquifer underlying the swamp supplies approximately 2 mgd (million gallons a day) to Amherst's wells, but the aquifer has the capacity to yield greater than 5 mgd. Lawrence Swamp is the discharge area for Lawrence Swamp Basin. The basin, drained by Hop Brook, covers an area of approximately 15.5 square miles. From the surface downward, Lawrence Swamp is underlain by (1) post-lake wind-blown sands with some dunes and alluvial sand and gravel forming the shallow unconfined aquifer; (2) Lake Hitchcock clays and silts separated in places by beds of sand; (3) the productive deeper stratified drift aquifer formed of sands and gravels; and (4) the bedrock aquifer consisting of Triassic rocks in the southern part of the swamp and crystalline bedrock of Ordovician through Devonian age in the northern part. The recharge area of the basin occupies the upland which flanks the swamp and consists of kame terraces and deltas of coarse-grained stratified drift which dip below the lake beds in the swamp to form the productive artesian aquifer supplying Amherst's wells. The stratified drift is called the primary recharge area because Hinkley and associated soils formed on the drift have high recharge rates from 60% to 70% of total precipitation. The secondary recharge area is underlain by crystalline bedrock and till and has recharge rates ranging from 15% to 30% of precipitation. There was relatively little town interest or concern for Amherst's water supplies and aquifers in the 1960s and early 1970s. However, the placement of a landfill in the northern part of the aquifer recharge area and the proposal of a major development with 200 units in the same area fostered citizens' interest and concern for their water resources. As a consequence of this interest, the following ground water studies have been made: Heeley, (1973); Motts (1974, 1975); Geraghty and Miller (1979,1985); Tighe & Bond, Inc. (1991); and most recently, an aquifer modeling study by Exarhoulakos and Heeley (1992).

MODFLOW Modeling of Lawrence Swamp Aquifer

At **STOP 3** we are looking down the steep northern rim of the Lawrence Swamp Basin, across the predominantly wooded swamp. Directly ahead beyond the swamp is Long Mountain in the Holyoke Range, which forms the southern basin boundary. Problems associated with ground water in this basin are illustrated by some of the physical features observed here. The steep slopes and shallow saturated thickness in the recharge areas surrounding the swamp created problems in the MODFLOW model simulation. The Triassic border fault passes just to the north of this stop. The Holyoke Range is also cross-cut by numerous faults. Problems associated with geologic features simulated in the modeling are discussed below (Exarhoulakos, Heeley, and Motts, 1992).

Determining the downgradient stagnation area of a municipal well field's zone of contribution in an artesian aquifer was complicated by basin margin geology and induced fracture recharge in the Lawrence Swamp Basin, Amherst and Belchertown, Massachusetts. Most recharge occurs along the basin margins where saturated thickness is insufficient and too discontinuous to be modeled, and where data is minimal due to a lack of exploratory drilling. Dry nodes appear in these areas during model development, thereby effectively removing recharge from the model. Recharge was simulated as wells injecting directly into the artesian aquifer.

On the southern and eastern basin flanks, faults which contribute significant recharge to the aquifer were modeled as general head boundaries, enabling validation of the MODFLOW model to observed stressed conditions. Water flow into and out of the modeled area was further complicated by two conditions. The impact of a partial stream diversion near the eastern basin boundary was established by a sensitivity analysis in which varying percentages of the stream flow were simulated as recharge to the aquifer. To account for outflow through a small channel incised into bedrock at the southwestern basin boundary, recharge was decreased.

Figure 4 illustrates the final delineation of Zones II and III for the municipal wells (Exarhoulakos and Heeley, 1992). Zone II is the area of stratified drift which supplies water for the wells under the most severe pumping and recharge conditions which can be realistically anticipated, defined as full safe yield pumping and 180 days with zero recharge. The downgradient limit of Zone II extends within 3000 feet of the surface water basin divide across the entire basin. Zone III consists of till and bedrock areas which contribute flow to adjacent Zone II areas. The modeled area is shown within the central Zone II area. However, most of the modeled area is actually a discharge area and is underlain by silt and clay beneath a shallow sand aquifer. The primary sand and gravel recharge

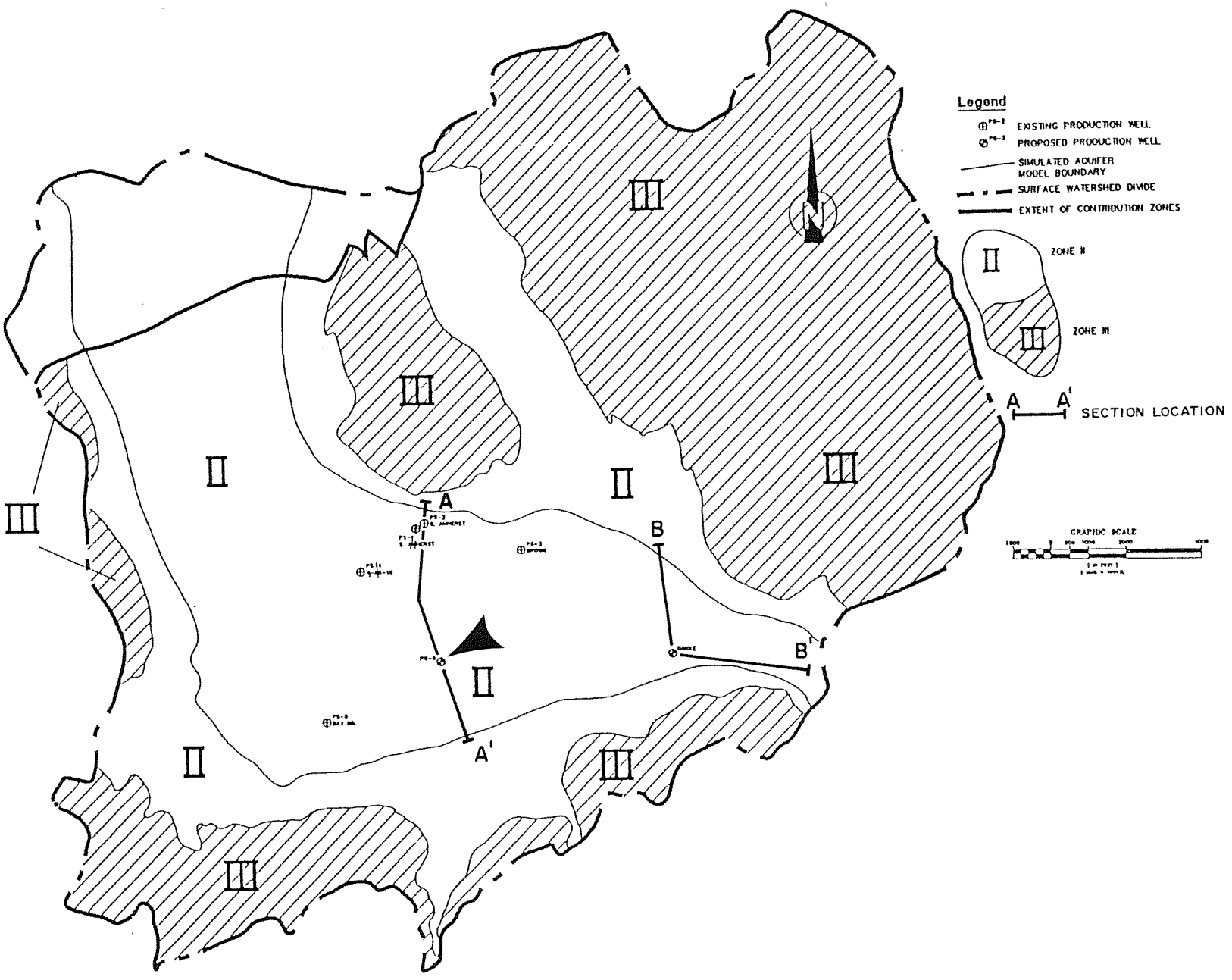

Figure 4. Site plan of Lawrence Swamp Basin, Hop Brook watershed. Location of **STOP 4** (Amherst Town Well #6) indicated by arrow. (adapted from Exharloukas and Heeley, 1992).

areas hydraulically connected to the lower artesian aquifer are primarily located between the modeled area boundary and the Zone III.

The Pump Test at Well No. 6

Well No.6 (**STOP 4**) is one of four large capacity (> 1.3 MGD), production wells which simultaneously pumped from the confined, Lawrence Swamp aquifer at a combined rate of 5.5 mgd, during a ten (10) day continuous aquifer pumping test during June, 1991. The test was conducted as part of the State regulated New Source Approval Process for the development of two (2) new large capacity wells in the Lawrence Swamp Aquifer. The two wells, Amherst well No.6 and Belchertown well PS #1, located approximately 6000 feet apart, each have proposed yields of 1.3 to 1.4 MGD.

The goal of the study was to gather empirical data from the aquifer under stressed conditions. These data would be used to calculate aquifer parameters of transmissivity (T) and storativity (S) and in the validation of the MODFLOW (numerical) analysis. The MODFLOW analysis was used to predict a combined "safe yield" for the basin by determining the Zone of Contribution (Zone II) for all wells pumping simultaneously. The Zone II could not include the landfills located downgradient of the production wells. In addition, the potential impacts on the irreplaceable Lawrence Swamp wetland resources were to be addressed.

Currently there are 5 supply wells located in Lawrence Swamp with a combined, approved yield of 4.4 mgd. Due to the presence of potential sources of contamination within the watershed (landfills and suburban development -- **STOP 5**), the State authority required that the test commence under "static" conditions and be conducted under the maximum stresses which could occur during actual production. However, due to design limitations of the existing production wells and distribution system, only 2 of the 5 existing supply wells were pumped simultaneously with the 2 test wells.

The expected complexities of conducting a pumping test and collecting and analyzing valid data for calculation of aquifer parameters, were therefore compounded by the need to adequately supply the Town with water during the test and to monitor for downgradient flow reversal near potential sources of contamination; by the number of monitoring locations (40) and the size of the basin (3.5 sq.mi.); and by pumping multiple wells (2 existing supply wells and 2 test wells) with overlapping cones of influence.

Stratigraphy was logged for all new wells drilled for this study and logs of existing wells were viewed (Fig. 5). Wells were completed in the confining clay unit as well as in the water-table and confined aquifers. Surface water staff gauges were installed in adjacent brooks and ponds and a lake and one well point was installed in the lake sediments. Locations and elevations of all water level measuring points were surveyed. Climatological data were recorded prior to the test to establish static conditions and continued during the pumping and recovery portions of the test. Water level data were collected electronically with data loggers employing pressure transducers at critical locations and manually at remote locations prior to and throughout the duration of the test. These data were collected so that potential effects onthe recorded water level elevations caused by aquifer barometric efficiency, partial penetration of the wells, artificial recharge from the test well discharge or precipitation,and interference from overlapping cones of depression could be evaluated and data adjusted as appropriate.

The existing supply wells were turned on five days prior to the actual test to create a "pseudo-static" condition. Drawdown effects related to the existing supply wells and the effects of the overlapping cones of drawdown between wells No.6 and PS #1 were calculated and found to be insignificant during the early times of the test; data were corrected for later effects. Effects of barometric changes, partial penetration and recharge also were determined to be insignificant. These early data were used to calculate aquifer parameters of T and S for various locations throughout the aquifer using the Theis and modified Theis solution.

The calculated values of T and S were in good agreement with previously estimated and calculated values. When used to predict short term drawdowns at various locations, the predictions generally agree with actual field conditions as measured during the pumping test. Therefore, althoughthe ideal conditions under which the analytical techniques of Theis, for calculating aquifer parameters, were not met during this test, the corrected data could be used to develop reasonably reliable values of T and S.

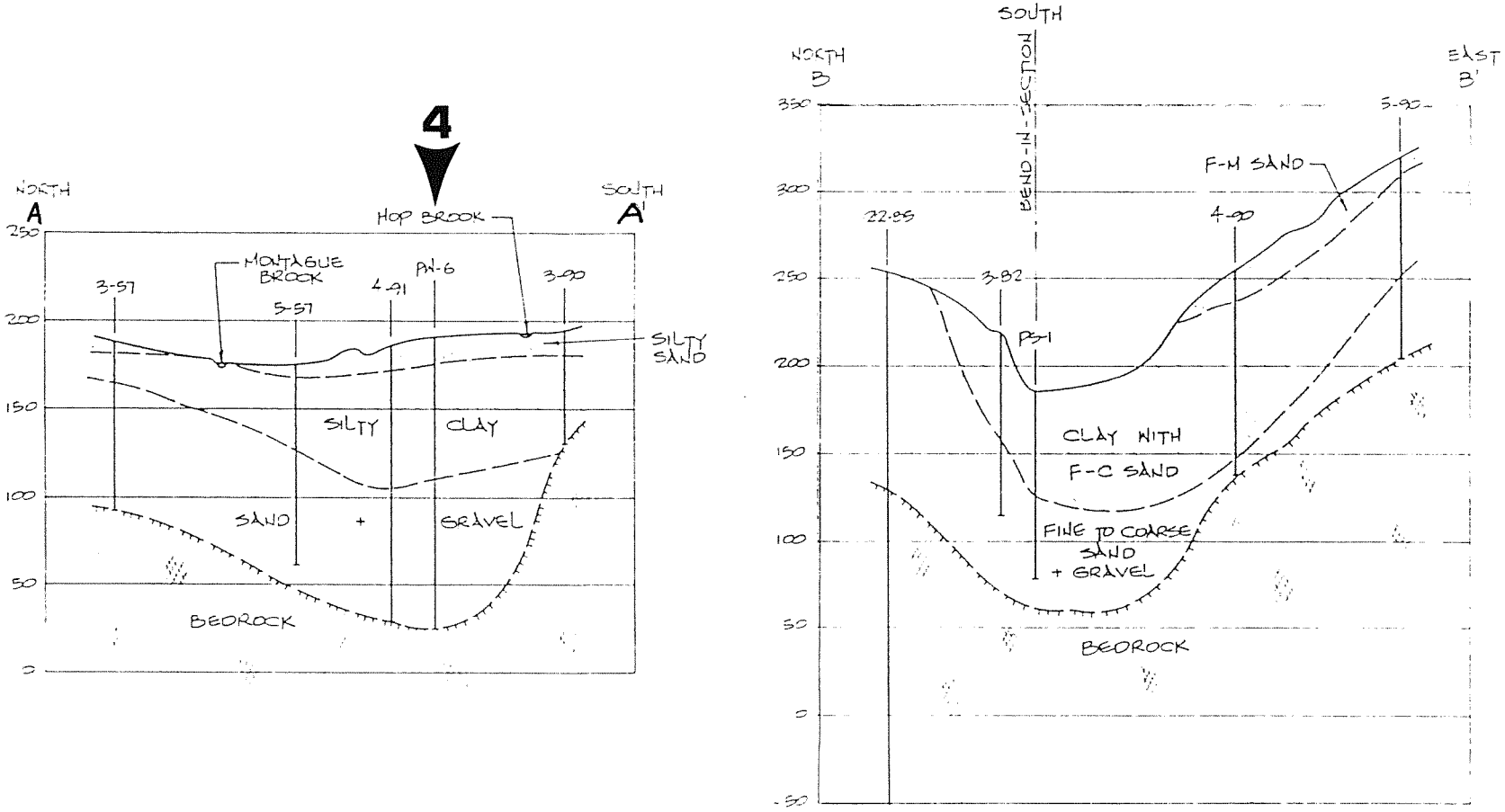

Figure 5. Cross-sections through the buried valley aquifer in the Lawrence Swamp basin. Elevations are in feet. Line of section A-A' and B-B' are shown on Figure 4; **STOP 4** (Town of Amherst Well #6) is indicated by arrow.

THE QUABBIN RESERVOIR

The Quabbin Reservoir, the largest and arguably the most important drinking water reservoir in the northeast, is located in the central highlands of Massachusetts (Fig. 1). The land around the reservoir, encompassing some 482 km^2 in area (186 square miles) is largely a protected catchment, managed as a forested watershed and recreational resource by the Metropolitan District Commission (MDC), although there is some residential development towards the boundaries of the drainage basin. Over the past ten years, there was a great deal of concern that the quality of water within the reservoir would be compromised by acid deposition because there had been an apparent decline in the alkalinity of the water since the reservoir was filled in the early 1940's. This was accompanied by problems in maintaining a viable trout and smelt fishery, since egg and fry mortalities seemed to be on the rise over time (Godfrey, 1988). Moreover, the geological terrain in central Massachusetts underlying the reservoir has often been classified as "sensitive to acidification," due to the presumably low reactivity of the silicate bedrock (DEQE, 1984). It seemed imperative to obtain a more precise understanding of the hydrology and aqueous geochemistry of the Quabbin Reservoir drainage basin.

We selected two small stream watershed systems for intensive study to evaluate their hydrology and hydrochemistry and to provide a first-order analysis of the likelihood of the reservoir to be threatened by watershed acidification. These stream watersheds are representative of the various types of bedrock geology and surficial deposits that occur in various parts of the Quabbin catchment (Leonard and othersl, 1984; Yuretich and others, 1986). The Carter and Mundberry Brook watershed was chosen as the representative of the eastern side of the reservoir in Petersham; the Cadwell Creek watershed was selected as a typical site on the western side in Pelham (**STOPS 6 and 7**). It is this latter site that we will visit on the field excursion today. Complete details of this study can be found in Batchelder (1991) and Yuretich (1992).

Cadwell Creek Watershed

Cadwell Creek watershed encompasses an area of 7.28 km^2 on the western shore of the Quabbin Reservoir (Fig. 6). It lies largely within the USGS Belchertown Quadrangle, with small segments in the Winsor Dam and Shutesbury Quadrangles. The watershed is forested, with the vegetation consisting of mixed softwood and hardwood, and large areas of planted red pine stand. The topography is typical of the dissected plateaus of the central Massachusetts uplands, with a maximum relief of 180 m. The drainage basin consists of several small tributaries separated by resistant bedrock rodges and sloping uplands covered by relatively thin till (Caggiano, 1977). However, the headwaters of the streams are often swampy low-gradient areas underlain by thicker surficial deposits.

Bedrock Geology .The bedrock geology is mostly Proterozoic and Paleozoic metamorphic rocks with a complex outcrop pattern typical of the mantled gneiss domes of Central New England. They possess a general north-south strike within the watershed. Four major formations are recognized: the Dry Hill Gneiss and Mount Mineral Formation (both Proterozoic Z to Cambrian); the Fourmile Gneiss (Proterozic Z to Ordovician); and the Partridge Formation (Middle Ordovician). In addition a small Mesozoic diabase dike has been mapped (Michener, 1983; Zen and others, 1983).

Surficial Geology and Soils. The surficial geology consists primarily of sandy glacial till blanketed by a thin aeolian mantle. Small areas of a clay-rich glacial till, stratified glaciofluvial sands, swamp deposits and alluvium are also present. The sandy till is believed to be younger and hence stratigraphically above the clay-rich till (Caggiano, 1977; Newton, 1978; Leonard, 1984). The upper till is a granular, poorly indurated, bouldery sandy till derived from the underlying bedrock. Textural analyses confirm a high sand content (67 to 81% sand; 16 to 27% silt and 1 to 6% clay) as well as a similarity to other tills in the region. Seismic surveys and well excavations place the till thickness at less than 5 meters, although most of the watershed has a till cover of less than 2 meters. The lower till is a jointed, fissile, compact clay- and silt-rich till with few large clasts. It is present only in the southeastern part of the catchment, where it has an estimated thickness of greater than 15 m. The small area of glaciofluvial deposit is a stratified, well-sorted, medium-grained sand; the thickness of this probable kame terrace approaches 5 m. An aeolian mantle of windblown silt and very fine sand blankets much of the landscape. It has an average thickness of 55 cm, although it is not present everywhere and commonly contains pebbles and boulders derived from the underlying till.

The soils within the watershed are predominantly glaciated upland soils belonging to the order Inceptisol. Most of the soils are Typic or Lithic Dystrocrepts, although some upland soils have

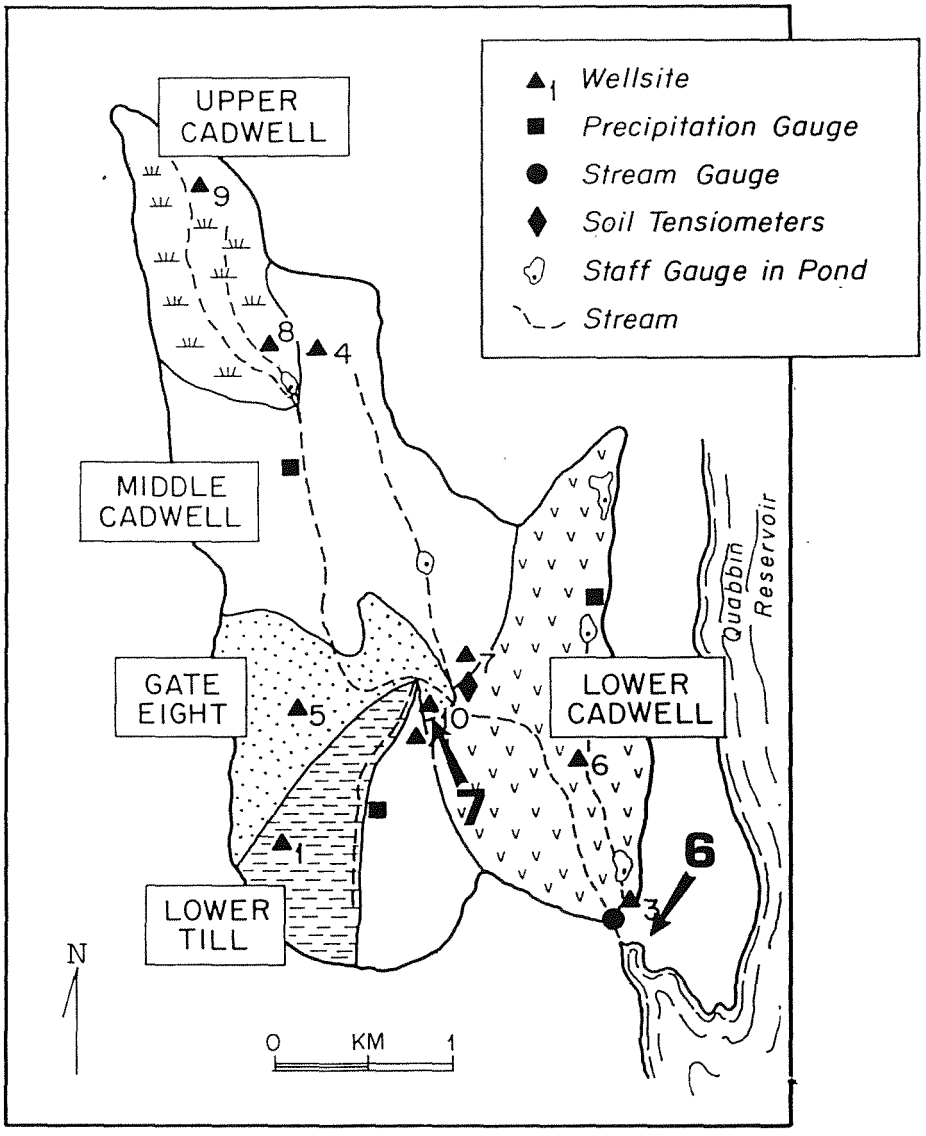

Figure 6. Cadwell Creek watershed showing monitoring network and hydrologic provinces determined from hydrogeological and topographic characteristics (from Yuretich and others, 1989). Bold numbers and arrows show the locations of **STOP 6** and **STOP 7**.

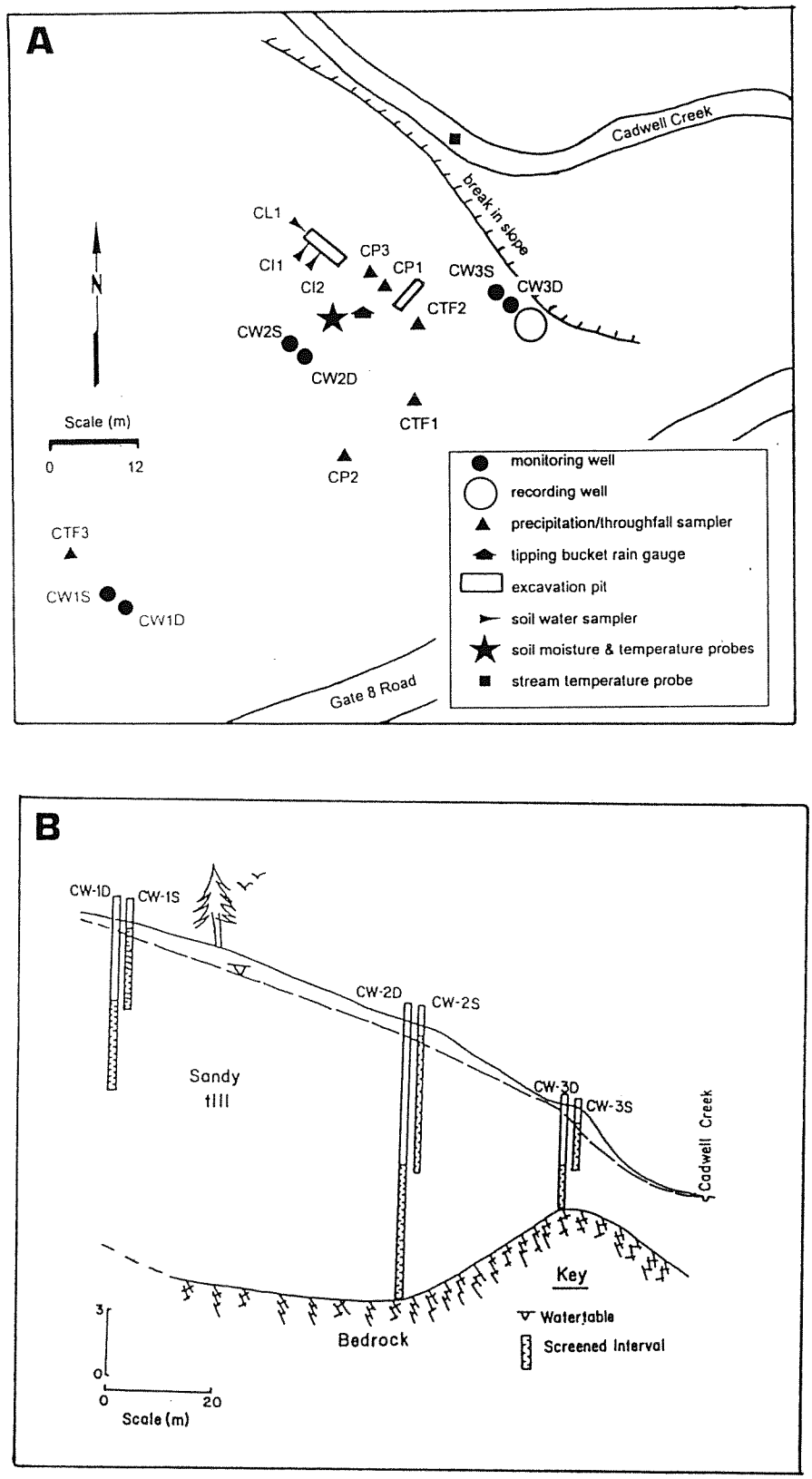

Figure 7. A. Detailed site map of instrumentation at **STOP 7**. B. Cross-section through the monitoring wells at the site (from Batchelder, 1992).

developed a compacted layer (fragipan) and are classified as Typic Fragiocrepts. A unique Entisol (Typic Udorthent), lacking good pedogenic horizons, has developed on the sand and gravel of the stratified drift deposit. Along the stream floodplains, poorly drained soils with fragipans (Aeric Haplaquepts) are common. All soils have coarse loamy to sandy textures and, except for the areally restricted floodplain soils, promote good drainage. A typical profile shows about 8 cm of O horizon, underlain by 13 cm of mineral soil containing high amounts of organic matter (A horizon). The B horizons average 43 cm thick and are composed of slightly altered material with coloration reflecting the presence of iron oxides.

Monitoring Network. This study consisted of two parts. The first was an analysis of the general characteristics of groundwater in Cadwell Creek watershed and its relationship to the stream water. The monitoring network for that purpose is illustrated in Figure 6. The second part of the study involved a more detailed examination of the pathways of precipitation as it moved from initial infiltration to discharge in the stream channel. For this project, stream gauges, closely spaced wells, soil-moisture collectors and an automatic weather station were established around our previous wellsite 10 (**STOP 7**) (Fig.7).

General Results. Cadwell Creek responds to storm events in about four hours, and the stream receives much of its water from the upper one to two meters of the water-table aquifer. This water is relatively unneutralized, giving Cadwell Creek a pH of 5.85 and an alkalinity of 26 μeq/l during 1988. The pH and alkalinity are very dependent upon stream discharge, yielding a strong inverse correlation. The hydrologic response of Cadwell Creek, especially during the periods of high antecedent soil moisture, is dominated by a build-up of the capillary fringe during storm events, with a subsequent rapid drainage to the stream channel. In addition, large differences in soil permeability and moisture regimes in the watershed enhance the contribution of variable source areas to the stream flow during storms of different size and duration.

Neutralization of precipitation acidity in Cadwell Creek is heavily weighted to silicate weathering, owing to a low cation exchange capacity in the soils and surfical materials (~1.5 meq/100g). Since well-neutralized groundwater is of less significance to the streamflow in this watershed, the alkalinity levels in the stream remain low. Silicate weathering is a slow process, as shown by scanning-electron micrographs of till grains. The correlation of groundwater composition with mineralogy of the till is poor, pointing the overiding importance of hydrology in dictating local variations in the groundwater chemistry.

Although the groundwater entering Cadwell Creek is relatively shallow and unneutralized, a large reservoir of deeper groundwater with higher pH (>7) and alkalinity (400 μeq/l) is present in the upper unconfined aquifer. Some of this water supplies baseflow to the stream in the uppermost headwaters region, but much of it apparently discharges directly into the reservoir. Such patterns of direct groundwater discharge into the reservoir may be quite common around the reservoir and will help alleviate any tendency toward acidification. Most streams entering the Quabbin Reservoir also typically have a higher pH and greater buffering capacity than Cadwell Creek (Yuretich, 1992).

ROAD LOG

Mileage

0.0 Assemble in E Lot north of University of Massachusetts Alumni Stadium. Head north on service road. Lot and stadium underlain by Lake Hitchcock clays which caused foundation problems at the stadium. Prof. Stanley Bemben (formerly Engineering Dept. U.Mass), who studied the clays, reports that they have among the highest moisture content of any similar-textured plastic clays in North America.

0.5 Left on Massachusetts Avenue.

0.6 Left (north) on Route 116.

2.0 Stop light - straight ahead, traversing Lake Hitchcock clays overlain in some places by post-lake dunes and alluvium.

7.5 Stop light, intersection with Route 47. Straight ahead, Sugarloaf Mountain and observation house at 1 o'clock ahead.

7.7 Crossing Connecticut River. This bridge is one of the most expensive of its size in New England because of construction problems on Lake Hitchcock clays (also see comments regarding these clays at start of trip). Sugarloaf Mountain on right.

8.3 Outcrop of Sugarloaf Arkose on right.

8.4 Entrance to Sugarloaf Reservation. Take two rights into reservation.

9.4 Park in parking area at top of Mount Sugarloaf.

STOP 1 - OVERVIEW OF PIONEER VALLEY. We will visit several different hydrogeologic environments and discuss a variety of flow and quality problems. Some of the topics of interest will include: aquifer contamination and protection; buried-valley aquifers; aquifer modeling; and hydrochemistry of shallow aquifers showing the influence of acid rain. At this splendid overview of the Connecticut Valley, we will first examine the geology and aquifers of the valley followed by a presentation of the hydrology and pesticide problems in Whately and South Deerfield. Some of the major aquifers in the field trip area, many of which will be examined during the course of the field trip, are buried valleys, deltas, high yielding stratified drift below lake clays, valley train aquifers of fluvial stratified drift or collapsed kame terrace, Triassic sedimentary bedrock, crystalline bedrock, and till.

10.2 Left on Route 116.
10.7 Connecticut River
13.6 Sunderland Hatchery - private fish hatchery on right. Warner Brothers Quarry in Sunderland Delta to left.
14.0 Left on East Plumtree Road.
14.3 Right into Cronin National Salmon Station, U.S. Fish and Wildlife Services.
14.4 Park vehicles in parking area next to building.

STOP 2A - GROUND WATER DISCHARGE AREA of SUNDERLAND DELTA. The Sunderland Delta is a prime illustration of the influence of glacial geology on water resources of the Connecticut Valley. The highly permeable soils along Long Plain Brook result in high amounts of recharge upstream from the former delta plain. Impermeable lake clays interfinger with the foreset beds near the toe of the delta, forcing water to the subsurface as the springs and streams which are put to good use by the fish hatcheries.

14.5 Leave station - right on East Plumtree Road.
14.7 Turn around and pull off road near entrance to old gravel pit.

STOP 2B - RECHARGE AREA OF SUNDERLAND DELTA. Proceed west on East Plumtree Road.

15.2 Turn left (south) on Route 116.
16.1 Turn left on North Pleasant Street. Podick Recreational Area (Town of Amherst) on right.
17.0 Traffic light, North Amherst, straight ahead.
18.0 Left on traffic light, Univ. of Mass. dormitories on left and right.
18.7 Right on East Pleasant Street.
18.8 Univ. of Mass. Wind-Energy Research Center on right.
19.1 Left on Strong Street.
19.2 Wildwood School (elementary) on right and pond on left. This is a perennial pond that's probably fed by faults from the bedrock aquifer. The pond is located on a surface-water drainage divide. Water drains north from the northern part of pond and south from the southern part.
20.1 Right on North East Street. We have entered an embankment underlain by Lake Hitchcock clays. Pelham Hills to the east.
20.8 Stop light - Main Street straight ahead.
21.1 Stop light - Route 9 straight ahead.
22.1 Junction of Hop Brook which is outlet for Lawrence Swamp Basin and Fort River, U.S. Geological Survey stage recorder is about 500 feet west on Fort River.
22.2 Railroad underpass.
22.4 Paleozoic bedrock outcrop on left. The western basin divide is formed by a bedrock ridge and drumlins like Mt. Castor and Mt . pollux.
Lawrence Swamp on left.
23.2 Divided road - bear left at South Amherst Common.
23.3 Immediate left on to Station Road, traveling from Basin Divide down into Lawrence Swamp Basin.
23.75 Beginning of Robert Frost Trail through part of Lawrence Swamp.
23.8 Hop Brook, Master Stream of Lawrence Swamp Basin - all surface and ground water discharged from the basin exits in this stream.
24.0 Flowing well screened in sand and gravel underlying Lake Hitchcock clays.
24.6 Cortland Drive - turn right and park.

STOP 3. OVERVIEW OF LAWRENCE SWAMP. The aquifers underneath Lawrence Swamp supply approximately 2 mgd to the water supply of the Town of Amherst, although the capacity is more than double this figure. Much of the water comes from a buried stratified-drift aquifer separated from the unconfined aquifer by Lake Hitchcock clays.

24.6 Turn around and continue east (right) on Station Road.
25.3 Turn right on Warren Wright Road.
25.45 On left, sandy and silty upper till with high infiltration capacity compared to most tills. Forms part of the secondary recharge area for Lawrence Swamp Basin.
25.6 On left, Kame Delta consisting of medium to very coarse sands interbedded with gravel. This stratified drift has a very high infiltration capacity and forms the primary recharge area.
25.8 On left, alluvial fan deposits.
26.0 On left, entrance to Brown Well which is screened in the deeper stratified drift aquifer. It has a yield of approximately 1ø mgd and is one of Amherst's most productive wells.
26.5 Turn right on dirt road - Robert Frost Trail. Follow along edge of field for approximately 0.5 miles.
26.7 On right, sand dune topography.
27.0 Park in field next to wells.

STOP 4, TOWN OF AMHERST WELL NO. 6 Well No. 6 is one of four large-capacity (> 1.3 mgd) production wells which simultaneously pumped from the confined Lawrence Swamp aquifer at a combined rate of 5.5 mgd during a continuous ten-day aquifer pumping test during June, 1991.

ALTERNATIVE STOP, IF WELL 6 ROAD IS WASHED OUT

26.5* Left on Goodell Road.
27.3 Turn right on Federal Street.
27.7 **ALTERNATIVE STOP 4 - DAIGEL WELL**
28.1 Turn left on Federal Street
28.5 Turn left on Goodell Road
29.3 Turn left on Warren Wright Road and continue to Atkins Farm Fruit Bowl by following original Road Log.

* The Road-Log mileage from the alternative stop is in parentheses.

27.0 Leave Stop 4 area.
27.5 (29.3) Turn right on Warren Wright Road
27.7 (29.5) Crossing Hop Brook.
27.8 (29.6) Right on Orchard Street.
28.8 (30.6) Right on Bay Road. Bay Road Well is in field behind houses to right.
31.3 (33.1) Left on Route 116 to Atkins Farm Fruit Bowl.
31.4 (33.2) Turn right into parking lot.

LUNCH STOP. Lunch may be purchased in left and rear part of store. Caravan will retrace route on Bay Road.

31.5 (33.3) Turn right (east) on Bay Road.
33.5 (35.3) Left (north) on South East Street. Lawrence Swamp on right. Pelham Hills in distance.
34.5 (36.3) Entrance to Well 4, Baby Carriage Brook on right, view of Lawrence Swamp.
34.7 (36.5) Drumlin (Mt. Pollux) on left.
35.4 (37.2) Bear right - South Amherst Commons.
37.1 (38.9) Right on Stanley Street.
37.8 (39.6) Right on Route 9.
38.8 (40.6) Right on Old Belchertown Road for parking.

STOP 5 - OLD AND NEW AMHERST LANDFILLS. The old Amherst landfill, on the west side of Route 9, was operated as an unlined sanitary landfill from 1971 until 1983. During the latter stages of its existence, some problems with leachate migration developed, causing the Brickyard Well Field (about 1000 m to the west of the landfill) to be shut down. Rust-colored water also appeared in springs emanating from the base of the landfill terrace in the vicinity of the Amherst Fields housing development. In general, the leachate from the landfill is confined to a small area and

is separated from the main aquifer for the town of Amherst by Lake Hitchcock clay (Schwalbaum, 1983). The new landfill, just across the road on the east side of Route 9, is a fully engineered landfill, with a clay liner and leachate recovery system.

Return to Route 9 and turn left (west).
40.4 (42.2) Bear rught at intersection by College Inn Apartments
40.5 (42.3) Turn right at stop sign onto South East Street heading north.
40.7 (42.5) Turn right at light onto Pelham Road.
41.0 (42.8) Crossing Fort River.
42.7 (44.5) Cemetery on right on kame terrace deposit.
43.7 (45.5) Hawley Reservoir, one of three surface water reservoirs for the Town of Amherst, is on the right.
46.2 (48.0) Turn right (south) on Route 202.
46.5 (48.3) Outcrops of Pelham Quartzite exposed on right.
48.8 (50.6) Turn left onto Gate 8 Road.
51.4 (53.2) Pull into parking area on left.

STOP 6 - QUABBIN RESERVOIR AND MOUTH OF CADWELL CREEK. This is the boat-launch area of the Quabbin Reservoir, which extends eastward from this site. Alkalinity has been showing an apparent decline in the reservoir water since the time of its initial filling, and trout hatchlings introduced into the reservoir in this area have had problems surviving. Cadwell Creek discharges just behind the boat launch, and both the pH and buffering capacity of this stream are low, contributing to the problem. Additional studies indicate that most of the streams entering the Quabbin Reservoir are better buffered.

Turn around and head back along Gate 8 Road.
52.6 (54.4) Pull off onto right side of road.

STOP 7 - MONITORING SITE AT CADWELL CREEK. Detailed studies of precipitation input, soil moisture, ground water and stream water at this site have revealed an exceedingly complicated hydrological relationship (Batchelder, 1991). Stream flow responds rapidly to precipitation, and only very shallow groundwater feeds the stream. Although there are sizeable volumes of water in the till, geochemical comparisons indicate that much of it underflows Cadwell Creek, presumably feeding the reservoir directly. Such patterns are responsible for the "flashiness" of Cadwell Creek and its poor acid-neutralization capacity, but the relatively good buffering of the deeper ground water helps prevent acidification of the Quabbin Reservoir.

This is the last geological stop of the field trip and the rest of the road log provides directions back to Alumni Stadium, the starting point of this trip.

Continue in same direction along Gate 8 Road.
54.0 (55.8) Intersection with Route 202, turn right (north).
56.7 (58.5) Turn left on Pelham Road towards Amherst.
62.3 (64.1) Traffic lights at intersection with South East Street , continue straight.
62.9 (64.7) Traffic lights at Triangle Street intersection; continue straight ahead.
63.3 65.1) Downtown Amherst intersection with North Pleasant Street; continue straight on Amity Street.
64.0 (65.8) Traffic light, turn right onto University Drive.
64.2 (66.0) Turn left toward Alumni Stadium
64.3 (66.1) Turn right onto service road.
64.4 (66.2) Turn left into E Lot.
END OF FIELD TRIP

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STAUROLITE PORPHYROBLASTS AND DEFORMATION IN THE GREAT HILL SYNCLINE, BOLTON, CONNECTICUT

by

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INTRODUCTION

An NEIGC field trip devoted to staurolite porphyroblasts? Given the current controversy surrounding porphyroblast rotation, the idea is not as absurd as it might seem. In a series of recent papers, Bell and his coworkers (Bell et al, 1986; Bell and Johnson, 1989; Johnson, 1990; Bell et al, 1992) have vigorously challenged the established view that curved inclusion trails document the rotation of a porphyroblast during its growth. They argue that strain partitions during deformation in such a way that porphyroblasts and other rigid objects remain fixed relative to geographical coordinates, if the matrix deforms in a ductile manner. The dispute is not trivial as it calls into question the use of inclusion trails as shear sense indicators. In fact, if Bell's views are correct, the conventional basis for the interpretation of the kinematics of deformation are suspect. Bell himself (Bell et al, 1992) considers his 'radical re-interpretation' of porphyroblast inclusion trails to be the foundation of a major paradigm change in structural geology.

Although Bell and his students have been exploring the implications of of these ideas since 1985, it is only recently that their basic assumption has received critical attention. Passchier et al (1992) have argued that the evidence for the irrotational behavior of porphyroblasts is unsound. Bell et al (1992) strongly dispute this conclusion and provide a mountain of new data to support their original interpretation. Part of the confusion seems unavoidable, as evidence from garnet porphyroblasts forms the basis for many of the arguments. Inclusion trails record only the relative rotation of the crystal and the foliation. For spherical porphyroblasts, such as garnet, the relative rotation is the same for all crystals, no matter what the actual kinematic situation. The ambiguity is resolved if the porphyroblasts are non-spherical. The motion of inequant rigid particles in a ductile matrix is complex and is a function of their dimensional orientation. Studies of syntectonic inclusion trails in non-spherical metamorphic minerals, such as staurolite, are thus one way to critically test Bell's ideas.

Large staurolite porphyroblasts occur abundantly throughout the Littleton Schist in the the Great Hill syncline in Connecticut. We have recently studied the nature of the inclusion trails in staurolites on both limbs of the syncline at Bolton, Connecticut (Busa and Gray, 1992), and concluded that they could not have been 'pinned' during growth, but rolled, yawed and twisted in a manner that depended on their initial orientation. The purpose of this trip is to examine these spectacular staurolite crystals in the field, and to explore the possible connections between their orientation, morphology, size, and the style of local deformation.

STAUROLITE PORPHYROBLASTS

Centimeter-sized staurolite porphroblasts in the Bolton area display four distinct habits (Figure 1a-d).

(a) anhedral sigmoidal cylinders; as large as 8 x 1.5 x 1 cm
(b) eudedral columnar prisms; as large as 15 x 3 x 2 cm
(c) radiating fan-shaped crystal aggregates; 2 to 4 cm in length and 0.5 cm thick
(d) lozenge-shaped paper-thin crystals. as large as 30 x 5 x 0.1 cm

The anhedral variety typically contain sigmoidal quartz inclusion trails that curve through angles of up to 135°. Serial sections of individual porphyroblasts demonstrate that the inflection hinge line and the statistical 'symmetry axis' of inclusion surfaces are centrally located along the long axis irrespective of its orientation. The apparent axial

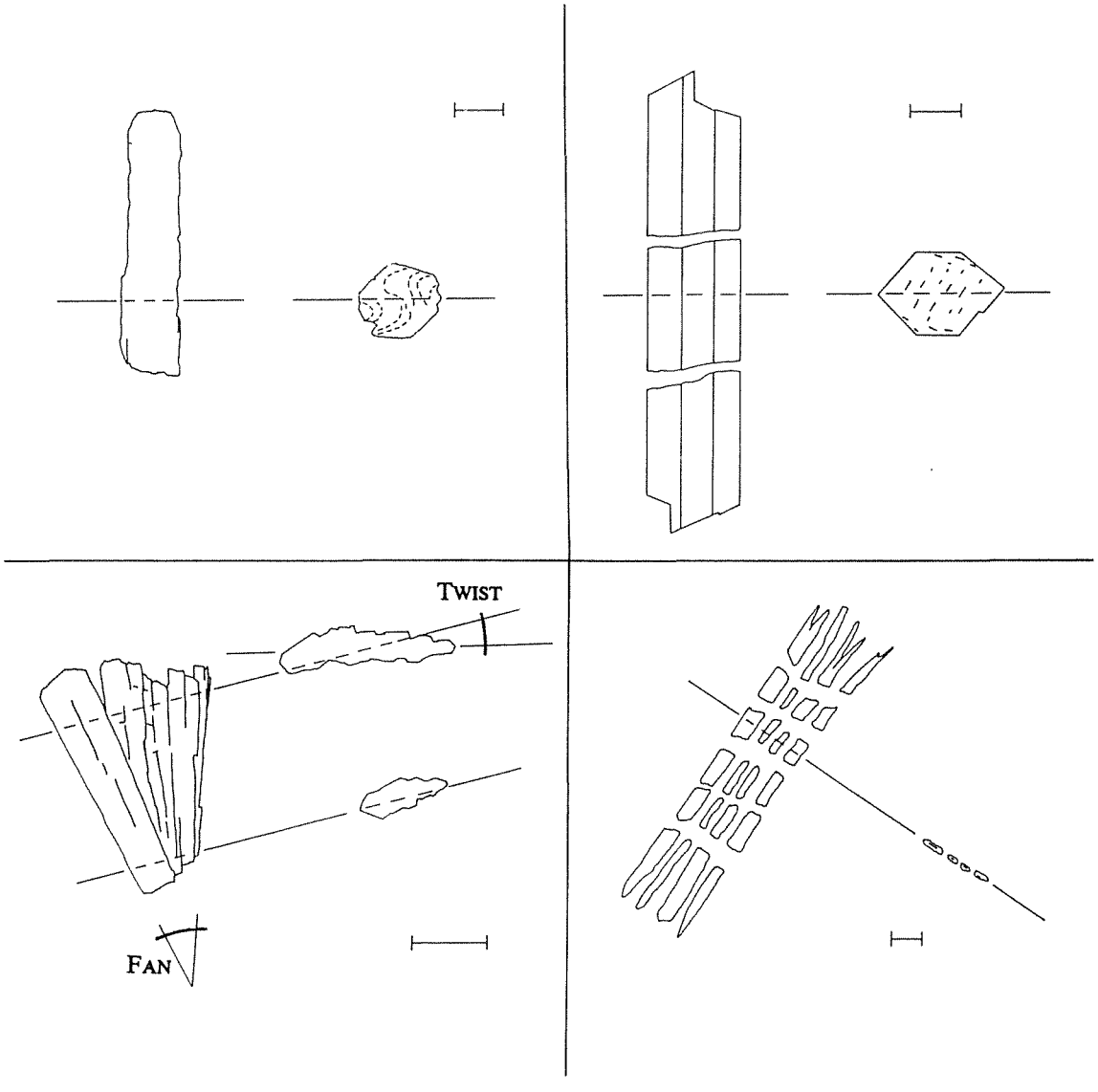
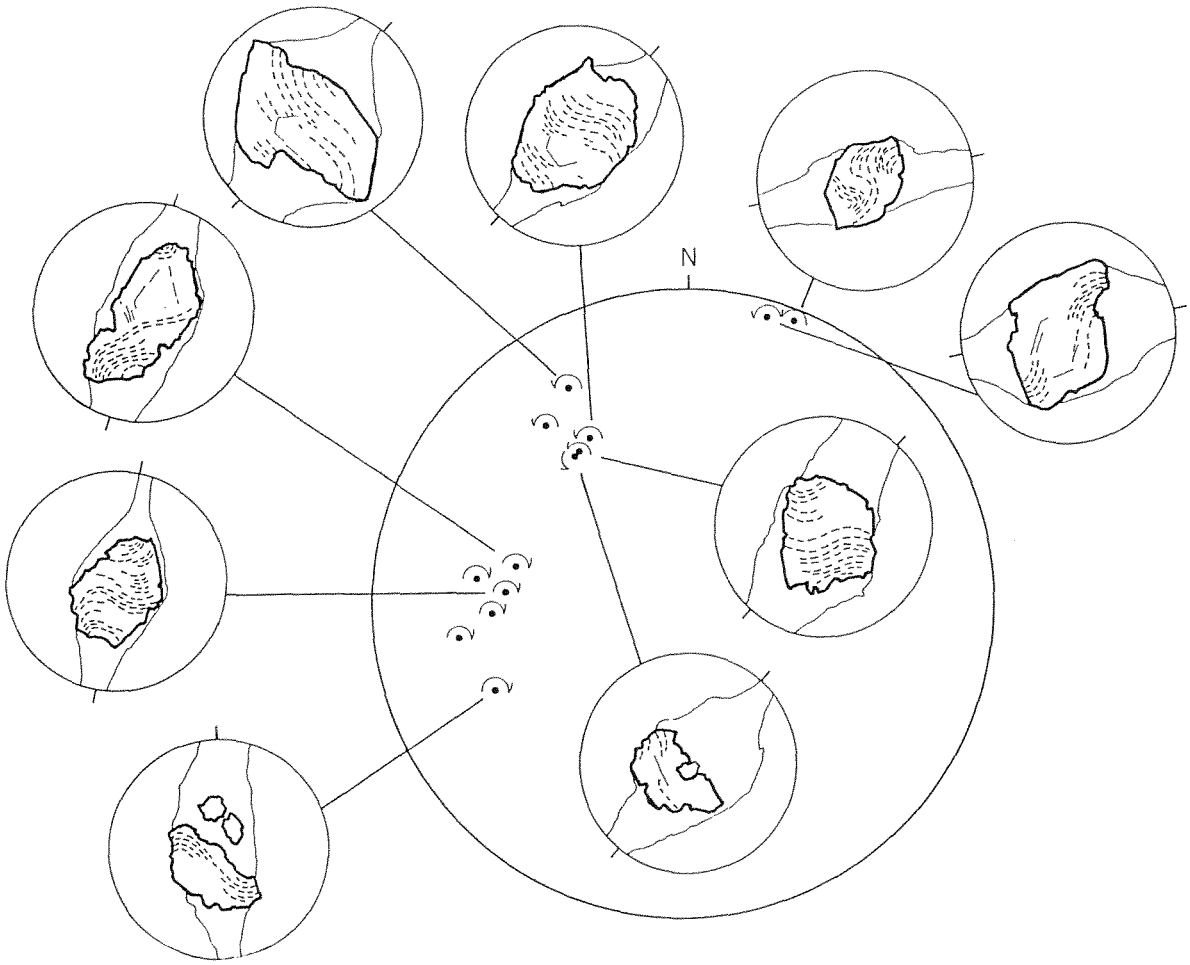
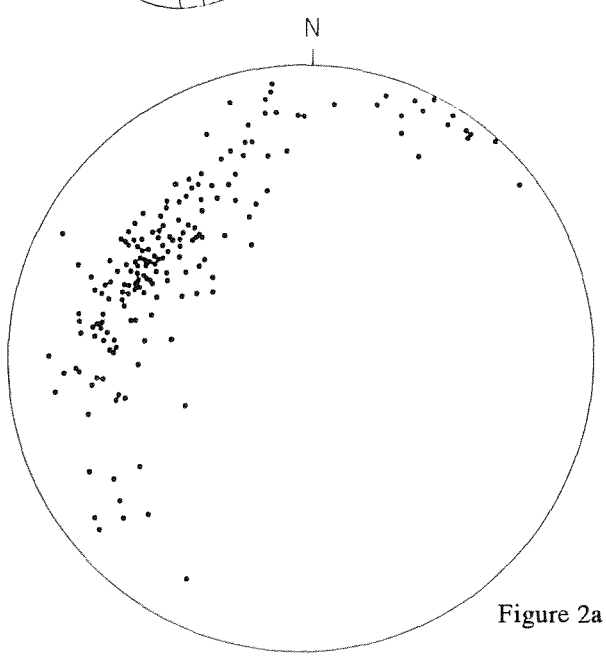


Figure 1

Staurolite porphyroblast various growth habits, Great Hill syncline, Bolton-Vernon, Connecticut. Plan and cross-sectional views. Scale bar is 1 cm.

(a) Anhedral staurolite porphyroblasts with sigmoidal-shaped inclusion trails
(b) Eudedral staurolite porphyroblasts with planar inclusion trails, except at the crystal margins.
(c) Fan-shaped staurolite porphyroblasts showing the 'fanning' of individual staurolite subcrystals which make up the porphyroblast
(d) Lozenge-shaped staurolite porphyroblasts. Muscovite fills the spaces between the boudinaged fragments.





EAST LIMB

Figure 2a Stereograms showing the preferred orientation, and the amounts and senses of rotation of staurolite porphyroblasts from the Great Hill syncline, Bolton, Connecticut (east limb)

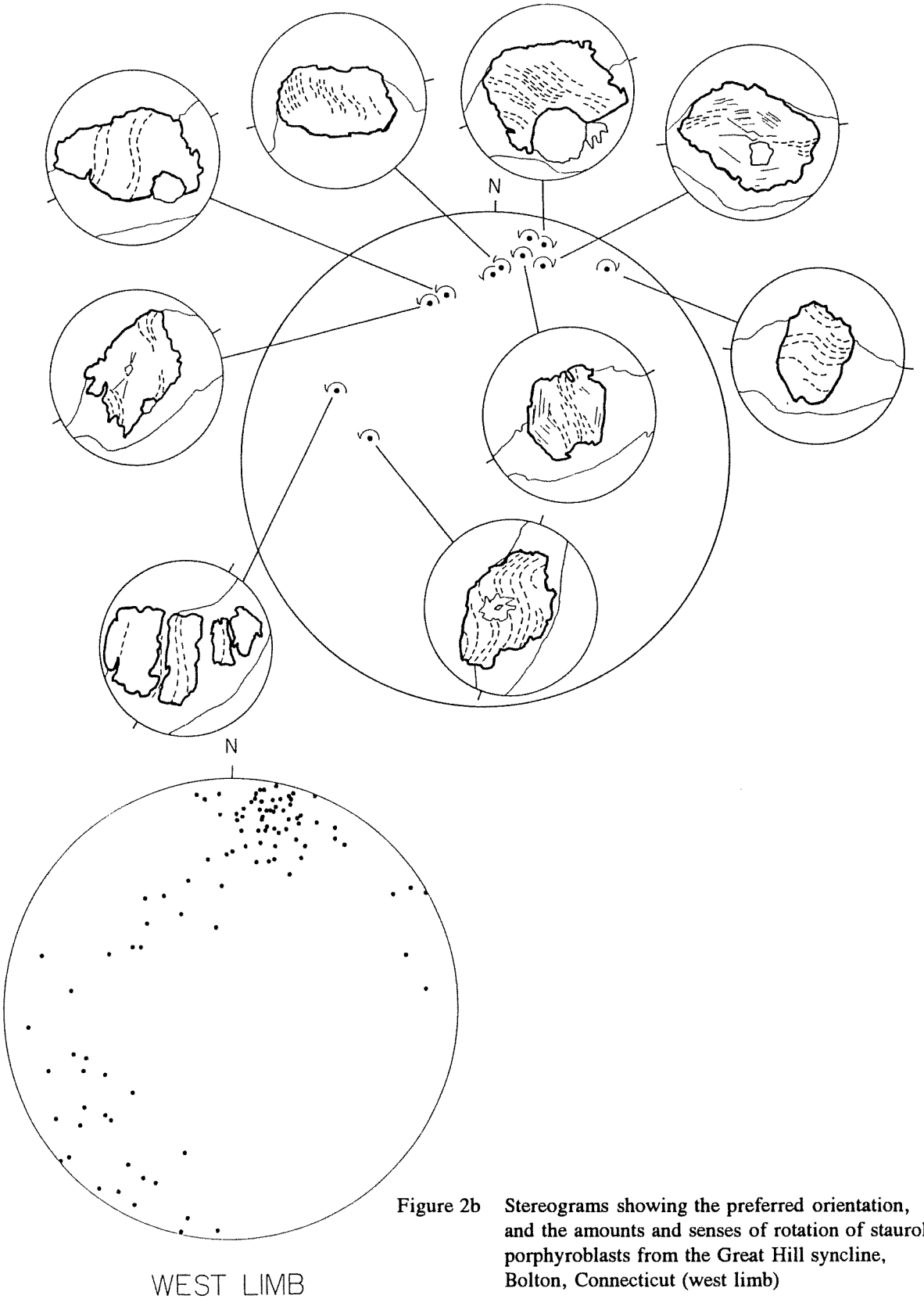


Figure 2b Stereograms showing the preferred orientation, and the amounts and senses of rotation of staurolite porphyroblasts from the Great Hill syncline, Bolton, Connecticut (west limb)

rotation defined by the curvature of inclusion trails is a maximum at the initial nucleation site and decreases towards both ends of a crystal. The sense and amount of rotation recorded by each porphyroblast is related to its orientation (Figures 2a and 2b). The transition from clockwise to counterclockwise rotation constrains the sense and direction of the shear that accounts for apparent rotation of the porphyroblasts. On the east limb of the Great Hill syncline, the inferred slip is in the direction 295° plunging 30°, with a top-to-the-west shear sense. Trails in the porphyroblasts on the overturned limb imply a top-to-the-south sense of shear in a direction 010° plunging 15°. In both cases, the weak preferred orientation of the anhedral porphyroblasts coincides with the inferred slip direction (Figures 2a and 2b). Busa and Gray (1992) have demonstrated that the morphology and orientations of the trails in the anhedral porhyroblasts are quatitatively consistent with deformation-induced rotation.

The euhedral porphyroblasts show no evidence of substantial rotation during growth. Their inclusion trails are largely planar, curving into parallelism with the matrix schistosity only at their very margins. Rotation, it would seem, inhibits the development of plane crystal faces. In the Bolton area, anhedral staurolite porphyroblasts are concentrated in a zone near the base of the Littleton. The large euhedral crystals are more commom higher in the section in what is now the core of the syncline. Perhaps the different morphologies can be used to map out the distribution of shear throughout the Littleton at the time of staurolite metamorphism.

The fan-shaped habit represent an extreme case of synrotational staurolite growth. Rather than a single anhedral crystal, these porphyroblasts are formed of radiating groups of bent and broken crystals. The orientations of the individual staurolite subgrains vary systematically across a fan, such that the cross section is subtly sigmoidal and the whole grouping has a helicoidal twist. It appears that the small fragments continually break off the twisting and yawing porphyroblast, only to be recaptured by its growth. Why some rotating staurolites grew as single crystals, while others broke-up to form fan-shaped aggregates is unclear. Perhaps the rate of deformation is a factor.

The lozenge-shaped staurolite porphyroblasts are restricted to papar-thin layers in fine-grained quartzites. Their size within these lamellae is quite extraordinary and is probably related to the difficulty of nucleating staurolite.

GREAT HILL SYNCLINE

The rocks comprising the Great Hill syncline are the Clough Quartzite, the Fitch Formation, and the Littleton Formation (Bolton Group - Rodgers, 1985). The Great Hill syncline is mapped as a tightly folded, overturned to the east syncline. The Clough is reported to be unconfomably underlain by the Glastonbury Gneiss, south of Bolton (Rosenfeld and Eaton, 1985), although we have not found the exposed contact. Aitken (1955) reports the Glastonbury Gneiss - Bolton group contact is conformable at the southwest portion of the Rockville quadrangle, but the contact becomes angularly unconformable in the northern portion of the quadrangle. The contact between the Clough and the Ordovician Collins Hill Formation can be observed at Bolton Notch (STOP 1, Figure 3). The syncline can be traced from Cobalt, Connecticut (approximately 25 km south-southwest of Bolton Notch) along a north-northeasterly direction to Monson, Massachusetts, "forming apparently the most extensive continuous Primary mountain range in the State" (Percival, 1842). The width of the syncline averages approximately 1 km, being the widest in the Bolton area (2 km), and the most narrow in West Stafford (1/2 km).

The structural and metamorphic history of the Great Hill syncline is complex and poorly understood. Even the timing of the staurolite metamorphic event is controversial. Rosenfeld and Eaton (1985) suggested that staurolite growth post-dated the formation of the syncline, whereas Robinson and Tucker (1982) argued that the folding post-dates the staurolite metamorphic event. As will be seen in the paragraphs to follow, neither argument agrees with the structural data inferred from the rotated staurolite porphyroblasts from Bolton, Connecticut.

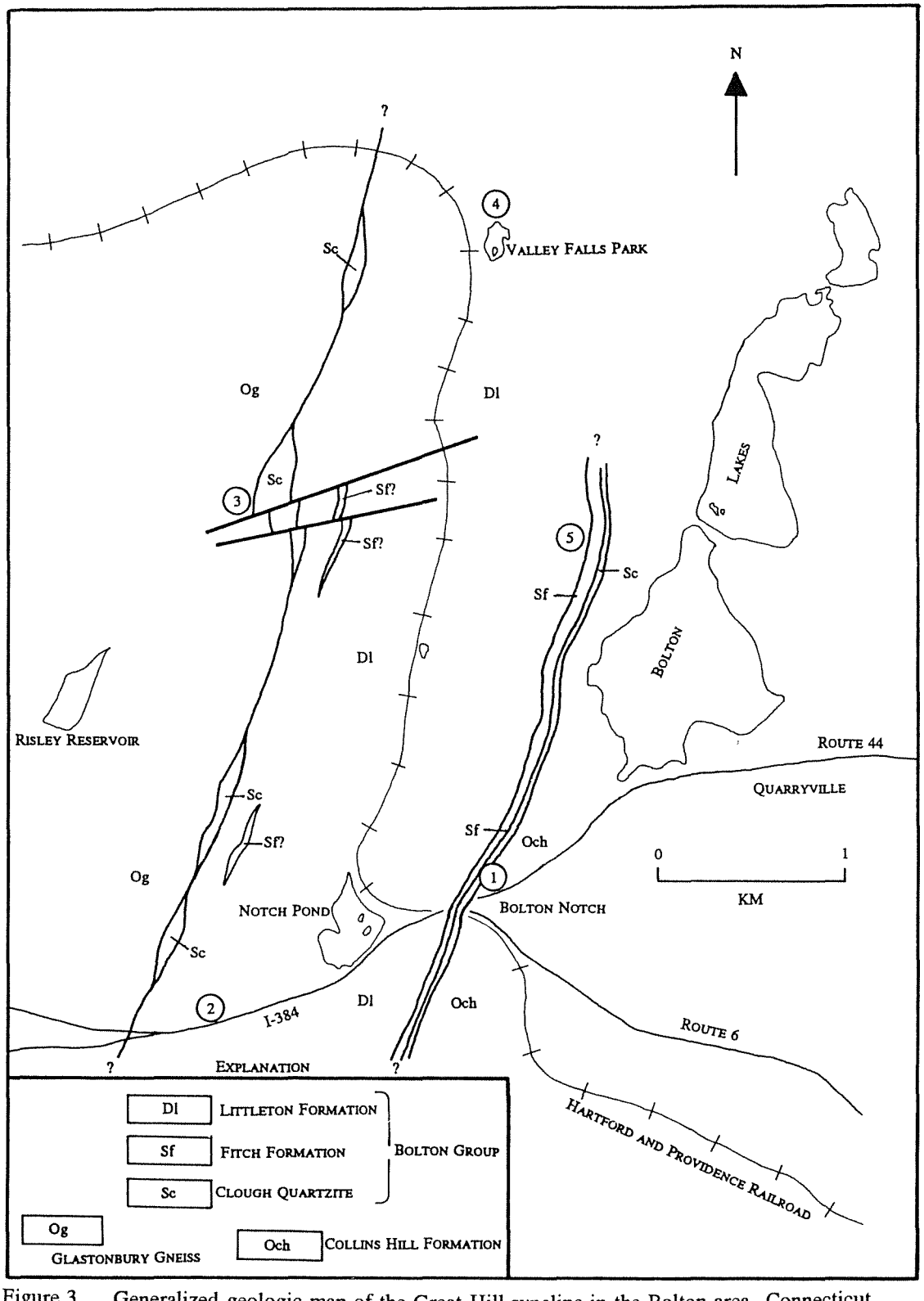


Figure 3 Generalized geologic map of the Great Hill syncline in the Bolton area, Connecticut. Numbers in circles designate stop area locations.

ROAD LOG

Bolton Notch, Connecticut is located approximately twenty miles east of Hartford. From Massachusetts, take I-91 south to Hartford, then take I-84 east to I-384. Stay on I-384 until the expressway ends/splits into Routes 6 and 44, at Bolton Notch. Take the Route 44 exit, stop at the first light and park in the W.H. England Plaza. The field trip begins 9:30 AM, Sunday October 11th. We will assemble and finish at the W.H. England Plaza. (The entire field trip will be in the Rockville quadrangle)

Mileage
0.0 Left onto Route 44 (west).
0.1 Turn into an unpaved empty lot on the right-hand-side of Route 44.

STOP 1. BOLTON NOTCH: SYNTECTONIC ANHEDRAL STAUROLITE PORPHYROBLASTS. (90 MINUTES) Park at the base of the advertising billboard. Follow the small path on the north side of the sign up the hill to the crest of the first ridge. The small outcrops along the path and on the top of the ridge are Ordovician sulfitic biotite gneisses. Leave the trail and walk to outcrops of Silurian Clough Quartzite at the foot of the cliff on the other side of the small valley. The contact of the Clough is not exposed here but we will have a chance to study it in the railroad cut on the south side of the highway, before returning to the cars. The detailed stratigraphy of the Clough is worth noting as it is duplicated on the west limb of the Great Hill syncline on Box Hill (STOP 3). The correlation is the best evidence for the overturned nature of the syncline, as minor structures and other stratigraphic evidence are ambiguous. Near its base the Clough here is characterized by deformed quartz pebbles, 1 to 3 millimeter-sized magnetite grains and the virtual absence of garnet. The pebbles and magnetite disappear a few meters up in the section and garnet makes its appearance as large, centimeter-sized flat, disk-shaped crystals confined to the thin muscovite schist lamellae. The average thickness of the quartzite beds decreases systematically from 20-30 cm at the base, to less than a centimeter near the top of the unit. The total thickness of the Clough in the Notch is twelve meters.

We will follow the quartzite along the slope to the south and then work our way up the edge of the cliff face. The top of the Clough here is marked by a 30 cm thick bedding parallel fault zone. Today it can be traced south along the cliff face, across the Notch, and into the modern quarry on the south side of the highway. In the 1800s a similar, if not the same, fault was exposed in quarries two miles north of here. Quarrymen at the time referred to the zone as the 'Diamond Reef' on account of the rhomboidal fragments into which it characteristically splits (Shepard, 1837). Fabrics in the fault zone indicate a top-to-the-south sense of shear.

Ten meters of a homogeneous, fine-grained muscovite-biotite-quartz gneiss overlies the Clough. The gneiss was formerly quarried for flags and grindstones as it splits readily into exceptionally uniform large planar slabs two or more meters across. Quarries both north and south of the Notch were active between 1812 and the late 1880s. Flagstones from Bolton were used for paving streets in Hartford, New York, Washington, Philadelphia, and Baltimore. Grindstones weighing several tons were shipped to Boston, New York and Albany.

Finely laminated, differentially weathered calcareous rocks of the Fitch Formation conformably overly the non-calcareous gneiss. A distinctive light gray calc-silicate layer marks its lower boundary. Alternating thin lamellae of a brown weathering calc-silicate and a non-calcareous biotite gneiss make up most of the ten meter thickness of the unit. Towards the top of the section, thick massive light green actinolite marble layers make an appearance. Meter size cavities are developed along some joints and small faults. The largest of these caverns extends some ten meters into the cliff. A number of colorful 'legends' are associated this cave which is known locally as 'Squaw Cave'.

A prominent overhang, two-thirds of the way up the cliff, marks the boundary between the calc-silicate rocks and the silvery-grey Littleton garnet- and garnet-staurolite schists. A dark green and white mottled amphibolite skarn, 1 cm to 1 m thick, is developed along the contact. Unlike its lower contact, the transition

from calc-silicate to pelitic schist is exceedingly sharp and abrupt. Staurolite crystals in the Littleton are only about 5 mm long at this contact but they rapidly increase in size toward the top of the cliff where centimeter-sized anhedral porphyroblasts are common.

The top of the cliff is reached following a crude trail (maintained by Introductory Geology students) south from the staurolite schist overhang. The route is somewhat precipitous and does involve some hand-over-hand climbing. An easier, but longer route, leads 100 meters north along the staurolite schist contact to a well-worn path that swish-backs up the slope and then runs south along the top edge of the cliff.

Follow the cliff edge to an overlook sixty meters to the south. From this vantage it is evident that the valley of the Hop River heads off to the southeast, directly away from the Notch. The Notch itself was probably a water-gap for a preglacial ancestor of todays Hop River. A small fault with a horizontal displacement of less than ten meters which runs along the base of the cliff through the Notch, probably controlled the location of the gap. To the west, the buildings of Hartford and the basalt ridges of the Connecticut River Valley are visible on the skyline. The north-south ridge closest to us is Box Hill, the opposite limb of the Great Hill syncline.

Our path now takes us 200 meters down the dip slope of the schist through a maze of poorly marked trails, past a small pond and up onto an outcrop ridge. The Littleton here contains large anhedral staurolite porphyroblasts in abundance. The sigmoidal inclusion trails in sections through variously oriented crystals from this locality are illustrated in Figure 2a. The amount and sense of rotation implied by these trails suggests the porphyroblasts grew whilst the matrix was undergoing simple shear with top-to-the-west slip. Against the backdrop of this strikingly porphyroblastic schist we will review the current controversy surrounding deformation partitioning and porphyroblast rotation. We will point out the features displayed by the inclusion trails (a hand lens is essential), the sector zoning and preferred orientations of these giant staurolite porphyroblasts that lead us to the conclusion that they rotated as they grew.

The crest of the ridge is capped by a large pegmatite body. At the cliff edge the mineralogy and texture of the pegmatite are completely obscured by thirty years of paint and graffiti. The overlying schist was neatly plucked away by ice so that the topography of the pegmatite outcrop mimics the form of its upper contact surface. The pegmatite has been boudinaged in a 'chocolate-block' fashion on a grand scale. The individual boudins are several meters across. The thinned down necks between the blocks form an orthogonal system of distinct furrows across the surface of the outcrop. Quartz veins up to a half meter wide fill rifts that formed along some of the more pronounced furrows.

While on this outcrop it is also worth noting that the staurolite porphyroblasts within a few decimeters of this pegmatite are pseudomorphed by muscovite. This one simple observation firmly establishes the relative ages of staurolite metamorphism, the pegmatite emplacement and the boudinage (which may be related to the rise of the Glastonbury dome and the overturning of the Great Hill syncline).

Now follow a well-marked trail 250 meters down the dip slope of the staurolite schist to the old Hartford and Providence Railroad. Proceed 100 meters north along the roadbed into the railway cuts. Staurolite schists and a well-bedded, fine-grained, grey quartzites are complexly folded together here. Isoclinal fold noses are preserved in large quartzite boudins which swim in a sea of schist. Locally the quartzite splits into remarkable paper-thin, bedding-parallel sheets, a meter or more across. Randomly oriented muscovite pseudomorphs (after staurolite ?, or kyanite ?) up to half a meter long, 5 mm wide and less than 1 mm thick ornament the surfaces of some sheets.

Return east along the railway bed to the 'giant' staurolite schist at the base of the cliff below the painted pegmatite. The abundance of large anhedral staurolite porphyroblasts in the schist here is quite extraordinary. The serially sectioned staurolite crystals illustrated in Busa and Gray (1992) were collected at this site.

Continue east along the railway bed through the tunnel. The lower contact of the Clough Quartzite is exposed on the north face of the open cut, twenty meters from the eastern portal. Although the lower part of the Clough is a pure, white pebbly quartzite, its base is marked by a half meter thick muscovite-rich quartz schist. Is the contact a simple unconformity or is it a major fault zone? A compete section through the Clough and Fitch is exposed in the dark of the tunnel. Most of the interesting features noted on the cliff; the 'Diamond Reef' fault, the light colored calc-silicate layers at the top of the non-calcareous portion of the Fitch and the mottled amphibolite skarn at the base of the Littleton, are also found in the tunnel. Return through the tunnel to a path leading from its west end up to the north side of the highway which we will follow back to our starting point.

Mileage

0.1 Right onto Route 44 (west).
0.6 Bear right and take the Middle Turnpike Exit (Routes 44 & 6).
1.1 Just before the overhead traffic light warning turn right into the United Artists Cablevision driveway.

STOP 2. UNITED ARTISTS: PRETECTONIC EUHEDRAL STAUROLITE PORPHYROBLASTS

(10 MINUTES) Park on the west side of the United Artists Cablevision lot. The outcrops of Littleton Schist in the woods to the west contain some the largest staurolite porphyroblasts in the Bolton area. The rocks here lie on the overturned limb of the Great Hill syncline and are stratigraphically higher than any seen at STOP 1.

The centimeter-sized crystals were originally euhedral. Some were even twinned. Most are now broken and microboudinaged, presumably by the same post-metamorphic flattening that affected the painted pegmatite in the Notch. The largest complete staurolite crystal collected here measured 20 x 2 x 3 cm. Why so large? What factors determine the size of metamorphic porphyroblasts? Why are the garnet crystals in this exceptionally porphyroblastic schist no more than a few millimeters in diameter? Perhaps it is significant that garnet growth in this rock predates the staurolite. Garnet inclusions in the staurolite are similar in size and crystal form to those in the matrix.

Graphite and ilmenite are the principal inclusions in the staurolite. Their trails are for the most part planar, curving through 90° into parallelism with matrix foliation only at the very edges of the porphyroblasts. Graphite trails in the included garnet crystals parallel those in the host staurolite. This observation suggests that the garnet-staurolite metamorphism at this site was here not associated with the intense penetrative deformation that produced the sigmoidal trails in the anhedral staurolites of the Notch.

Mileage

1.1 Turn right onto East Middle Turnpike (Routes 44 & 6).
1.4 Traffic lights at the Route 85 intersection. Continue straight on East Middle Turnpike (Routes 44 & 6).
2.3 Left into the parking lot of the Shady Glen Restaurant.

LUNCH

Mileage

2.3 Right onto East Middle Turnpike. Quickly move into the far left lane in order to make a left turn onto Lake Street.
2.4 Left onto Lake Street.
3.1 Stop sign. Route 85 joins Lake Street. Continue straight.
3.6 Vernon town line.
4.2 Take a sharp right onto Box Mountain Drive.
4.8 Left into Box Mountain Quarry.

STOP 3. BOX HILL: SYNTECTONIC FAN-SHAPED STAUROLITE PORPHYROBLASTS. (60 MINUTES) Park just outside the gate. Most of the Clough Quartzite quarried here is used locally for patios, fireplaces and stone walls. Selected thin slabs with large flat pink garnets are marketed nation-wide as a veneer stone. The stone is used as a facing on one national chain of gas stations and it adorns many buildings in the midwest. Disseminated arsenopyrite causes some discoloration problems. The yellow-green stain which colors the exposed surfaces of some slabs is the mineral scorodite ($FeAsO_4 \cdot 2H_2O$), a weathering product of the arsenopyrite. With time the green color turns to rust as the scorodite weathers to limonite. The stratigraphically highest quartzites in the quarry seem to be the most susceptible to scorodite discoloration. Angry customers whose expensive retaining walls turned an ugly green less than a year after they were constructed have made the quarrymen very wary of the upper quartzite.

The Clough here is on the overturned limb of the Great Hill syncline. The stratigraphy noted on the right-side-up limb at Bolton Notch is inverted here and can be matched in most detail. The lowest quartzites are pebbly, magnetite rich, and thick-bedded. Garnet makes its appearance higher in the section and becomes particularly abundant once magnetite disappears. Euhedral, honey-yellow staurolites occur along with large flat pink garnets in thin schistose interbeds between the pure quartzite layers. The total section is thirty-five meters thick. The top of the Clough is marked by a tectonic melange of dark colored graphitic quartzite and black garnet-biotite schist similar to the 'Diamond Reef' in the Notch. The deformation here however is much more intense, and the layering is completely transposed. Similar graphitic zones are present at all outcrops of the Clough on the overturned limb of the syncline between here and Great Hill.

A homogeneous lineated plagioclase gneiss, the Glastonbury Gneiss, overlies the base of the Clough here. The contact is not visible, in fact to our knowledge it is not exposed anywhere in Connecticut. The Glastonbury outcrops on the small ridge just west of the quarry.

We will work our way up the northern quarry face from the pebbly, magnetite quartzite just 'above' the Glastonbury contact to the terrace near the top of the quarry where the black graphitic quartzite is exposed. The detailed stratigraphy is worth examining. Also of intrest are several kaolinitized pegmatite boudins within the quartzite. Feldspar kaolinization is common in pegmatites which cut the Clough all along the overturned limb of the Great Hill syncline. Is it a result of Tertiary weathering or is it hydrothermal? Quartz veins near the top of the quarry may contain coarse arsenopyrite and pyrite mineralization. One sample of scoroditzed massive arsenopyrite from here assayed 1.5 oz of gold per ton. The visible native gold at Great Hill has a similar parageneiss.

The view from the crest of the hill overlooking the quarry is spectacular. On a clear day the basalt ridges of the Hartford basin are plainly visible. The Sleeping Giant and the Hanging Hills of Meriden are on the southwest horizon and Avon and Talcott Mountains on the northwest.

From here follow the trail that leads north from the summit of the quartzite ridge. About 250 meters north the trail joins another that runs east-west along a fault-line valley. Follow this path eastward for 300 meters. Outcrops of Littleton Schist along the way contain 1 to 2 cm long anhedral staurolite porphyroblasts. These crystals are small-fry compared to those we seek. We are after much bigger game. The trail also passes small outcrops of various calc-silicates and muscovite-biotite gneisses. Are these exposures of the Fitch on the overturned limb or are they part of the Littleton? Staurolite schists outcrop on both sides of the calc-silicate rocks, but the Clough Quartzite is only a few hundred meters away. Leave the trail at its high point in a col and ascend the hill on its north side. The summit is marked '789' on the topographic quadrangle map. Staurolite porphyroblasts several centimeters long are remarkably abundant in outcrops on hill top. This band of giant anhedral staurolites can be traced the entire length of Box Hill. Its stratigraphic position and textural features are strikingly similar to the zone of equally giant staurolites below the painted pegmatite in Bolton Notch. However the shapes of the porphyroblasts are not quite identical. The staurolites in the zones on both limbs are distinctly anhedral but the crystals here are much less cylindrical. Triangular fan-shaped

porphyroblasts formed of radiating groupings of curved staurolite crystals are common. The orientations of the individual staurolite crystals vary systematically across a fan such that cross sections are sigmoidal and the whole grouping twists along its long axis, much like a propeller blade. The morphology of the porphyroblast aggregates is consistent with shear-induced surface breakage during growth. Small growing fragments break off and rotate independently until rigidly intergrown with the rest of the porphyroblast. Why on the eastern limb the rotating staurolites grew as single crystals while on the west they continually fractured is unclear. Perhaps the rate of deformation is a factor.

The orientation of the fan-shaped porphyroblasts show no strong preference for a particular direction. Bow-tie (i.e. a two isosceles triangles joined at their apex) crystals are rare. Those that do occur display a distinct curvature and as a result look more like twisted, swept-back wings than simple double-bladed propellers. These features are also consistent with shear induced surface breakage during growth.

On our way back we will examine muscovite pseudomorphs after staurolite in the aureole of a large pegmatite on the south side of the col near where we left the trail. In the same general area staurolite schists adjacent to large boudinaged quartz veins are converted to a banded tourmaline-quartz schist.

Rather than retracing our route along the trail, we will follow the chain of calc-silicate outcrops 150 meters south to a logging road which leads west to a private driveway on Box Mountain Road. Both the eastern and western contacts of the Fitch (?) with the Littleton Schist are exposed in outcrops along the logging road. Follow Box Mountain Drive west back to the Quarry.

Mileage
4.8 Return west on Box Mountain Drive.
5.4 Right onto Lake Street.
5.7 Stop sign. Turn right onto Tunnel Road.
6.5 Stop sign at a one lane tunnel. Proceed through tunnel but prepare to turn right at the first intersection.
6.7 Right onto Valley Falls Road.
7.8 Turn right at entrance to Valley Falls Park.

STOP 4. VALLEY FALLS PARK: LOZENGE-SHAPED STAUROLITE PORPHYROBLASTS
(45 MINUTES) Park in the public lot overlooking the lake. Follow the path southwest across the dam and small bridge along the west side of the lake. About halfway along the shore the trail splits. Follow the western branch up a steep slope across several outcrops of well-laminated, fine-grained grey quartzites and staurolite-garnet schists onto an old railway bed. Structurally these rocks are situated in the core of Great Hill syncline and represent the youngest Littleton exposed in the Bolton area.

Walk north along the railway bed. The first schist outcrop in the woods just to the west contains small euhedral stubby staurolite crystals and millimeter-sized garnet porphyroblasts. Continue north another fifty meters. Leave the railway bed and head to the outcrops along the steep slope to the west. The grey fine-grained micaceous quartzite in these outcrops splits easily into large thin sheets. The surfaces of some sheets are covered with large lozenge-shaped staurolite porphyroblasts. The crystals are typically no thicker than 1 mm, but range in size up to 30 cm in length and 10 cm in width. Crystals are randomly oriented within the bedding plane and all are boudinaged in a 'chocolate-block' fashion along both their long and short dimensions. Fractures are filled with muscovite. The flattening was apparently isotropic as no orientation dependence of the boudinage is evident.

As the porphyroblasts grew into each other with no change in width and no obvious 'zone of depletion', material transport could not have been rate limiting. Perhaps the extraordinary size of these crystals is related to an absence of appropriate nuclei.

Retrace our route back to the jeep trail which crosses the railbed near the junction of the path we followed from the lake. Follow the jeep trail north to Valley Falls Road. Just before this road the trail passes over a one meter thick schistose zone containing 3 to 4 centimeter size anhedral staurolite crystals. Below this zone the staurolites are euhedral and only 1 to 2 cm long. Why the size and morphology difference? Is it controlled by difference in the bulk composition or is it related to the intensity and style of deformation? The schists and granitic dikes just above the anhedral staurolite layer are amongst the most highly folded in the area.

Return east along Valley Falls Road to the parking lot.

Mileage
7.9 Right onto Valley Falls Road.
8.1 Stop sign. Turn right onto Vernon Road.
9.3 Vernon Road. Turn right into a private driveway at 797 Vernon Road.
This is a BLIND corner, BE CAREFUL.

STOP 5. FITCH & CO. FLAGSTONE QUARRY: THE FITCH FORMATION

(30 MINUTES) Beware, this is one of Connecticut's leading POISON IVY nurseries. Park on the grass on the north side of the driveway. Follow the dirt road west 120 meters along the edge of a hay field. Leave the road and bear south on a small but well worn trail which leads to the northern end of a large abandoned quarry. Although the area is completely overgrown (by poison ivy especially!) the extent of the exposed workings and the size of the tailings piles suggest something of the scale of the original operation. In the mid 1800s the non-calcareous muscovite-biotite gneiss of the Fitch Formation was quarried all along the Bolton ridge. By an odd coincidence this particular quarry was owned and operated by Patten Fitch (No relation to the Formation!).

By all accounts the muscovite-biotite gneiss of the lower Fitch was an ideal flagstone. It splits into unusually large thin slabs whose surfaces, according to Shepard (1837), are "as true and smooth as any granite or sandstone could be rendered by the nicest process of dressing". The stone was much in demand and was shipped to points as distant as New Orleans. Even today, the operating quarry at Bolton Notch can get an excellent price for large slabs of this muscovite-biotite gneiss. The quarries' decline in the late 1800s was not due to lack of demand but rather to the difficulty of obtaining sufficient material. The rocks dip into the hill and the quarrying was forced underground. In the early 1800s water was removed "by means of lead-syphon tubes of the largest size" (Shepard, 1837). Steam powered pumps were employed in later years but the underground workings were always dangerous places and collapses were common.

Today the pits are filled with water and only the upper calc-silicate portion of the Fitch Formation is well-exposed. Nonetheless, the penetrative planar foliation and the widely-spaced joints that made the Fitch an excellent source of large flagstones are still visible.

Mileage
9.3 Right onto Vernon Road
9.8 Bear left at this 'Y' junction. Vernon Street becomes Quarry Road.
10.4 Route 44 traffic lights. W.H. England Plaza on south side.
TRIP ENDS.

ACKNOWLEDGEMENTS

The Bolton Notch area is the traditional site of the University of Connecticut's Field Geology course. We thank the instructors, Tony Philpotts and Randy Steinen, who have been involved in teaching that course for over fifteen years, and are indebted to them for sharing with us their intimate knowledge of the geology of the area.

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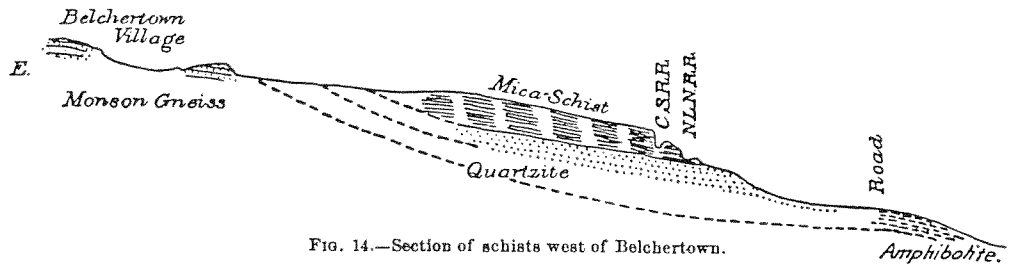

FIG. 14.—Section of schists west of Belchertown.

RECENT STUDIES IN THE ACADIAN METAMORPHIC HIGH, SOUTH-CENTRAL MASSACHUSETTS

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INTRODUCTION

The Acadian metamorphic high of southern New England contains the most intense Acadian metamorphism in the Appalachian orogen. Regional metamorphic zones for central Massachusetts, based on assemblages within pelitic schists, are shown in Figure 1 (Tracy, 1975, 1978; Tracy et al., 1976; Robinson et al., 1978, 1982, 1986a). A broad area in the middle of the metamorphic high is in Zone VI, characterized by the assemblage sillimanite + K-feldspar + garnet + cordierite. Metamorphosed igneous rocks in Zone VI include assemblages of orthopyroxene + augite + plagioclase and orthopyroxene + K-feldspar, indicative of the lower granulite facies (Hollocher, 1985). Calc-silicate rocks in this zone include the assemblages plagioclase + quartz + calcic scapolite (A.B. Thompson in Goldsmith and Newton, 1977) and wollastonite + anorthite (Berry, 1989, 1991; Robinson, 1991). Leucocratic layers in migmatitic gneisses represent a diverse suite of fluid-absent partial melts that formed over a range of P-T conditions (Robinson et al., 1982; Tracy and Dietsch, 1982; Thomson, 1992).

Recent field guides for this area give a more regional perspective of the high grade region (Robinson et al., 1982, 1986a, 1989). The present field trip will emphasize recent information without attempting to reiterate previous field guides. We will start out in lower grade rocks along the west margin of the granulite-facies high, where there is an apparent steep field gradient, and work eastward across the high, examining both pelitic and more unusual assemblages and their partial melts (Fig. 1).

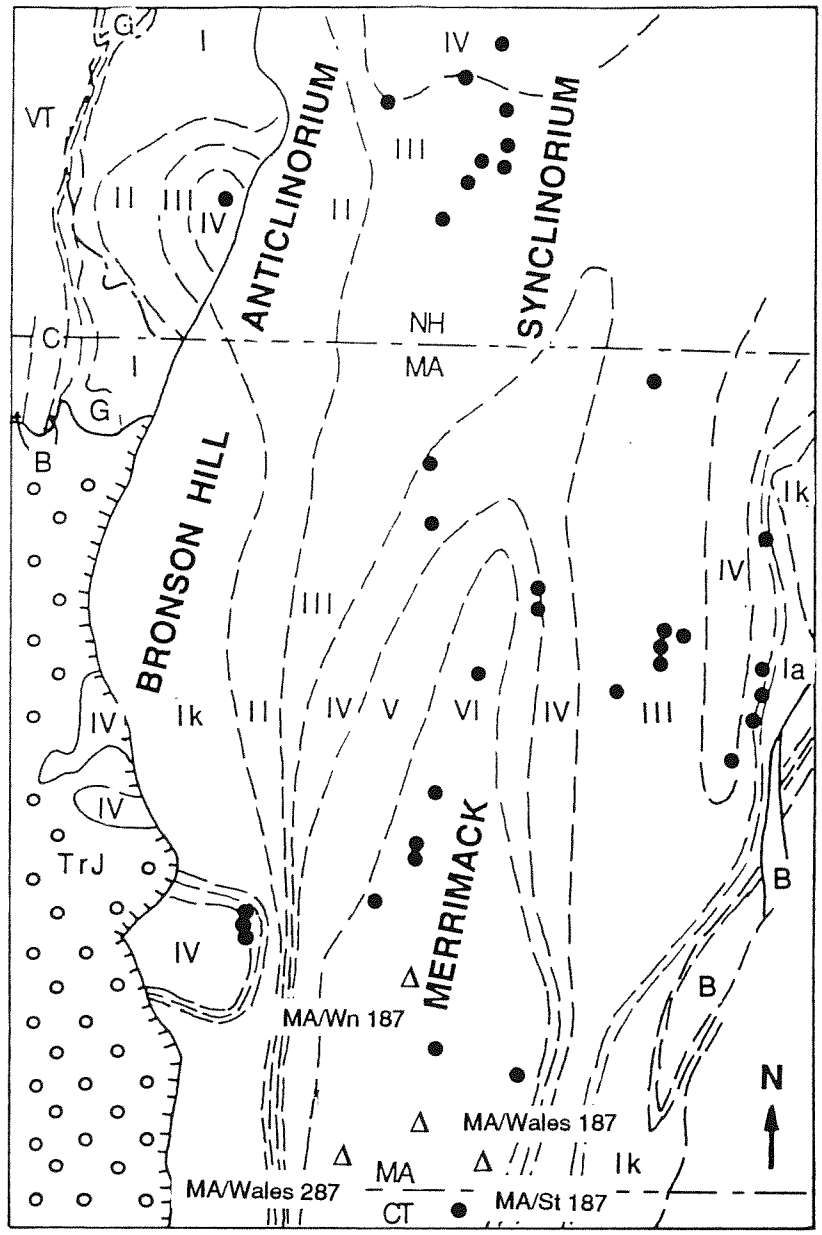
REGIONAL SETTING

The rocks visited on this trip lie within the Bronson Hill anticlinorium and Merrimack synclinorium (Fig. 1), two major tectonostratigraphic features of the New England Appalachians, regionally deformed and metamorphosed during the Devonian Acadian orogeny (Hall and Robinson, 1982). The Bronson Hill anticlinorium is partly defined by a series of structural culminations, exposing Late Proterozoic and Late Ordovician basement (Thompson et al., 1968; Tucker and Robinson, 1990). The Merrimack synclinorium contains highly deformed and metamorphosed Silurian - Devonian sedimentary rocks with minor volcanics (Hall and Robinson, 1982).

In central Massachussetts, the Acadian orogeny has been divided into three generalized stages of deformation: the nappe stage, backfold stage, and dome stage (Robinson, 1979; Berry, 1987, 1989). Westward transport of material during the nappe stage produced large-scale fold- and thrust-nappes resulting in an inverted metamorphic gradient in some places (Thompson et al., 1968; Robinson, 1979; Thompson, 1985; Elbert, 1988; Robinson et al., 1988, 1991). During the backfold stage, the tectonostratigraphy along the eastern margin of the Bronson Hill anticlinorium was overturned to the east, accompanied by emplacement of some gneiss bodies in the Bronson Hill anticlinorium (Robinson, 1979; Robinson and Hall, 1980; Robinson et al., 1982; Thompson, 1985; Peterson, 1992). Continued upward as well as lateral transport of basement gneisses during the culminating dome stage produced the final geometry of the gneiss domes of the Bronson Hill anticlinorium (Robinson, 1963, 1979).

METAMORPHISM

Differences in the character of metamorphic facies and in the metamorphic history are observed between the Merrimack synclinorium and Bronson Hill anticlinorium (Fig. 2). Within the Merrimack synclinorium in Massachusetts, high temperatures and moderate pressures were achieved through early high temperature/low pressure Buchan-style metamorphism followed by heating to peak thermal conditions and then compression during slow cooling (Schumacher et al., 1989, 1990; Robinson et al., 1986b, 1989).



Metamorphic Zones and Features:

VI	Garnet-Cordierite-Sillimanite-K-feldspar
V	Sillimanite-K-feldspar
IV	Sillimanite-Muscovite-K-feldspar
III	Sillimanite-Muscovite
II	Sillimanite-Staurolite
I_k	Kyanite-Staurolite
I_a	Andalusite-Staurolite
G	Garnet
B	Biotite
C	Chlorite
TrJ	Mesozoic Sedimentary and Volcanic Rocks
●	Sillimanite Pseudomorphs after Andalusite

Figure 1. Generalized map of metamorphic zones in central Massachusetts and southern New Hampshire based on pelitic rock assemblages (from Robinson et al., 1982). Sample numbers are localities of rocks that have been dated using U-Pb systematics (see text).

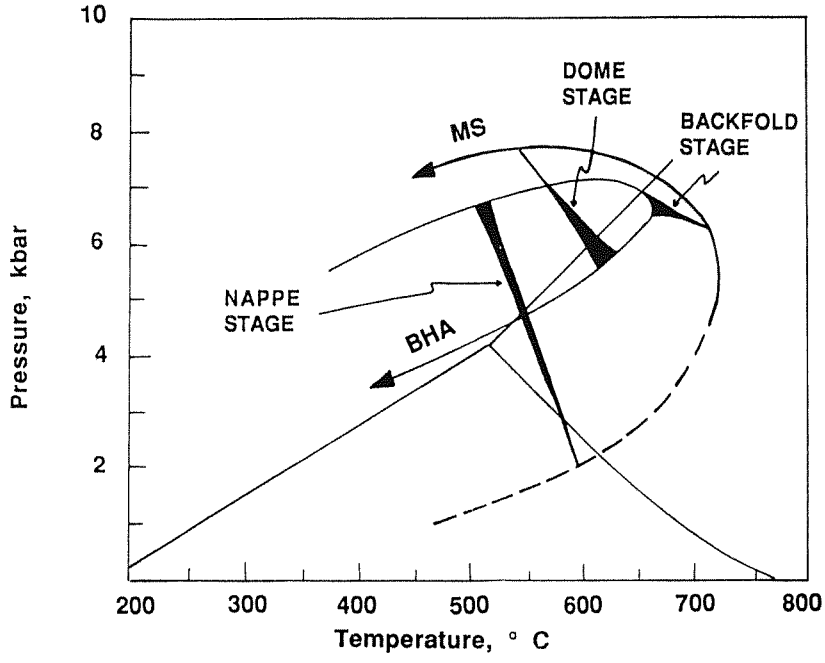


Figure 2. Integrated contrasting P-T paths for the Keene dome of the Bronson Hill anticlinorium (BHA) and Merrimack synclinorium (MS) (from Schumacher et al., 1989). Thick ends of tie lines lie on trajectory of the BHA; thin ends on trajectory of the MS.

Evidence for this counterclockwise P-T path has been found in independent petrologic investigations of pelitic rocks, calc-silicate rocks, partial melts, and fluid inclusion studies (Tracy and Dietsch, 1982; Robinson et al., 1982, 1986a, 1988, 1989; Schumacher et al., 1989; Berry, 1989, 1991; Thomson, 1989, 1992; Winslow et al., 1991). The metamorphic features preserved in Zone VI of the Merrimack synclinorium can be grouped informally into early metamorphic features, peak metamorphic features, and retrograde features. These features are summarized below, and in more detail later in the manuscript.

In contrast, rocks from the Bronson Hill anticlinorium appear to have followed a clockwise P-T path involving compression, followed by cooling and decompression (Robinson et al., 1982; Schumacher et al., 1989, 1990; Schumacher, 1990; Tracy et al., 1991), however, interpretation of the metamorphic history of this belt is complicated, however, by a Late Paleozoic metamorphic overprint in certain parts of the anticlinorium (Robinson and Tucker, 1991; Robinson et al., this volume).

Early Metamorphic Features

An early low-pressure high-temperature history that involved partial melting is suggested by the occurrence of sillimanite pseudomorphs after andalusite in pelitic schists and gneisses (Fig. 1) and by mineral reactions in cordierite±garnet-bearing pegmatites (Tracy and Dietsch, 1982; Thomson, 1992). Analysis of both melt- and solid-state mineral reactions in the pegmatites suggest that the cordierite formed earlier and at lower pressures than the peak of metamorphism (Thomson, 1992). Some pegmatites contain cordierite that has broken down to aggregates of garnet + sillimanite + quartz, including samples collected from STOP 6. A more complete discussion of cordierite±garnet-bearing pegmatites, including the melt reactions responsible for their genesis and the P-T conditions followed through the later stages of metamorphism is presented below and in Thomson (1992).

Peak Metamorphic Features

Garnets that belong to the prograde assemblages of Zone VI and parts of Zone V are characterized by homogeneous interiors with rims which differ in composition only where adjacent to other ferromagnesian minerals such as biotite or cordierite (Hess, 1969, 1971; Tracy et al., 1976; Robinson et al., 1982; Tracy, 1982). It has been suggested that prograde metamorphism and partial melting at temperatures of 675 to 730°C flushed the system of

metamorphic fluids so that later localized retrograde ion exchange reactions occurred in the absence of significant amounts of metamorphic fluid (Robinson et al., 1986a). Partial melting of pelitic schists and gneisses of appropriate bulk composition resulted in the formation of leucocratic garnetiferous melt segregations, the genesis of which is discussed below.

Peak granulite-facies metamorphic conditions of pelitic schists and gneisses have been estimated using a variety of thermobarometric techniques applied to garnet and cordierite core compositions and isolated matrix biotite. The calculated pressures and temperatures have been estimated at up to 740°C and 6.5 kbar (Robinson et al., 1986a, 1989; Thomson, 1989, 1992; Peterson and Thomson, 1991; Peterson, 1992). The univariant assemblage anorthite + wollastonite + grossular + quartz from a rock south of Holland, Massachusetts (Berry, 1989; Robinson, 1991) requires temperatures in excess of 730°C assuming 5 kbar, or in excess of 760°C assuming 6 kbar (Berman, 1988). The univariant assemblage corundum + garnet + sillimanite + spinel in a rock near the Holland-Sturbridge line in southern Massachusetts, provides a temperature estimate of 750°C assuming 5 kbar, or 800°C assuming 6 kbar (for further details, see STOP 7).

Retrograde Features

On the whole, peak metamorphic features in Zone VI are remarkably well preserved. Retrograde effects such as chloritized biotite or sericitized feldspar are rare as compared, for example, to rocks in lower grade zones. The most common post-peak retrograde effects appear to be ion exchange reactions that took place at lower temperatures, affecting compositions of garnet, biotite, and cordierite near mutual grain boundaries. (Tracy et al., 1976; Robinson et al., 1986a, 1989; Peterson and Thomson, 1991; Peterson, 1992; Thomson, 1992).

Petrographically discernable retrograde reaction textures in Zone VI include: fine-grained cummingtonite along orthopyroxene grain boundaries (Hollocher, 1985; Robinson et al., 1986a, 1989); pale green biotite on edges and fractures of cordierite and/or garnet near large K-feldspar grains within partial melts (Tracy and Dietsch, 1982; Robinson et al., 1989; Thomson, 1989, 1992); garnet + sillimanite + quartz aggregates entirely within large pegmatite cordierite grains (Tracy and Dietsch, 1982; Robinson et al., 1989; Thomson, 1989, 1992); muscovite-bearing pegmatites and granitic dikes (Robinson et al., 1989; Berry, 1989; Thomson, 1992; Peterson, 1992); and secondary epidote veins (Robinson et al., 1989). While most of the small-scale retrograde features are related to post-Acadian cooling, some features, particularly the latter two, could be late Paleozoic in age.

DISTRIBUTION OF METAMORPHIC FACIES ALONG THE WESTERN MARGIN OF THE GRANULITE FACIES HIGH

The Zone V - Zone VI boundary in south-central Massachusetts is marked by the first appearance of cordierite in garnet-biotite-sillimanite-orthoclase-bearing pelitic schists. Along the western margin of the Zone VI granulite-facies high, this boundary coincides approximately with the eastern margin of the Conant Brook shear zone, a zone of focused shear strain during the Acadian orogeny (see Peterson, this volume). The Conant Brook shear zone is a kilometer-wide zone, bounded to the west by the main body of Monson Gneiss (see Peterson, this volume). The Zone V pelitic schists discussed here are from within the shear zone. The nature of the Zone V - Zone VI transition in this area was evaluated through comparison of mineral compositions, zoning character and estimates of temperature conditions. The Ti-contents of biotites from Zone V assemblages within the shear zone are more similar to those of Zone VI than to previously analyzed samples from Zone V outside of the shear zone (Peterson, 1992, unpub. data). However, detailed comparisons show that there are distinct differences in mineral compositions between the shear zone rocks and Zone VI rocks. In Zone VI, biotites have a higher Mg/(Mg+Fe) content (Fig. 3) and garnets have a slightly higher pyrope and grossular content and lower spessartine content (Fig. 4). In addition, it appears that the cordierite-bearing assemblages of Zone VI are stabilized at a slightly lower μH_2O than those in Zone V within the shear zone (Fig. 5).

Detailed analysis of several garnets from samples collected throughout the shear zone shows that the pattern of garnet zoning is the same as that observed in garnets from Zone VI as described by Tracy et al. (1976). These garnets were homogenized at peak metamorphic conditions and show zoning due to retrograde re-equilibration only in the vicinity of biotite (Fig. 6). As noted above, this type of zoning has been interpreted to reflect a scarcity of intergranular fluid available to assist diffusion of species during retrograde re-equilibration of the rocks on cooling (Tracy et al., 1976; Tracy, 1982). Elsewhere in Zone V, the retrograde zoning is more continuous along thegarnet grain boundary, probably due to a greater availability of intergranular fluid (Tracy, 1982). Biotites also show evidence of compositional variation within a thin section. Within a given thin section, the composition of coarse

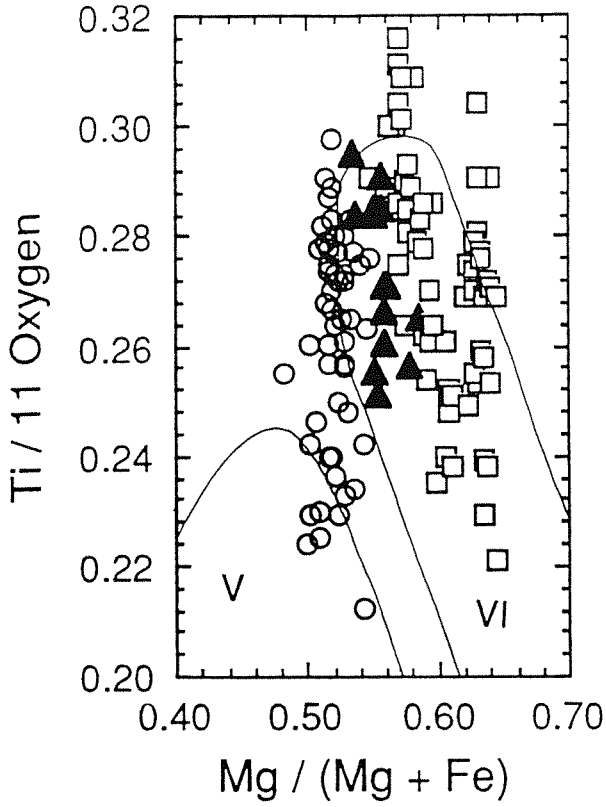


Figure 3. Comparison of atomic percent Ti vs. Mg/(Mg+Fe) in biotites from schists, saturated with Ti in the form of ilmenite or rutile and Al in the form of sillimanite or kyanite. This figure shows a range of biotite compositions for schist samples in Zones V and VI. Circles are from Zone V assemblages in the Conant Brook Shear Zone (CBSZ). Triangles are from a cordierite-free schist in Zone VI. Squares are from cordierite-bearing schists from Zone VI. Biotites from Zone V within the CBSZ have a distinctly lower Mg/(Mg+Fe) than those from Zone VI, but a similar range of Ti content. Compared with previously analyzed samples, indicated by outlined fields (from Robinson et al., 1982), they are more similar to the range of Zone VI than Zone V.

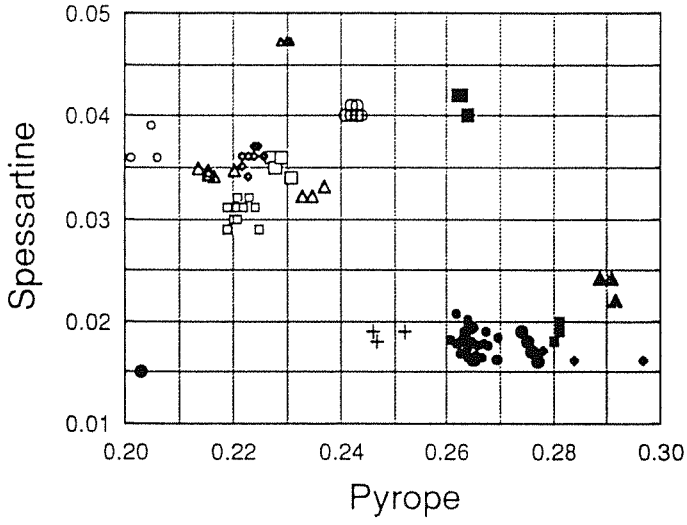


Figure 4. Plot of pyrope vs. spessartine content of garnets from analyzed samples in Zones V and VI. Hollow symbols represent garnets from Zone V assemblages and filled symbols represent garnets in cordierite-bearing Zone VI assemblages. Crosses are garnets from a cordierite-free sample from Zone VI. The pyrope content of Zone VI garnets is distinctly higher than Zone V garnets. The spessartine content of Zone V garnets is generally, but not consistently, higher than Zone VI garnets.

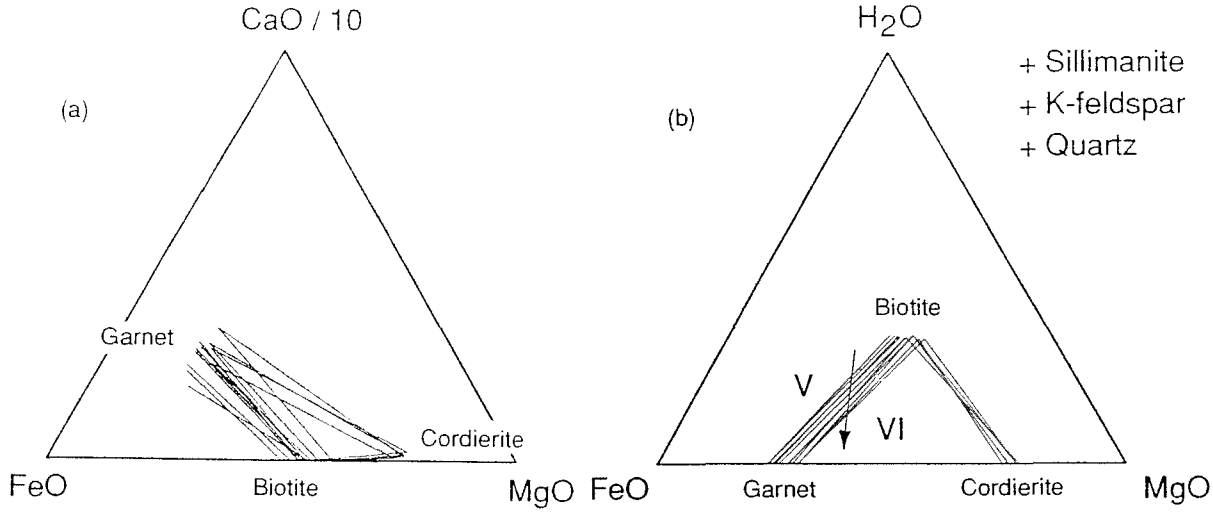


Figure 5. Comparison of assemblages from Zones V (Bt-Grt) and VI (Bt-Grt-Crd) samples. (A) FeO-MgO-CaO/10 comparing the grossular component of garnets from Zones V and VI. The CaO corner of the ternary is divided by 10 due to the low concentration of grossular in these samples. (B) FeO-MgO-H_2O projection showing assemblages from Zones V and VI. The water content in each biotite was assumed to be 1 H_2O molecule per formula. This projection does not account for octahedral vacancies in biotite, particularly due to Al and Ti substitution so that the water content appears much higher than it would in typical biotite. The apparent variations in H_2O shown on the projection are caused by slight variations in these substitutions. The arrow points in the direction of lower chemical potential of H_2O in the cordierite-bearing assemblages.

matrix biotite grains is consistently lower in Mg/Fe than analyses of biotite directly adjacent to garnet (Fig. 7). This is further evidence of inefficient intergranular diffusion within the matrix during cooling of these rocks.

In order to compare peak metamorphic temperatures across the Zone V - Zone VI boundary, calculations were made with several garnet-biotite thermometers (Peterson, 1992) using garnet core compositions and coarse matrix biotite, in some cases encased in large orthoclase grains. There is no apparent difference in temperatures calculated across the Zone V - VI boundary using the empirical thermometer of Thompson (1976) (Fig. 8) or other thermometers.

These observations may indicate that granulite facies conditions, typical of Zone VI, persisted west to the eastern margin of the Monson Gneiss and that the reaction that defines the Zone V - Zone VI assemblage boundary:

$$\text{Sil} + \text{Bt} = \text{Grt} + \text{Crd} + \text{Kfs} + H_2O \qquad (1)$$

may be more sensitive to subtle differences in bulk composition and μH_2O than to changes in temperature (mineral abbreviations after Kretz, 1983). It is also possible that a low μH_2O may be the factor or "extra component" that permits the existence of the four phase assemblage garnet-biotite-cordierite-sillimanite across the region of Zone VI. There may also be an intimate relationship between deformation within the Conant Brook shear zone and the location of the Zone VI isograd boundary. Focused deformation within the shear zone could have produced subtle differences in bulk composition and may have helped enhance fluid mobility. Alternatively, slight differences in composition and a slightly higher μH_2O may have helped to focus deformation within this zone.

Along the eastern margin of the Monson Gneiss are mafic and felsic gneisses of the Ammonoosuc Volcanics, as well as a few garnet-bearing felsic gneisses within the Monson Gneiss that provide some constraints on their conditions of metamorphism (STOP 1). These rocks lie a few hundred meters west of rocks of the Conant Brook shear zone (Peterson, this volume) which have yielded temperatures in the range of 700 - 750°C for peak metamorphic conditions. Equilibrium assemblages present include garnet-biotite-muscovite-sillimanite-K-feldspar in aluminous felsic gneisses and gedrite-cordierite-sillimanite-staurolite in altered mafic gneisses. The conditions suggested by these assemblages, as well as temperatures estimated at 600 - 650°C, are consistent with conditions for regional Zone IV (Sil-Mu-Kfs Zone). The apparent metamorphic discontinuity between the eastern Monson Gneiss - Ammonoosuc Volcanic package and the Conant Brook shear zone suggests the presence of a late or post-metamorphic fault . The presence of this fault, named the Quabbin Park fault, has not been substantiated in the field. Other supporting evidence is discussed by Peterson (1992).

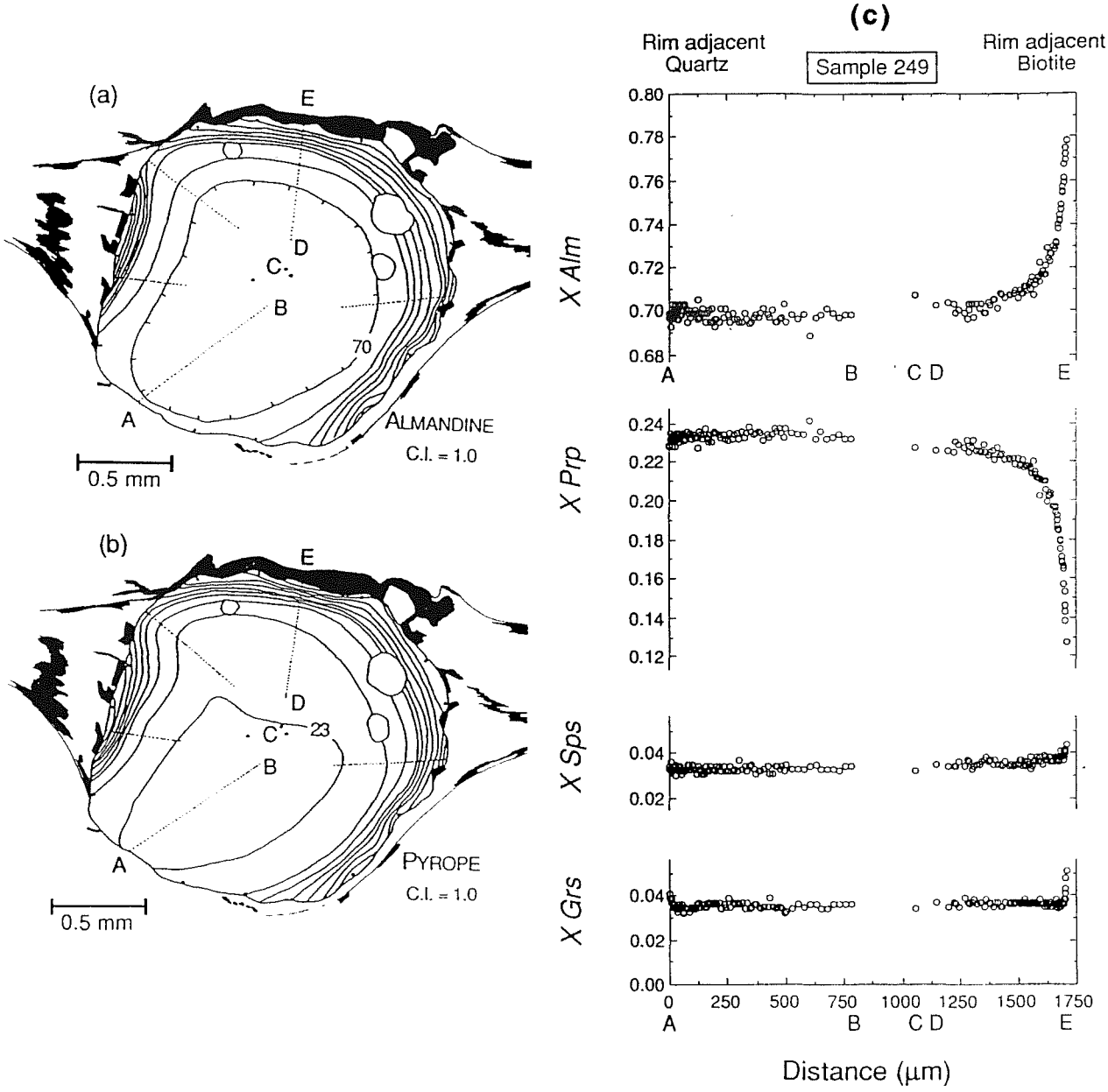


Figure 6. Contoured maps of almandine (A) and pyrope (B) contents (mole percent) of a garnet from the Conant Brook shear zone based on from traverse data (lines) and individual analyses (black dots). Areas of solid black show location of biotite grains. (C) Rim-to-rim compositional profiles from this garnet along the path A-B-C-D-E. Note homogeneous garnet interior, strong Mg and Fe zoning near biotite, and lack of zoning near quartz.

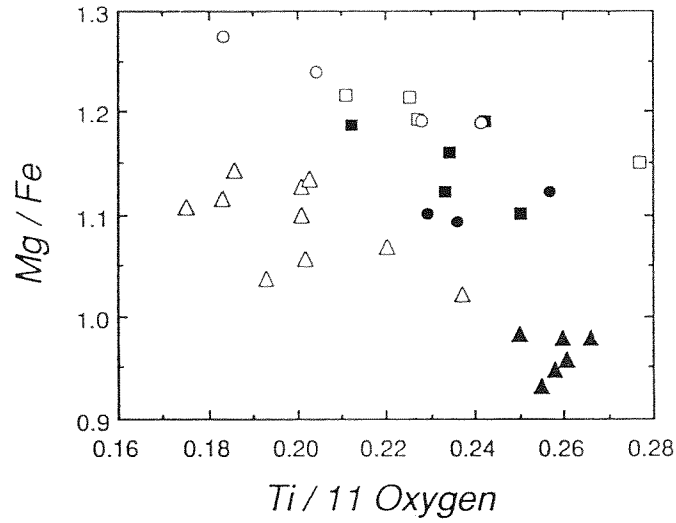


Figure 7. Comparison of matrix biotite compositions with those of biotite adjacent to garnet for three representative pelitic schist samples from Zone V within the CBSZ. This plot shows that biotite compositions are generally higher in Mg/Fe and lower in Ti adjacent to garnet. Matrix biotites = solid symbols; Adjacent biotite = hollow symbols.

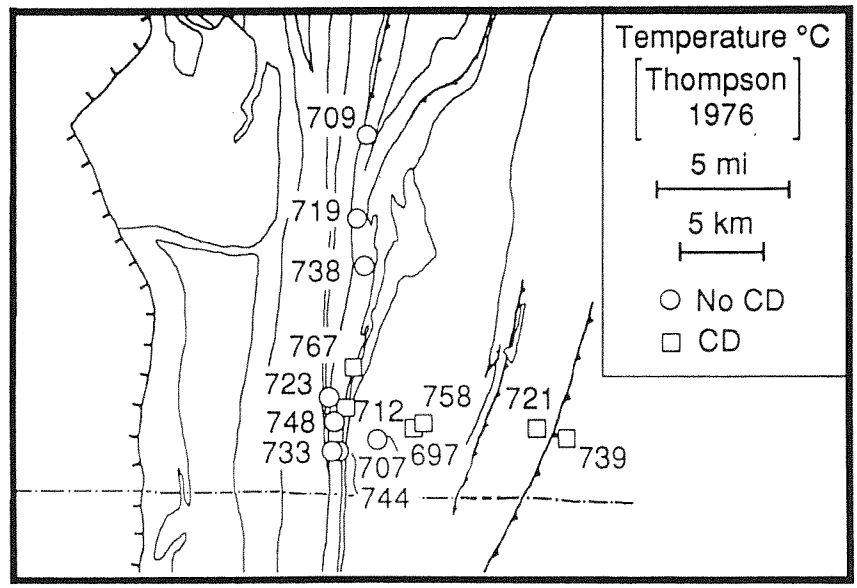


Figure 8. Map distribution of temperatures for pelitic schists of Zones V and VI based on the calibration of Thompson, 1976. Samples include those discussed by Peterson (1992) and Thomson (1992).

PARTIAL MELTING OF PELITIC SCHISTS AND GNEISSES

Evidence of partial melting of pelitic schists and gneisses lies in the presence of cordierite±garnet-bearing pegmatites and leucocratic garnetiferous melt segregations (vein-type migmatites) within the gneisses of the central Massachusetts metamorphic high. The pegmatites and garnet melt segregations are subparallel to parallel to the major foliation of their country rocks and have variable thickness (generally less 10 cm). Petrographic features indicate that they have been somewhat deformed. Despite the lack of primary igneous textures, due to subsequent metamorphism, the pegmatites and garnet melt segregations are interpreted to be the result of partial-melting of the surrounding host gneisses (Thomson, 1992). When compared to associated gneisses, the pegmatites and leucocratic melt segregations are coarser grained, typically with more abundant quartz and feldspar and less abundant biotite.

The pegmatites are primarily quartzofeldspathic rocks with varying amounts of cordierite and garnet, and minor amounts of biotite and sillimanite (Thomson, 1992). All share in common quartz, plagioclase, orthoclase, biotite,

sillimanite, ilmenite, and graphite and differ mainly in whether they contain both garnet and cordierite, cordierite only, or garnet only. Pyrrhotite and rutile also occur in some samples. The leucocratic melt segregations typically consist of euhedral garnet crystals 2 - 4 cm in diameter set within a coarse-grained matrix consisting of quartz, orthoclase, minor plagioclase, coarse sillimanite, little biotite, opaque phases (Ilm-Po-Gr±Ccp) and, in rare cases, cordierite.

Biotite within the pegmatites and melt segregations occurs in two main optical and textural varieties. Red-brown biotite occurs as inclusions in quartz, orthoclase, plagioclase, garnet and cordierite, and as isolated groundmass plates. Red-brown biotites in the pegmatites have X_{Mg} = 0.504 - 0.618 and Ti/11 oxygens = 0.088 - 0.276 whereas those in the melt segregations have X_{Mg} = 0.559 - 0.645 and Ti/11 oxygens = 0.235 - 0.309. Nearly all cordierite-bearing pegmatites contain pale green biotite (X_{Mg} = 0.562 - 0.647; Ti/11 oxygens = 0.000 - 0.045) intimately intergrown with fine-grained prismatic sillimanite and quartz on the edges of cordierite grains, typically in the vicinity of large K-feldspar grains. Similarly, all leucocratic melt segregations contain pale green biotite on garnet edges and fractures (not analyzed). Pale green biotite has not been observed in associated gneisses.

Cordierite grains within the pegmatites are up to 8 cm across although pinitized remnants of what were originally cordierite grains have been found up to 15 cm across (STOP 6). Cordierite grains in the pegmatites are typically homogeneous with X_{Fe} = 0.274 - 0.347. Cordierite grains in the melt segregations have X_{Fe} = 0.237 - 0.284, similar to cordierite grains in associated gneisses (X_{Fe} = 0.227 - 0.294). Local, very minor compositional deviations occur in cordierite only near grain boundaries and fractures which have been affected to a certain degree by pinitization and/or where partially rimmed by pale green biotite + sillimanite + quartz. Cordierite near these retrograde intergrowths has slightly lower X_{Fe} than cordierite not associated with this alteration.

In contrast to the homogeneous cordierite discussed above, some cordierite grains in several cordierite-garnet-bearing pegmatites have pronounced zoning. These samples have cordierite grains partially or entirely surrounding euhedral to subhedral garnet + sillimanite + quartz aggregates. The cordierite is strongly zoned, particularly within 1000 µm of the aggregate (Fig. 9a,b). Cordierite X_{Fe} far from the aggregate is typically 0.28 - 0.32, comparable to the X_{Fe} of homogeneous cordierite grains discussed above. The X_{Fe} of the cordierite decreases dramatically as the aggregates are approached and becomes as low as 0.16 directly adjacent to garnet (Fig. 9c). Such cordierite zoning features were also observed by Tracy and Dietsch (1982). These garnet + sillimanite + quartz aggregates within large cordierite grains have only been observed in pegmatite samples.

Isolated garnet grains within the pegmatites and melt segregations are homogeneous except where adjacent to biotite, cordierite, or Fe-Ti oxides. Garnet core compositions in the pegmatites range $Alm_{63-71}Prp_{24-29}Sps_{2-6}Grs_{3-4}$ whereas those in the melt segregations range $Alm_{65-71}Prp_{26-29}Sps_{2-5}Grs_{2-4}$ (mineral abbreviations after Kretz, 1983). In contrast to isolated garnet grains in the pegmatites, those that are present either partially or completely within large cordierite grains, show pronounced compositional zoning and, in general, have a higher pyrope content (Prp > 30). The garnet zoning is fairly complicated (Fig. 10) and suggests that both net transfer and ion exchange reactions operated to cause the observed zoning. There is no way to distinguish between the garnet-producing reaction and later ion exchange, nor to determine which reaction was more important in causing the observed zonation (Thomson, 1992).

Melt-forming Reactions

The cordierite±garnet-bearing pegmatites are interpreted to have been produced by biotite-consuming, melting reactions of the surrounding host gneisses, particularly fluid-absent or dehydration-melting reactions which generate melt in the absence of a fluid phase (Thomson, 1992). It is suggested that cordierite "phenocrysts" in the pegmatites grew in contact with a liquid generated by melting of gneisses with moderately magnesian bulk compositions. The major cordierite + melt-forming reaction responsible for the genesis of cordierite-bearing pegmatites, particularly those with cordierite only, is the fluid-absent biotite-melting reaction:

$$\text{Qtz} + \text{Als} + \text{Bt} = \text{Kfs} + \text{Crd} + \text{M}. \qquad (2)$$

Rocks with a slightly more Fe-rich bulk composition may have originally melted via Reaction (2) to form cordierite and melt, followed shortly thereafter by intersection with the biotite-melting reaction:

$$\text{Qtz} + \text{Als} + \text{Bt} = \text{Kfs} + \text{Crd} + \text{Grt} + \text{M} \qquad (3)$$

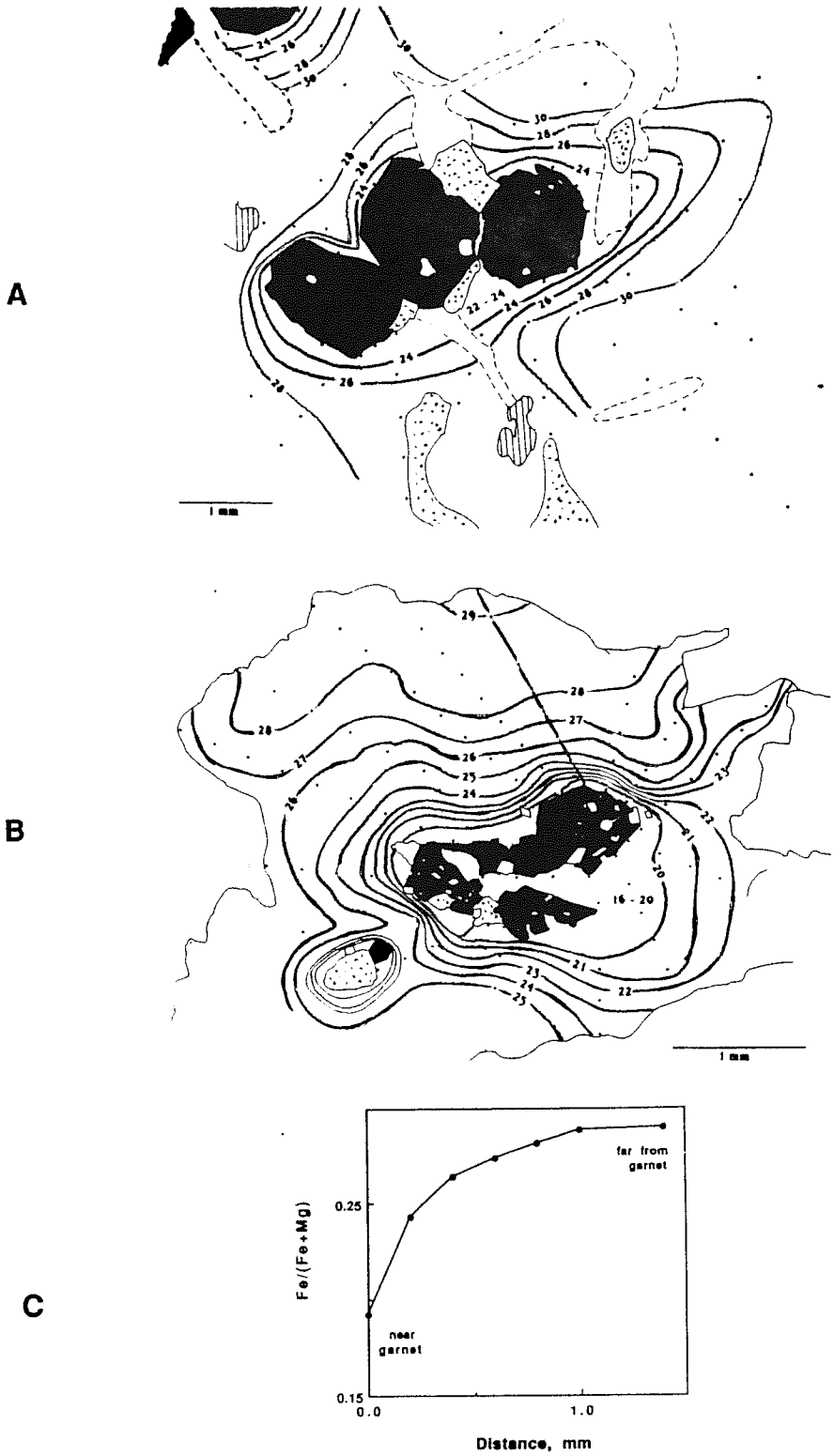


Figure 9. Concentric X_{Fe} contours in cordierite completely surrounding aggregate of Grt (black) + fine-grained Sil (enclosed by dashed lines in A; gray in B) + Qtz (stippled). Red-brown biotite inclusions are shown in striped pattern. Analysis points in cordierite shown as small black dots: (A) Sample Sturb2 collected from STOP 6; (B) Sample P11-1. Traverse of cordierite compositions along black line in (B) is shown in (C) (from Thomson, 1992).

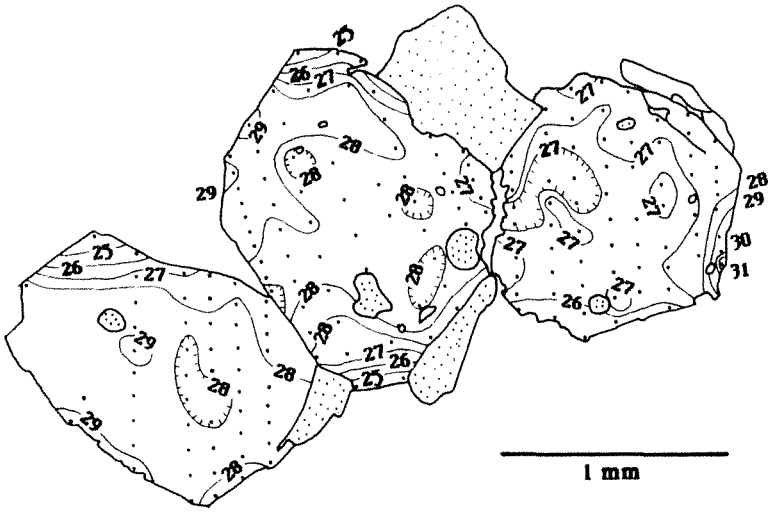


Figure 10. Contours of mole percent of pyrope in garnet in aggregate entirely within large cordierite from sample Sturb2 (STOP 6). Qtz (stippled), Crd, and Grt (unpatterned). Analysis points shown as small black dots. See Figure 9b.

to form garnet phenocrysts. These reactions are depicted on the T-X_{Mg} diagram in Figure 11.

The unusually Fe-rich cordierite compositions preserved in the pegmatites suggest that pegmatite genesis took place at lower pressures than the peak granulite-facies conditions recorded in the surrounding gneisses. Conditions of pegmatite genesis based on the core compositions of isolated garnet and cordierite are 710 - 800°C and 4 - 5 kbar estimated from garnet-cordierite thermometry (Thompson, 1976) and cordierite-sillimanite-garnet-quartz barometry (Bhattacharya, 1986). Peak metamorphic conditions, as recorded in the gneisses, are 710 - 740°C and 6 - 6.5 kbar based on garnet-biotite thermometry (Thompson, 1976) and cordierite-sillimanite-garnet-quartz barometry (Tracy et al., 1976).

The garnetiferous melt segregations are also thought to be the result of biotite-consuming, melting reactions of the surrounding host gneisses (Thomson, 1992). However, these melting reactions occurred at slightly higher pressures than those responsible for pegmatite genesis. Hence, the phase relations on the T-X_{Mg} diagram in Figure 12 have moved to higher X_{Mg}. The garnet "phenocrysts" in the leucocratic segregations are thought to have grown in contact with a liquid generated by melting of gneisses with a moderately iron-rich bulk composition. The major garnet + melt forming reaction responsible for the genesis of garnetiferous leucocratic melt segregations is the fluid-absent biotite-melting reaction:

$$\text{Qtz} + \text{Als} + \text{Bt} = \text{Kfs} + \text{Grt} + \text{M}. \qquad (4)$$

Rocks with a slightly more Mg-rich bulk composition may have originally melted via Reaction (4) to form garnet and melt, followed shortly thereafter by intersection with the fluid-absent biotite-melting reaction:

$$\text{Qtz} + \text{Als} + \text{Bt} = \text{Kfs} + \text{Crd} + \text{Grt} + \text{M} \qquad (3)$$

to form cordierite phenocrysts.

The fact that cordierite-bearing melt segregations contain cordierite with compositions similar to cordierite in surrounding host gneisses suggests that they formed under similar P-T conditions. Conditions of melt segregation genesis are 700 - 720°C and 6 kbar, based on garnet-biotite thermometry (Thompson, 1976) and cordierite-sillimanite-garnet-quartz barometry (Tracy et al., 1976).

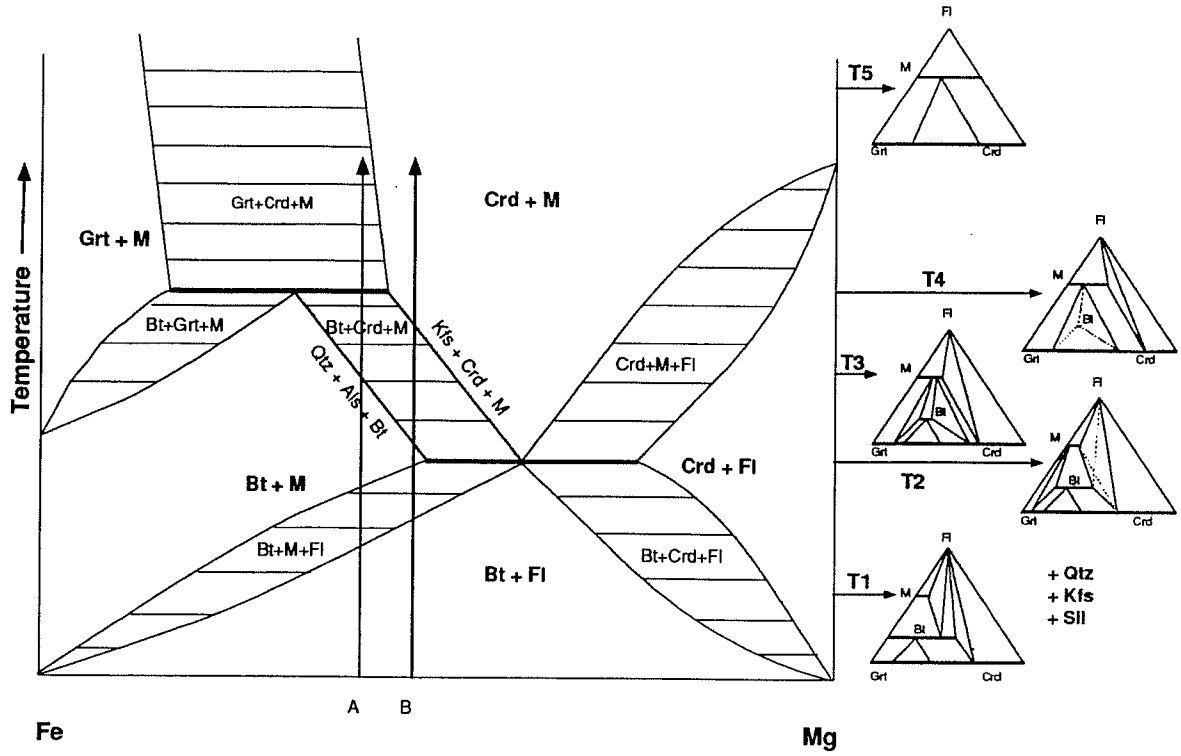


Figure 11. T-X_{Mg} diagram and compatibility (Fl-FeO-MgO) diagrams at a variety of temperatures, T1 - T5. Fl=fluid, M=melt. Arrows show two possible bulk compositions, A and B, which may have melted to generate Crd + Grt + Kfs + M (A) or Crd + Kfs + M (B) pegmatites (from Thomson, 1992).

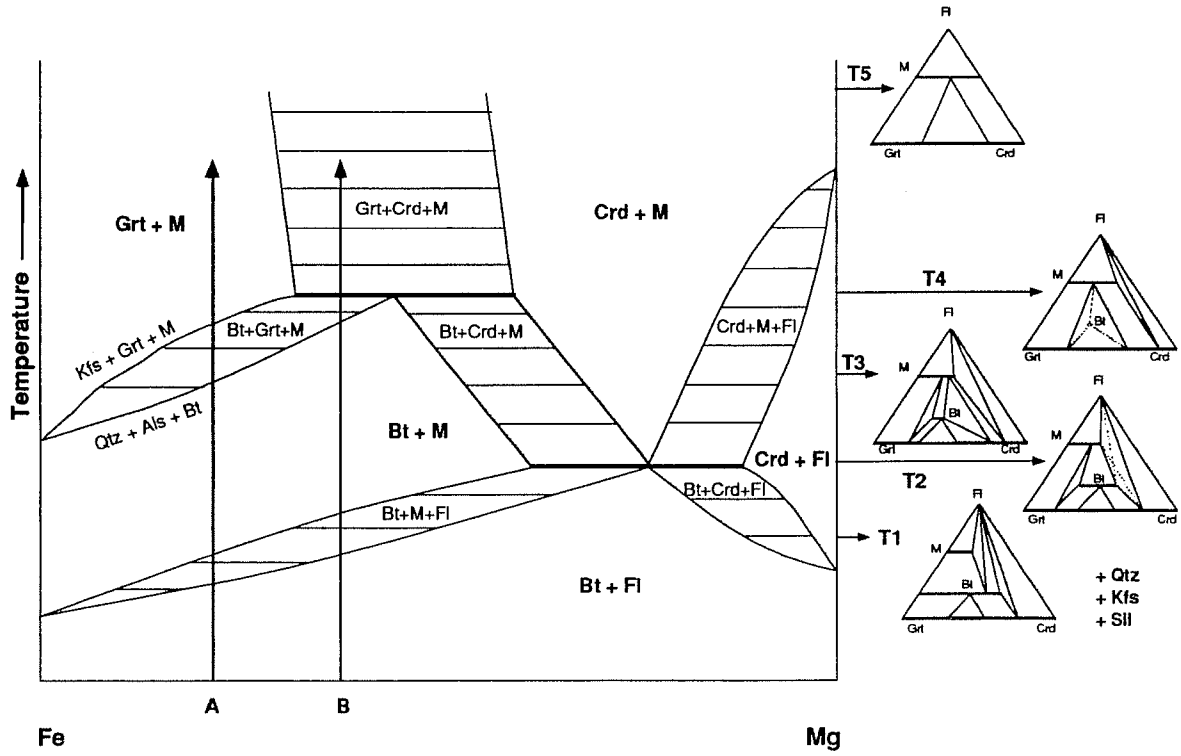


Figure 12. T-X_{Mg} diagram and compatibility (Fl-FeO-MgO) diagrams at a variety of temperatures, T1 - T5. Fl=fluid, M=melt. Arrows show two possible bulk compositions, A and B, which may have melted to generate Grt + Kfs + M (A) or Grt + Crd + Kfs + M (B) melt segregations (from Thomson, 1992).

Retrograde Reactions and History

Petrographic observations reveal at least two retrograde reaction textures in the pegmatites and one in the leucocratic melt segregations. The most common reaction texture observed in all cordierite-bearing pegmatites is the partial rimming of cordierite by fine-grained intergrowths of pale green biotite + sillimanite + quartz on the edges and fractures of cordierite grains, generally in the vicinity of large K-feldspar grains. Similarly, garnets in the melt segregations are partially replaced by pale green biotite on grain edges and fractures. The reactions responsible for the formation of these replacement textures are:

$$\text{Crd} + \text{Kfs} + H_2O\ (M?) = \text{Bt} + \text{Als} + \text{Qtz} \qquad (5)$$

and

$$\text{Grt} + \text{Kfs} + H_2O\ (M?) = \text{Bt} + \text{Als} + \text{Qtz} \qquad (6)$$

respectively. Both of these reactions are essentially the retrograde equivalents of the biotite-consuming melting Reactions (2) and (4), respectively, responsible for melt genesis. These reactions and the resulting textures may have resulted by either reaction between cooling melt and crystallizing minerals (Ashworth, 1976; Corbett and Phillips, 1981; Ashworth and McLellan, 1985; Stevens and Van Reenan, 1992) or by a very localized influx of externally-derived water during retrogression. It could be that late aqueous fluids followed structural channelways confined in the pegmatites and melt segregations and did not affect cordierite or garnet in the surrounding gneisses. The formation of late, retrograde assemblages by high temperature, structurally channeled fluids has been suggested by Waters and Moore (1985), Waters (1988), and Stevens and Van Reenan (1992). As the melts crystallize, the residual melt becomes water saturated and aqueous fluids may react with crystallizing phases. The fact that these replacement textures are not observed in associated gneisses suggests that aqueous fluids were confined within the pegmatites and leucocratic garnetiferous melt segregations.

The second reaction observed in some pegmatites involves the formation of aggregates of subhedral to euhedral garnet + sillimanite + quartz either partially or entirely within large cordierite grains. The reaction responsible for the formation of these textures is the breakdown of cordierite via:

$$3\ \text{Crd} = 2\ \text{Grt} + 4\ \text{Sil} + 5\ \text{Qtz}. \qquad (7)$$

The Fe-rich pegmatite cordierite formed at lower pressures than cordierite found in the adjacent gneisses. Thus, it is expected that early formed, Fe-rich cordierite would break down to garnet + sillimanite + quartz at higher pressures. Conditions of the "beginning" and "end" of cordierite breakdown to garnet + sillimanite + quartz were calculated using the garnet-cordierite thermometer of Thompson (1976) and the cordierite-sillimanite-garnet-quartz barometer of Bhattacharya (1986). Conditions of the onset of cordierite breakdown, calculated from aggregate garnet "core" and cordierite far from the aggregate, are 760 - 810 °C and 5 - 5.5 kbar. The last recorded conditions of cordierite breakdown, estimated from aggregate garnet rim and adjacent cordierite mineral pairs, are from 570 - 600 °C and 6 - 7 kbar. The results suggest that the cordierite pegmatites record part of a P-T path in which compression with heating appears to have been followed by further compression with cooling (Fig. 13).

The last recorded conditions of cordierite breakdown (570 - 600°C, 6 - 7 kbar) lie within the kyanite field on a P-T diagram (Fig. 13). Only sillimanite has been found associated with the aggregates within the pegmatite samples. One gneiss sample, collected by John T. Cheney, contains both kyanite and staurolite, suggesting that the rocks may have crossed the sillimanite-kyanite transition. The new geobarometer calibrated by Mukopadhyay and Holdaway (1991) may place these P-T points in the sillimanite field, resulting in a more isobaric cooling path. Calculations have not been done at the time of this writing.

TIMING OF METAMORPHISM WITH RESPECT TO DEFORMATION

Berry (1989) has suggested that peak granulite-facies metamorphism followed the nappe-stage of Acadian deformation in the Brimfield-Sturbridge area of south-central Massachusetts. This conclusion is based on the fact that Acadian tonalites, which cross-cut the fold- and thrust-nappes, have been affected by the peak of granulite-facies metamorphism.

Two phases of shearing observed within the Conant Brook shear zone have been tied to the regional backfold- and dome-stages of deformation. The two phases of shearing produced fabrics that are virtually identical, distinguished primarily by the orientation of their associated stretching lineations, which are defined in part by coarse

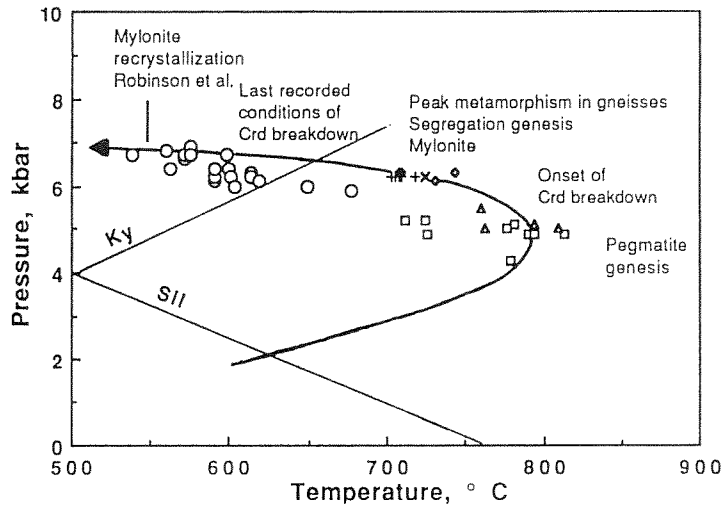


Figure 13. Counterclockwise P-T path of the Merrimack synclinorium based on the results of Thomson (1992). Squares=pegmatite genesis; triangles=onset of Crd breakdown; diamonds=peak conditions in gneisses; crosses=melt segregation genesis; X=mylonite sample; circles=last conditions of Crd breakdown; bar at 550°C, 7 - 8 kbar=conditions of mylonite recrystallization (Robinson et al., 1986a). Aluminosilicate triple point after Holdaway (1971).

prismatic sillimanite (Peterson, this volume). Textural observations indicate that peak metamorphic minerals make up the shear fabrics, so that both phases of shearing took place after the peak of metamorphism. The predominance of dynamic over static recrystallization textures may also reflect the continuation of deformation beyond the metamorphic peak. The absence of late metamorphic fluids during deformation may have been important in preventing both extensive retrograde reactions during shearing and extensive annealing after deformation ceased.

Some observations suggest that shearing deformation took place at relatively high temperatures. The dominant deformation mechanism in quartz and feldspar within these rocks is dislocation creep (Peterson and Robinson, in press). The mechanism of dislocation creep is active in naturally deformed feldspars at temperatures above 450 or 500°C (Tullis, 1983). The extensive recrystallization of feldspar under relatively dry conditions may suggest conditions of deformation that were comfortably above this limit. The apparent activity of prism slip mechanisms in quartz from these rocks (Peterson and Robinson, in press) gives further evidence for high temperature deformation conditions. Ave'Lallement and Carter (1971) showed that under laboratory conditions, prism slip is only active at temperatures greater than 800°C. Lister and Dornsiepen (1982) suggest that this transition to prism slip under natural conditions takes place between 600 -700°C at 6 kbar.

The textural similarity between the two lineations and associated mylonitic fabrics suggests that the metamorphic conditions during both phases of deformation were essentially the same and possibly near peak temperature conditions for this area. This may indicate that the rocks in this zone remained near the peak of metamorphism for a relatively long time, or that the time between the two phases of deformation was relatively short.

The steep metamorphic field gradient observed along the western margin of the granulite facies high in this regions may result in part from backfold-stage overturning of an inverted metamorphic gradient, produced during the nappe stage. This effect may be complicated by the late or post metamorphic Quabbin Park fault.

MONAZITE U-Pb DATING OF ZONE VI METAMORPHISM

Samples

Monazites have been separated and analyzed from four Zone VI metamorphic rocks. The sample localities are indicated in Figure 1. Three of the localities form a traverse along our trip route, with one toward the western limit

of Zone VI, one in the heart of Zone VI, and one toward the eastern limit of Zone VI. The fourth sample is from the central region of Zone VI about 8 miles north of our route.

The samples were collected from roadcuts or freshly blasted bedrock at localities where the metamorphism is well characterized. All the samples are from pelitic schists interpreted to be metamorphosed Silurian sediments. The fact that monazite is unable to survive greenschist facies metamorphism (Smith and Barreiro, 1990) implies that no detrital monazites are preserved, and that the monazites in these rocks formed during high-grade metamorphism.

We will visit two of the localities from which monazites have been analyzed (sample MA/Wales 287 at STOP 4, and sample MA/St 187 at STOP 6). Rock descriptions, mineral chemistry, and petrologic interpretations are described below for these stops. Between STOPS 4 and 5, as indicated in the road log, we will drive past the outcrop from which sample MA/Wales 187 was collected. An outcrop description and mineral analyses for this locality are given by Robinson et al. (1982, their STOP 8). The fourth monazite sample (MA/Wn 187) is from a roadcut on the southeast side of Rt. 67 north of the town of Warren in the Warren, MA quadrangle. This locality was mapped and described by Mike Field (1975), the garnet zoning was studied by Tracy et al. (1976, sample FW407), and lithologic and petrologic descriptions are given by Robinson et al. (1982, their STOP 5).

Based on these previous studies and the new data reported here, the peak metamorphic conditions experienced by the monazite samples can be estimated as follows:

Sample	Peak temperature (°C)	Method, Reference
MA/Wales 187	710	Gar-Biot, Thomson (1992), this paper STOP 4
MA/Wales 287	745	Gar-Biot, Robinson et al. (1982)
MA/St 187	760	Gar-Cord, Thomson (1992), this paper STOP 6
MA/Wn 187	680	Gar-Biot, Tracy et al. (1976)

These temperature estimates are internally consistent, based on the calibrations of Thompson (1976), and are given here to indicate the relative intensity of peak metamorphism among the samples. Other calibrations may give different absolute temperatures.

Methods

Two or three fractions were obtained from each of the four samples. Each fraction was dissolved in bombs in HCl, aliquoted, and spiked with a 208Pb - 235U solution. Pb and U were separated using standard anion exchange techniques. Pb and U blanks were below 100 pg. Isotopic ratios were measured on a VG Sector mass spectrometer, and corrected for fractionation by comparison to NBS 981 and U500 standards. Ages have been calculated using decay constants of Steiger and Jager (1977).

Results

Results are shown in Table 1. Most analyses show reverse discordance typical of monazites; that is, the 206Pb*/238U age is slightly greater than the 207Pb*/235U age. This discordance is attributed to excess radiogenic 206Pb (206Pb*) derived from 230Th, and therefore the 207Pb*/235U is considered to be the most reliable age.

The most concordant and reproducible analyses are from MA/Wales 287. We take its age to be 363±1 Ma. Sample MA/Wales 187 shows slightly more discordance, and the third fraction did not reproduce the age of the first two fractions. We assign it a "best guess" age of 367±2 Ma based on the first two analyses. For the northern sample, MA/Wn 187, the two fractions do not agree very well, suggesting a mixed age. Both fractions from this sample give ages older than the first two samples, suggesting that at least some of the monazites in MA/Wn 187 are distinctly older than the apparently homogeneous monazite population in the first two samples.

For the eastern sample, MA/St 187, the three fractions did not reproduce well, again suggesting the possibility of a mixture of different-aged monazites. But in contrast to MA/Wn 187, the ages for all fractions of MA/St 187 are distinctly younger than the first two samples. The fact that the first fraction from MA/St 187 is concordant may indicate that 345 is a geologically meaningful age, but the data from the other two fractions cast doubt on any simple interpretation.

Table 1. Monazite data from metamorphic Zone VI in south-central Massachusetts.

Sample	206Pb*/238U	207Pb*/235U	207Pb*/206Pb*	206Pb*/238U age (Ma)	207Pb*/235U age (Ma)	207Pb*/206Pb* age (Ma)	Best age
MA/Wales 287	0.05780	0.4291	0.05384	362.2	362.5	364.2	
MA/Wales 287	0.05795	0.4299	0.05380	363.2	363.1	363.0	
MA/Wales 287							363±1
MA/Wales 187	0.05854	0.4332	0.05367	366.8	365.5	357.3	
MA/Wales 187	0.05896	0.4367	0.05372	369.3	367.9	359.2	
MA/Wales 187	0.05800	0.4257	0.05323	363.5	360.1	338.0	
MA/Wales 187							367±2
MA/Wn 187	0.05924	0.4408	0.05397	371.0	370.8	369.8	
MA/Wn 187	0.06069	0.4531	0.05415	379.8	379.5	377.4	
MA/Wn 187							"375±5"
MA/St 187	0.05497	0.4041	0.05332	344.9	344.6	342.6	
MA/St 187				354.2	351.5	333.4	
MA/St 187				351.3	343.4	290.2	
MA/St 187							"345"

Interpretation

On the basis of petrologic continuity, we believe that the peak metamorphism in Zone VI was produced in a single event. The concordant ages from MA/Wales 187 and MA/Wales 287 show internal consistency and close agreement between samples. From this, we take the age of peak metamorphism in Zone VI to have occurred at 362 - 369 Ma. These ages are the same as peak metamorphism to the north in central Massachusetts (see Robinson et al., this volume), and the age of M3 metamorphism and associated plutonism in western Maine (Smith and Barreiro, 1990).

The interpretation of MA/Wn 187 is less clear, but assuming that the Zone VI metamorphism there was also about 362 - 369 Ma, then it may be true that some of the monazites in MA/Wn 187 formed earlier than the peak metamorphism. The petrology indicates that there was a regional metamorphic event in this region that preceeded the peak metamorphism. It might be that, because peak temperatures were lower here than for the other samples, some of the earlier monazites have survived. The regional andalusite-forming event in western Maine (M2), that is similar in many ways to the early event in central Massachusetts, has been dated at around 400 Ma (Smith and Barreiro, 1991).

Similarly, if the peak metamorphism at MA/St 187 is taken to be about 363 - 369 Ma, then the monazite data suggest that a subsequent event may have affected these rocks. While the data from this one sample are not sufficient to indicate two separate events, let alone what their ages might be, the possibility of a late Paleozoic overprint on Acadian metamorphism could account for the apparent age of 345 Ma.

ACKNOWLEDGEMENTS

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ROAD LOG

Assemble at the UMASS stadium parking lot at 8:00 AM. We will give a brief introduction, then hit the road at 8:30 AM. The road log begins at the intersection of Rte 9 and 202 in Belchertown, MA (9 miles from the center of Amherst).

Mileage

0.0 Head south on Rte 202 to Belchertown center.
0.9 Traffic light, intersection with Rte 181. Go straight through intersection on Rte 181 south.
1.2 Sign for Three Rivers. Bear left, continuing on Rte 181.
6.5 Rte 181 makes sharp left into Bondsville.
6.8 Center of Bondsville, T intersection. Turn right.
7.0 T intersection at church. Turn right, continuing on Rte 181.
7.6 Angle left at sign for Thorndyke (Thorndyke Street).
9.2 Crossing western contact of Monson Gneiss with Ammonoosuc Volcanics.
9.9 Intersection with Rte 32. Bear right on Rte 32.
10.3 Traffic light at shopping plaza. Continue straight.
10.7 Pass Entrance to Massachusetts turnpike.
11.3 Traffic light, junction Rte 20. Continue straight.
11.5 Traffic light, center of Palmer. Turn left and follow Rte 32 south through the town of Monson MA.
12.2 Note huge sandpit to left.
16.0 Blinking light, Monson MA., bear left (continue on Rte 32)
17.6 Angle left onto Cedar Swamp Rd.
18.1 Monson sand and gravel pit.
18.5 Cross Tenneco Gas Corporation pipeline.
19.3 Peck Hill Rd to right. Bear left, continuing on Cedar Swamp Rd.
19.8 Park on the right side of the road, just before intersection with Ayres Rd on the left.
(Driving time to stop ≈ 35 MINUTES)

STOP 1: COARSE GEDRITE-CORDIERITE-SILLIMANITE-STAUROLITE GNEISS, AMMONOOSUC VOLCANICS (30 MINUTES) (Monson Quadrangle). There are two outcrops here, a small roadcut to the east and a larger outcrop below the road to the west. We will look primarily at this larger outcrop, but please restrict hammering to abundant loose pieces or the smaller roadcut.

These outcrops are among the few exposures of the Ammonoosuc volcanics found along the eastern margin of the main body of Monson Gneiss. The contact with the Monson Gneiss lies within 50 meters to the west and is constrained by outcrops within a few meters on Peck Hill. The exposures just to the southwest on Peck Hill are dominated by orthoamphibole-bearing felsic gneisses.

This rock is a heterogeneous, coarse-grained mafic gneiss, dominated by radiating sprays, several centimeters long, of brown gedrite. The matrix contains cordierite, quartz, sillimanite, biotite, staurolite, ilmenite, rutile, and sulfides. Locally, gedrite grains are smaller and matrix minerals, particularly cordierite and quartz dominate the rock. An east-facing wall in one part of the outcrop is dominated by coarse blue cordierite and quartz. An isolated patch of garnet is present near the top of the outcrop. Garnet and gedrite are found to coexist without cordierite, sillimanite, or staurolite in an outcrop up the hill. A block of this garnet-bearing rock can be found here, near the base of the outcrop. The heterogeneity in assemblages or in proportions of the different phases within an assemblage suggests that there are distinct bulk compositional differences across the outcrop.

The presence of gedrite, sillimanite, staurolite, and cordierite and absence of garnet and plagioclase in the dominant rock in this outcrop place constraints on the P-T location of this assemblage on the petrogenetic grid of Spear and Rumble (1986) (Fig. 14). Estimated conditions from this grid are ≈ 600 - 630°C and 4 - 5 kbar. Using a calibration of Ti in orthoamphiboles from this region (Schumacher et al., 1990), temperature estimates are ≈ 620°C. Garnet-biotite temperatures estimated for two felsic gneisses from the Ammonoosuc Volcanics and Monson Gneiss, collected just northwest on Peck Hill, are between 640 and 650°C. One of these gneisses contains the assemblage garnet-biotite-sillimanite-muscovite-quartz-K-feldspar, which is typical of schists from Zone IV. The temperature estimates for these rocks are consistent with or slightly low for Zone IV.

The four phase assemblage in this rock, gedrite-sillimanite-staurolite-cordierite, is univariant in P-T space (Figs. 14 and 15). Compositional data and textures observed in thin section provide clues to the possible reaction preserved in this rock. Staurolite and sillimanite both appear to be reaction products. Staurolite grains have euhedral boundaries and appear to be spatially associated with ilmenite and rutile. Thick mats of fibrolitic sillimanite are ubiquitous and in many places appear to overgrow gedrite. In some places, coarse sprays of gedrite surround cordierite and appear to have grown quickly, whereas in other parts of the rock, gedrite appears to be breaking down in favor of sillimanite. Some gedrite grains are zoned toward lower Na and Al and higher Mg and Fe with a very small increase in the Mg/Fe ratio from core to rim. Throughout the sample, there are also relatively homogeneous grains with a range of composition among them that is the same as that of the individual zoned grains. It appears that Na-Al-richer, Fe-Mg-poorer gedrites were breaking down in some places and Na-Al-poorer, Fe-Mg-richer gedrites were growing in others. The zoned grains may reflect local disequilibrium.

Peterson (1992) has suggested a possible reaction for this rock:

Na-Al-richer gedrite + cordierite + ilmenite + H_2O =>
Na-Al-poorer gedrite + staurolite + sillimanite + rutile + quartz. (8)

The cordierite balanced in this reaction is anhydrous, so that water can be considered either as a reactant phase or as a component of cordierite. Na is balanced by a greater abundance of Na-poorer gedrite in the product. Ilmenite appears to supply much of the necessary Fe to form staurolite and it is possible that the reaction is driven by an ilmenite to rutile reaction or crossing of an ilmenite-rutile fence. A first approximation of the direction and slope of this reaction in P-T space (Fig. 14) was made using a possible balanced reaction and the thermodynamic database of Holland and Powell (1990).

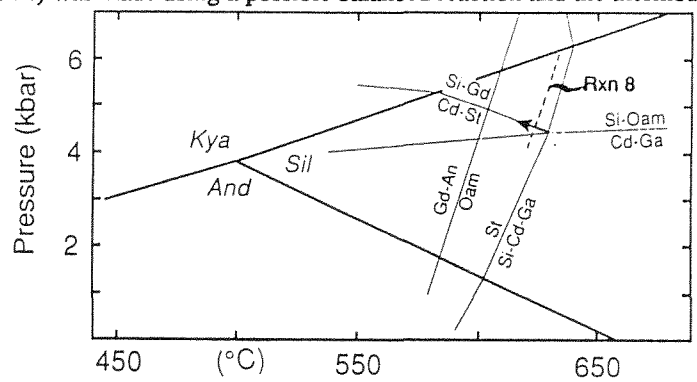

Figure 14. Selected reactions from petrogenetic grid of Spear and Rumble (1986). The assemblage described in sample 272 lies within the sillimanite field, above the orthoamphibole solvus, above the Cd + Ga = Si + Oam reaction and probably near the Si-Gd-Cd-St univariant. Possible reaction, Cd + Gd => Si + St + Gd shown as Rxn 8 (dashed line) may proceed in the direction of the arrow.

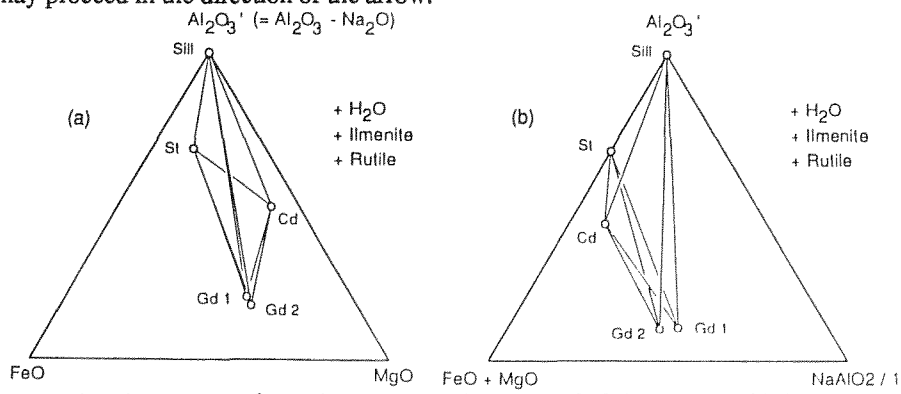

Figure 15. Diagrams showing nature of reaction observed in sample 272 (Peterson, 1992). Data points in (A) and (B) are representative microprobe analyses of the reactant and product phases. Sill = sillimanite, St = staurolite, Cd = cordierite, Gd1 = gedrite reactant, Gd2 = gedrite product. (A) Al_2O_3' - FeO - MgO projection shows small change in Mg/Fe ratio in gedrite. (B) Al_2O_3' - MgO+FeO - $NaAlO_2$/10 projection shows change in Na content of gedrite. Both (A) and (B) show a decrease in the abundance of cordierite and increase in abundance of staurolite as reaction proceeds.

Take an immediate left from Cedar Swamp Rd onto Ayres Rd.
20.7 Turn left at T intersection onto Moulton Rd.
21.7 Take right fork in the road.
21.9 Turn left onto Wales Rd.
22.1 Park on right side of the road, just beyond the outcrop.
(Driving time to stop ≈ 10 MINUTES)

STOP 2: "TYPICAL" ZONE V ASSEMBLAGE FROM CONANT BROOK SHEAR ZONE (30 MINUTES) (Monson Quadrangle). The best exposures here are on the north side of the road. This roadcut includes both gray-weathering graphitic schists and rusty-weathering sulfidic schists of the Silurian Rangeley Formation (see Peterson, this volume - STOP 7). On the east end is a highly deformed pegmatite. Here we can see assemblages and deformational features typical of Zone V pelitic schists within the Conant Brook shear zone.

The assemblage in this, typical of Zone V within the shear zone, includes biotite, quartz, orthoclase, sillimanite, garnet, plagioclase, graphite, and ilmenite. Garnet is less abundant in the rusty-weathering schist. Garnet and feldspar form porphyroclasts in this highly deformed schist. In thin section, these have rounded shapes and asymmetric matrix tails. Coarse sillimanite and stretched quartz define a strong steeply west-plunging lineation in the outcrop and in thin section. Biotite is deep red-brown. Quartz most commonly forms finely recrystallized ribbons. Composite asymmetric porphyroclasts and lenses of quartz and feldspar appear to be attenuated and disaggregated pieces of pegmatite. Granitic veins are also disaggregated and in one place, on top of the outcrop, a vein is folded and truncated.

The pegmatite at the east end of the outcrop contains quartz, plagioclase, K-feldspar, sillimanite, and minor biotite. In thin section, the feldspar is strongly recrystallized and pulled apart into small boudins. Temperature and pressure conditions obtained using garnet-biotite thermometry and garnet-sillimanite-quartz-plagioclase barometry from a similar sample just to the north are 733°C at 6.2 kbar (Williams and Grambling, 1990 with Ti correction; Koziol and Newton, 1989)

Continue northwest on Wales Rd.
22.8 Cross Tenneco Gas Corporation pipeline.
24.6 Sharp right turn onto Munn Rd.
25.5 Turn left onto East Hill Rd.
27.4 Turn right onto Brimfield Rd.
27.9 Turn left into private driveway just before town line and park. Please note that this is private property.
(Driving time to stop ≈ 15 MINUTES)

STOP 3: CORDIERITE-GARNET-RICH SCHISTS FROM ZONE VI, POSSIBLE UPPER WARNER FORMATION, AND A GARNETITE FROM THE COYS HILL GRANITE (45 MINUTES) (Monson Quadrangle). The outcrops of interest are in the woods, near Brimfield Rd., east of the driveway. The first rocks encountered are coarse schists, typically rich in sillimanite and loaded with big garnets. This unit has been tentatively correlated with the uppermost Silurian Warner Formation (see Peterson, 1992 and this volume). The assemblage present in these schists, which includes garnet, quartz, biotite, sillimanite, cordierite, orthoclase, plagioclase, ilmenite, and graphite is typical of Zone VI pelites. In thin section these garnets are commonly large and poikiloblastic. Orthoclase forms small porphyroclasts and occurs in the matrix. Biotite, quartz, sillimanite, and cordierite dominate the matrix. Plagioclase is not abundant. Temperature and pressure conditions obtained for this outcrop are approximately 746°C at 6.5 kbar. Note that these rocks are not as highly deformed as those observed at the last stop.

Continuing east, we cross the contact with the Coys Hill granite, which is very thin (less than 100 meters thick) here. Several meters below the contact is a layer of garnetite, approximately 1 meter thick and continuous for several meters. Locally this layer is composed of several thinner layers of concentrated garnet. Note that the color of these garnets is much redder than the purplish garnets observed in the schist above. Similar garnet rich layers have been observed in the Kinsman Granite in New Hampshire and may represent some kind of "restite" (Thompson, 1985).

Turn around in driveway, return to Brimfield Rd and turn left.
28.6 Turn right on Sutcliffe Rd.
31.5 Turn left onto Hitchcock Rd.

32.0 Cross town line.
34.1 Hitchcock Rd, Tenneco Gas Corporation pipeline.
(Driving time to stop ≈ 15 MINUTES)

STOP 4: RANGELEY FORMATION AND LEUCOCRATIC GARNETIFEROUS MELT SEGREGATION (30 MINUTES) (Wales Quadrangle). This stop provides the opportunity to collect fresh samples of Qtz-Pl-Kfs-Sil-Bt-Grt±Crd gneiss of the Lower Silurian Rangeley Formation. In addition, the locality has a spectacular exposure of a leucocratic garnetiferous melt segregation thought to be the result of fluid-absent biotite-dehydration melting reactions of the surrounding gneisses. The mineral assemblage within the segregation is dominated by garnet, orthoclase and quartz. The surrounding gneiss, on the other hand, contains a greater proportion of hydrous minerals, particularly biotite, and lesser amounts of quartzofeldspathic minerals. The garnets within the concordant to slightly discordant segregation are coarser grained (2 cm) than garnets in the surrounding host gneiss.

Samples collected from a tiny roadbed outcrop (since obliterated) near the intersection of Hitchcock and McBride roads, are vein-type migmatites with similarities to the rocks observed on the pipeline. The samples contain euhedral garnet crystals ($Alm_{69.3}Prp_{26.1}Sps_{4.8}Grs_{2.8}$) to 4 cm in diameter set within a coarse-grained matrix consisting of quartz, orthoclase and sillimanite. In addition, the samples contain abundant sillimanite pseudomorphs after andalusite up to 3 cm in diameter and 9 cm in length.

Samples of pelitic schist and and adjacent melt segregation from a site just west of Mt. Hitchcock Road on the pipeline were analyzed in detail by electron microprobe. Biotite analyses reveal that the most Fe- and Ti-rich biotites are those present in the leucocratic segregation as inclusions in K-feldspar (XMg = 0.498 - 0.521; Ti/11 oxygens = 0.292 - 0.335). The most magnesian biotites are isolated matrix grains within the leucocratic segregations (X_{Mg} = 0.552 - 0.573; Ti/11 oxygens = 0.258 - 0.283). Gneiss matrix biotites have an intermediate X_{Mg} (X_{Mg} = 0.534 - 0.544; Ti/ 11 oxygens = 0.251 - 0.267). The biotite inclusions in K-feldspar may represent compositions that existed in the gneisses just as melting to form the segregations began. The compositions of garnet grains within the melt segregation ($Alm_{70.7}Prp_{25.6}Sps_{1.9}Grs_{2.3}$) differ slightly from that of the associated gneiss ($Alm_{69.7}Prp_{25.2}Sps_{2.3}Grs_{3.7}$), although this may not be true for all melt segregations and gneisses. Garnets in both the melt segregations and the gneisses are homogeneous except where adjacent to biotite and cordierite. When compared to average garnet core compositions, garnet compositions adjacent to biotite are characterized by higher X_{Fe}, higher almandine and spessartine content, similar to slightly lower grossular content, and lower pyrope content.

Thermobarometric calculations based on garnet core and isolated matrix biotite compositions of the gneiss and adjacent melt segregation suggest that the melt segregations formed under conditions similar to the peak granulite facies metamorphism recorded in the host gneisses. The garnet-biotite calibration of Thompson (1976) yields peak metamorphic temperatures of 700 - 710°C. Minimum pressure estimates based on garnet compositions and the calibration of Tracy et al. (1976) are 6.2 - 6.3 kbar. Temperatures calculated using the Fe- and Ti-rich biotite inclusions in K-feldspar within the segregation together with garnet core compositions range from 750 - 780°C. These would represent maximum temperatures for leucocratic garnet melt segregation formation if it is assumed that the biotite inclusions did not re-equilibrate and that they are, in fact, in equilibrium with garnet core compositions.

Sample MA/Wales 287 of coarse-grained sillimanite-orthoclase-garnet-biotite schist was collected here. Monazites from this sample give a concordant age of 363±1 Ma (Table 1), showing that the partial melting and metamorphism here is Devonian.

Return to cars.
Continue south on Hitchcock Rd.
34.2 Turn left onto McBride Rd.
34.3 Big rocks on both sides of road. Analyses of Robinson et al. (1982) are from here. One piece of sillimanite-garnet-biotite gneiss collected here by Jack Cheney contains retrograde kyanite and small, euhedral staurolite grains. Rocks containing large sillimanite pseudomorphs after anadalusite are present 100 feet to the west under the road, in former outcrop.
34.5 End of McBride Rd. Turn left (east) on Monson Rd.
35.3 Stop sign. Turn left onto Rt. 19N, toward Brimfield.
35.7 Stay on Rt. 19N, bearing right at monument.
36.9 Take second right onto Holland Rd., Wales, toward the Wales Country Lounge.
37.6 Continue straight at town line. Now on Wales Rd., Holland.

38.0 Natural outcrops on left, of sillimanite-orthoclase-garnet-cordierite schist in the heart of Zone VI.
38.4 Stop sign. Turn right on Brimfield Rd. toward Holland.
38.8 Overgrown roadcut on right at inside of curve in road was Stop 8 of the 1982 NEIGC trip by Robinson, Tracy, Hollocher, and Dietsch. This outcrop gave their highest reliable garnet-biotite temperature estimate for Zone VI (745 °C), and a pressure estimate of 6.4 kb. Sample MA/Wales 187 was also collected from this outcrop, and it yielded a monazite age of 367 +/- 2 Ma (Table 1).
40.0 Cross pipeline trench. A wollastonite-bearing sample was collected from the Rattlesnake Mtn. segment of the trench, about 3500 feet west of here (P-44 of Berry, 1989; 1991).
40.1 Bear left at fork, staying on Brimfield Rd.
40.6 Continue straight at blinking light in Holland. Now on Mashapaug Rd.
41.3 Hamilton Resrvoir on left.
41.5 Park off road to right. This is private property in the vicinity of unattended expensive summer cottages close to hunting season, so local security concerns are high. Take the "no tresspassing" signs seriously. Follow gravel road up hill to abandoned shack. Walk due west about 1000 feet, passing abundant but irrelevant outcrops to the top of the hill.
(Driving time to stop ≈ 25 MINUTES)

STOP 5: ULTRAMAFIC ROCK IN LEADMINE POND GNEISS (45 MINUTES) (Wales Quadrangle). Near the top of the hill are two outcrops of dark brown to black orthopyroxene-olivine hornblendite (Fig. 16). The ultramafic rock has 1 to 2 cm, amber-brown orthopyroxene megacrysts set in a medium-grained groundmass of shiny, black hornblende. Olivine occurs as smaller rounded grains, more difficult to identify in hand specimen. The poikilitic orthopyroxene megacrysts are slightly elongate and define a weak, but clearly discernable foliation. Some megacrysts contain black streaks of hornblende parallel to their length. In thin section, the orthopyroxene is a pale plum color, and the hornblende is yellow-brown to slightly greenish-brown. The smaller, round olivine grains are scattered throughout the thin section. Dark green spinel makes up several percent of the rock.

The outcrop just east of the ultramafic rock is amphibolite. The outcrop 120 feet along strike to the south, on the west side of the gulley, is layered plagioclase-quartz-biotite-hornblende gneiss. These enclosing rocks belong to the Leadmine Pond Gneiss, so the ultramafic rock is assigned to the Leadmine Pond Gneiss as well. A second occurrence of orthopyroxene hornblendite within the Leadmine Pond Gneiss is exposed 7 miles north of here in Brimfield, on the south side of the Massachusetts Turnpike near the east end of the Chamberlin Mtn. road cut (Berry, 1989). These two occurrences are similar to ultramafic pods in the Monson Gneiss to the northwest (Tracy, et al., 1984), although the rock here is at significantly higher metamorphic grade.

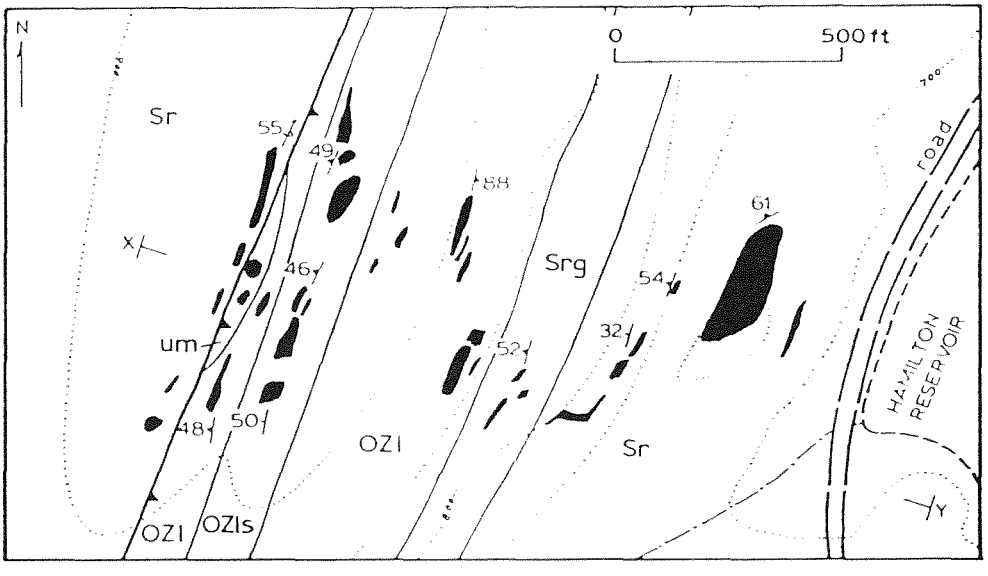

Figure 16. Detailed map showing the ultramafic rock at STOP 5. Symbol "um" indicates ultramafic rock. Outcrop shown in black on map. Map units beginning with "Ozl" are members of the Leadmine Pond Gneiss. Units beginning with "Sr" are assigned to the Rangeley Formation (from Berry, 1989).

No grand tectonic significance is attached to this ultramafic rock. The field relations suggest that the ultramafic pod here and the one in the Turnpike cut are merely minor compositional variants in a sequence of layered gneisses. No evidence of an ultramafic complex or ophiolite suite, for example, has been recognized. The suggestion that the hornblendites represent metamorphosed lavas or shallow intrusives of very mafic or ultramafic composition (Tracy et al., 1984), is tentatively adopted here.

Return to cars.

41.9 Cross Hamilton Reservoir Causway. The samples studied by Tracy and Dietsch (1982) were collected by Robinson and Klepacki from the loose blocks along the causeway (Robinson et al., 1982).

43.5 CT State Line - well layered hornblende-clinopyroxene-plagioclase-scapolite gneiss with secondary epidote + cross-cutting quartz-plagioclase-clinopyroxene pegmatite dikes (see Robinson et al., 1986a, 1989).

43.8 Two pyroxene granulite with veins of tonalitic partial melt (see Robinson et al., 1986a, 1989).

44.0 Turn left onto I-84 East.

45.0 Massachusetts Welcomes You!!

48.1 Exit 1 - Mashapaug Rd. - Southbridge.

48.2 Take right at end of exit ramp onto old Rte 15.

48.5 Pass under I-84 East.

48.9 Park cars near the entrance to the dead Sturbridge Isle. Walk to outcrops and blasted rock behind old, abandoned fuel island.

(Driving time to stop ≈ 5 MINUTES)

STOP 6: CORDIERITE±GARNET-BEARING PEGMATITE, SCHISTS, AND GRANULITES OF THE SILURIAN(?) PAXTON FORMATION (45 MINUTES) (Southbridge Quadrangle). This stop will allow you to collect pristine samples of cordierite±garnet-bearing pegmatite thought to be the result of fluid-absent, biotite-dehydration melting reactions of the surrounding gneisses. The host rock at this locality consits of quartz-sillimanite-garnet-cordierite schists and interlayered biotite and calc-silicate granulites of the Silurian(?) Paxton Formation.

The pegmatite consists of the assemblage Qtz-Pl-Kfs-Sil-Bt-Crd-Grt. Cordierite within the pegmatite is blue to dark lavender in color and up to 8 cm across. Some samples from this locality show large dark patches up to 15 cm across that appear to be a pinite alteration of original cordierite.

Samples from this locality show evidence of cordierite breaking down to aggregates of garnet + sillimanite + quartz. The cordierite in the vicinity of these aggregates is zoned, particularly within 1000 μm of the aggregate (Fig. 9). Cordierite far from the aggregate has an XFe of 0.30 - 0.31. The XFe of the cordierite decreases dramatically as the aggregates are approached and becomes as low as 0.22 directly adjacnt to garnet. Garnet within the aggregates is also strongly zoned (Fig. 10). Pyrope contents are generally between 26 and 28%.

Conditions of the onset of cordierite breakdown were estimated at 762°C and 5 kbar using the aggregate garnet "core" compostion and the cordierite composition far from the aggregate. The last recorded conditions of cordierite breakdown have been estimated at 617°C and 6 kbar based on the compositions of the aggregate garnet rim and the adjacent cordierite. All calculations used the calibrations of Thompson (1976) and Bhattacharya (1986). Conditions of the onset of cordierite breakdown recorded in a number of samples from the pipeline are between 760 and 800°C, 5 - 5.5 kbar, similar to those recorded at this locality. However, the sample from this locality records "final" conditions of breakdown that suggest lower pressures and higher temperatures than those recorded in the west. The last recorded conditions of cordierite breakdown are between 570 and 590°C and 6.6 - 6.7 kbar in samples 10.5 miles to the west. This data may suggest that, during the late stages of cordierite breakdown, temperatures were slightly lower and pressures slightly higher in the west than in the east.

One monazite fraction from this locality gave a concordant age of 345 Ma, but two other monazite fractions from the same sample gave discordant analyses (Table 1). The fact that two of the three fractions were discordant and that the three fractions did not agree with each other suggests that the monazite population is not homogeneous. It is possible that the condordant analysis at 345 Ma is dating a geologic event. Alternatively, if the monazites have experienced a more complicated geologic history, as the data suggest, the 345 Ma age may represent a meaningless "average age" based on analysis of grains with different isotopic characteristics.

Return to cars, continue south on old Rte 15.
50.5 Leadmine Rd., stay straight on old Rte 15.
51.3 Carefully park on right side of road, in former exit ramp to abandoned parking lot. Watch out for potholes and broken glass. Walk south along road .
(Driving time to stop ≈ 5 MINUTES)

STOP 7: CORUNDUM-BEARING GNEISS IN LEADMINE POND GNEISS (45 MINUTES) (Southbridge Quadrangle). The rock is a medium-gray to light-gray, granular textured, biotite-plagioclase-orthoclase-garnet-corundum-sillimanite-spinel gneiss. The red-brown garnets, 0.5 to 1 cm in diameter, stand out from the smoothly weathered outcrop surface. There is no prominent layering, but there are lighter and darker areas and the concentration of garnet is different from place to place, suggesting there is a small range in bulk composition within the outcrop. A weak foliation is locally present, defined by wispy sillimanite-rich streaks (Fig. 17).

The coexistence of garnet, corundum, sillimanite, and a spinel mineral suggests that the metamorphic conditions may have approximated the end-member reaction: Almandine + 5 Corundum = 3 Hercynite + 3 Sillimanite, which has been determined experimentally by Shulters and Bohlen (1989). For a chosen pressure of 6 to 6.5 kb, which is the pressure suggested from nearby pelitic schists, the reaction curve in the end-member system (FeO-Al_2O_3-SiO_2) passes through about 800 to 825°C. Electron microprobe data show the spinel from this locality to have a composition of $Hercynite_{80}$ $Spinel_{17}$ $Magnetite_2$ $Ghanite_1$, with only trace amounts of Cr and Ti, and no detectable Mn. Garnets have a composition of $Almandine_{76.7}Pyrope_{16.7}Grossular_{4.8}Spessartine_{1.8}$ (Berry, unpub. data). The presence of accessory graphite and ilmenite in the thin section implies that ferric iron is probably not a significant thermodynamic complication. Based on this preliminary work, it seems that the mineral assemblage does indeed approximate the calibrated end-member reaction. Allowing for some uncertainty in the equilibration temperature due to activities less than unity of almandine and hercynite in the natural sample, the phase relations in this rock are compatible with the peak metamorphic conditions estimated from schists, mafic rocks, and calc-silicate rocks in Zone VI.

The bulk composition of this rock, as reflected by the presence of the silica-free minerals corundum and hercynite, has an unusually low silica : alumina ratio compared to normal igneous and sedimentary rock compositions Some process has presumably depleted the rock or its protolith in silica. For example, chemical weathering may have produced a deposit enriched in aluminous clay minerals, or hydrothermal alteration in a volcanic environment may have produced a silica-depleted rock of appropriate bulk composition. Another possibility, suggested by Peter Robinson (pers. commun., 1992), is that chemical interaction of a tonalitic liquid with adjacent wall rock could have "desilicated" the wall rock, leaving an aluminum-enriched restite. This is an attractive suggestion because the mapped unit immediately to the west of the corundum-bearing gneiss contains partial melt veins of tonalite (Hollocher, 1985; Robinson et al., 1986a).

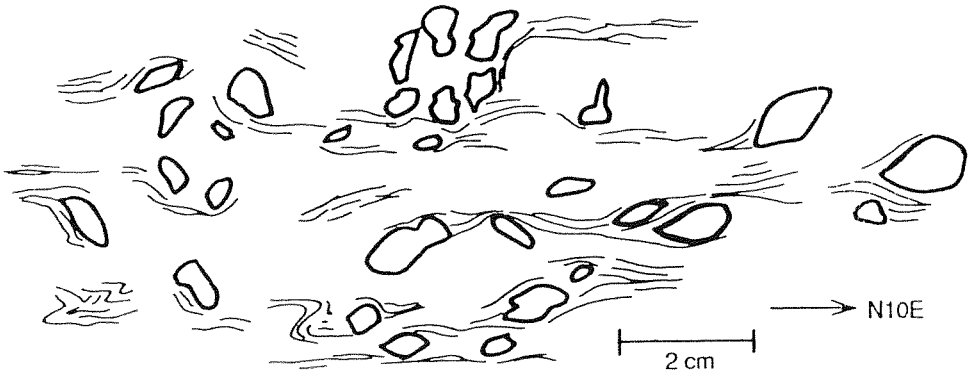

Figure 17. Sketch of corundum-bearing gneiss from a pavement outcrop at STOP 7. Wispy trails, rich in sillimanite and biotite are indicated by lines; garnet megacrysts, about 0.5 to 1 cm in width, are outlined. Remainder of rock is granular K-feldspar, corundum, biotite, and spinel.

Walk S10W about 250 feet to float blocks of calc-silicate granulite.

These white rocks with dark green spots represent the calc-silicate unit immediately to the east of the corundum-bearing gneiss. Although most of the pieces exposed here are loose, outcrop mapping shows that they are approximately in place. The characteristic rock type is a scapolite-plagioclase-quartz-diopside-calcite-sphene±graphite granulite that probably represents a metamorphosed siliceous limestone. Wollastonite has not been found in this unit to date, suggesting high CO_2 activity in the sampled rocks. Samples that might contain wollastonite at this metamorphic grade must have had a lower activity of CO_2 and are more likely to contain grossular than scapolite as the aluminosilicate phase.

Unsavory retrograde features such as green amphibole rims on diopside, and calcite veins through scapolite have begun to permeate this rock. Such blatant retrograde textures are decidedly rare elsewhere in Zone VI. But in this area, near the eastern edge of Zone VI, the high grade rocks contain widely spaced muscovite-bearing granite and pegmatite dikes, local retrograde textures, and slightly young, perhaps disturbed monazite ages (sample MA/St 187, Table 1). Guidotti, Herd, and Tuttle (1973) observed that Barker's (1962) report of pegmatites with microcline rather than orthoclase in the high-grade gneisses northeast of here, may indicate polymetamorphism. None of these features have been examined in detail, but taken together they suggest that the place where the intensity of Acadian metamorphism begins to decrease toward the east may be about where Late Paleozoic thermal effects begin to increase toward the east. This is a good place to end our trip. Thank you for coming.

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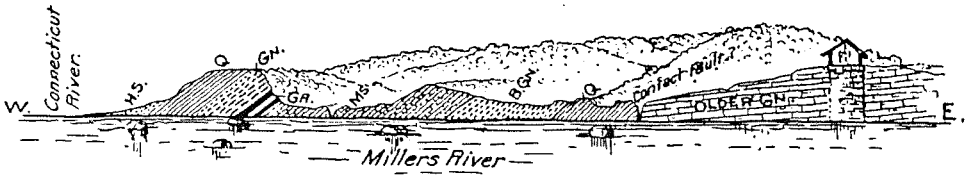

FIG. 21. Sketch of rocks at mouth of Millers River, looking northeast from B on map, fig. 20. Scale, 1: 2000. H. S.= hornblende-schists; Q.=quartzite; GN.=gneiss; GR.=granite; M. S.=mica-schist; B. GN.=biotite-gneiss; OLDER GN.=Monson Cambrian gneiss.

Fig. 1.

45 in. x 44.

Fig. 2.

Box 4 ft. 10 in. x 20.75 in.

Fig. 3.

4 ft.

L.M. Bradford & Co's Lith.